Karst Bauxites

Bauxite Deposits on Carbonate Rocks

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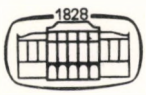
GYÖRGY BÁRDOSSY



AKADÉMIAI KIADÓ · BUDAPEST 1982

THIS BOOK IS THE COMPLETED AND TRANSLATED VERSION OF
KARSZTBAUXITOK
AKADÉMIAI KIADÓ, BUDAPEST 1977

TRANSLATED BY
B. BALKAY

JOINT EDITION WITH ELSEVIER SCIENTIFIC PUBLISHING COMPANY
Amsterdam–Oxford–New York

ISBN 963 05 2480 5

AKADÉMIAI KIADÓ, BUDAPEST 1982

PRINTED IN HUNGARY

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INTRODUCTION

A description of all the karst bauxite regions of the world would require a multi-volume monograph whose great bulk would hamper the highlighting of the main features of interest. For this reason, the writer has striven above all to compare bauxite regions on a quantitative basis wherever possible, relying in doing so on the methods of mathematical statistics. This quantitative processing and assessment of the information available is the most essential novel feature of this book.

The first part of this volume presents observational and analytical evidence concerning karst bauxites. The typical features of bauxites were analysed from a variety of aspects and results from different bauxite regions were compared. Much attention was paid to Hungarian bauxite deposits and it is these that figure in the greatest detail in the comparative assessment. More or less metamorphosed karst bauxites were also considered, as there is no sharp limit between metamorphic (emery) and non-metamorphic bauxites: the transition being gradual, any delimitation would be arbitrary.

The second part is concerned with the conditions of formation of karst bauxites and with the factors controlling their geographic and stratigraphic distribution. The treatment is of a synthesising type, intended to elucidate the interaction in space and time between all the known factors.

The writing of this book was much facilitated by the fact that, in recent years, the writer has had occasion to visit the principal karst bauxite regions of Austria, Bulgaria, France, Greece, Iran, Italy, Jamaica, Rumania, Spain, the USA, the USSR, the Socialist Republic of Vietnam, and Yugoslavia. Bauxite samples from all other karst bauxite regions of the globe were made available to the writer by courtesy of bauxite specialists at home and abroad. Comparisons with lateritic bauxites were made possible by visits to the principal bauxite zones of Brazil, the Guinean Republic and India.

Over and above the viewpoint of scientific research, the writer has paid special attention to aspects useful for practical bauxite prospecting and processing.

The writer wishes to express his sincere gratitude to Academician Elemér Szádeczky-Kardoss who authorised the research and supported him with his advice. In the analytic work, in data processing and in statistical calculations, the writer was assisted by Miss Mária Tóth, Miss Ilona Buday and Miss Olga Komoróczy. The photos were prepared by Mr Lajos Klinda. Thanks are due to Miss Anikó Tóth for the painstaking preparation of the drawings.

The scientific readers of the book were Dr. Kálmán Balogh and Dr. Vilma Széky-Fux; the author takes this opportunity to express his gratitude to them.

He is indebted to Bálint Balkay, György Komlóssy, Péter Ottlik and Ferenc Zenkovics for reading chapters of this book, for their helpful advice and remarks. Last but not least, it would be quite a task to enumerate all those geologist, mineralogist and technologist colleagues who have contributed ideas, remarks and friendly assistance in the preparation of this book, but sincere thanks are expressed to them all.

No international consensus has as yet evolved concerning the terminology of bauxite geology. The writer therefore deems it necessary to specify the meaning of the principal terms as they are used throughout this book.

Bauxite. A residual or sedimentary rock in which the aggregate content of the oxide and/or hydroxide minerals of aluminium, iron and titanium is greater than 50 %, and aluminium minerals are more abundant than the minerals of the two other elements together. The question as to what is a bauxite, then, must be decided on the strength of the quantitative mineralogical composition. The definition of bauxite as an ore of aluminium is equivocal because the limit between ore-grade and sub-grade bauxites is in continual change, depending on the development of refining technology and of demand. The types of bauxite encompassed by this broad definition are discussed in § 4.1.

Bauxite ore body. That part of the bauxite complex (see below) which is made up of ore grade or potentially ore-grade material (bauxite, clayey bauxite, bauxitic clay).

Bauxite genesis. The totality of those processes of alteration and sedimentation which lead to the formation or rocks enriched in aluminium, including processes of weathering, transportation, and deposition.

Bauxitisation. That phase of bauxite genesis by which the material of some rock is converted into bauxite or by which the conversion into bauxite of a rock already partially bauxitised is continued.

Lateritisation. A form of superficial weathering resulting in an increase of aluminium and titanium contents, also frequently but not invariably of iron contents, and in a decrease of alkali, alkali earth and silica contents. Bauxitisation is a special case of lateritisation, characterised by a prevalence of aluminium enrichment. *Laterite* is a generic term for the rock products of lateritisation.

A hierarchy of four bauxite geological/depositional units in ascending order of size is distinguished:

(1) *Bauxite deposit.* A spatially coherent (uninterrupted) depositional unit of uniform genesis, composed of bauxite and bauxitic rocks (e.g. the bauxite lens Darvastó VI). In certain depositional types, several bauxite ore bodies can be distinguished within a bauxite deposit; these may be separated by bauxitic rocks, and possibly by clay, sand or lignite. Even in this case, however, the bauxite deposit must be genetically coherent.

(2) *Bauxite deposit group.* Several bauxite deposits lying close together in the horizontal sense but separated by a comparatively broad barren zone from any bauxite deposits not belonging to the group (e.g. Nyirád bauxite deposit group). Typical deposits have an areal extent in the range 10–100 km^2. In a bauxite deposit group featuring a largish number of deposits, it is usual to distinguish subgroups for better orientation. In the extreme case, even a single bauxite deposit may be regarded as a

'group' if it is geologically separate and geographically remote from some other groups. In some areas, bauxite formation repeatedly took place in different geological periods. In the writer's experience, deposit features and genetics in each of these periods tend to differ rather substantially; they are accordingly regarded here as separate groups (e.g., in the Obrovac–Novigrad–Drniš region of Yugoslavia, a Paleocene and a Middle Eocene deposit group are distinguished).

(3) *Bauxite region.* A complex of geologically related bauxite deposit groups, separated from neighbouring bauxite regions, if any, by a barren zone or by a change in geology (e.g. the Transdanubian Hills bauxite region of Hungary). Their areal extent is in the range of from several hundred to several thousand km^2. Kirpal' (1977) distinguished bauxitic zones within bauxite regions, comprising deposit groups that constitute elongate bands or zones. In the writer's opinion, this hierarchic sub-unit is unnecessary except in some bauxite regions of exceptional extent. A bauxite region, as a unit of higher rank both in the spatial and in the temporal sense, may encompass several bauxite deposit groups of different ages.

(4) *Bauxite belt.* A coherent group of bauxite regions in a second- or third-order geotectonic unit (e.g. the Mediterranean bauxite belt), characterised by a uniform geostructural history. The areal extent is from some hundred thousand to a few million km^2 with a length or width of several thousand km. Somes bauxite belts have had a single episode of bauxite formation; in others, two or more such episode may have taken place.

Bauxite complex. A term coined by Vadász (1951) for the rocks making up a bauxite deposit; it includes in addition to bauxite also all other rocks genetically connected with it.

Bauxitic formation. A stratigraphic succession, one or more horizons of which include one or more bauxite deposits. The facies of the bauxitic formation are genetically related to the bauxite deposit and, in fact, control its depositional features.

1. CLASSIFICATION OF BAUXITE DEPOSITS

A much debated issue, the classification of bauxite deposits has not as yet produced a consensus. This is due to the manifold features of the deposits which can thus be classified from a number of different aspects. What the specialists cannot agree on is the relative importance of these features.

The most striking difference recognised early in this century is between bauxite deposits overlying silicate rocks on the one hand and carbonate rocks on the other. It was on this basis that Harrassowitz (1926) divided bauxites into silicate bauxites and calcareous bauxites. Rather widely accepted since, this subdivision is still current especially in textbooks written in the German language (Petrascheck and Petrascheck 1950, Schneiderhöhn 1962, Cissarz 1965).

Fox (1932) subdivided bauxite deposits on a genetic basis. In his opinion, lateritic bauxites ('aluminous laterites') are the products of the lateritic weathering of silicate rocks, whereas terra-rossa bauxites derive from terra rossa, the weathering residue of limestones and dolostones. His subdivision was adopted also by De Weisse in his monograph of 1948.

Vadász (1951) rejected the terms 'silicate' and 'calcareous' bauxite as conducive to misunderstandings, and introduced instead the terms 'lateritic bauxite' and 'karst bauxite', appreciating in the latter them of the role played by the karstification of a carbonate bedrock in the evolution of some deposits. This simple and easy-to-apply subdivision became fairly widely accepted in the literature (cf e.g. Zans 1959, De Weisse 1964).

Roughly at the same time, some other classifications were also elaborated, without eliciting much response. For example Harder (1952) proposed the following classification based on parent-rock composition:

(1) Deposits derived from alkali- and aluminium-rich silicate rocks.
(2) Deposits connected with limestones.
(3) Deposits connected with sedimentary clays.
(4) Deposits connected with intermediate and basic eruptives.
(5) Deposits due to the vigorous weathering of rocks of medium aluminium content.

The groups thus defined cannot be unequivocally distinguished; the place of a given deposit in this or that group is often debatable.

Hose's (1960) classification rests on a morphological basis:

(1) Deposits developed on peneplains.
(2) Deposits formed on volcanic domes and plateaux.
(3) Deposits on karsted limestone plateaux.
(4) Sedimentary reworked deposits.

This classification also failed to gain broad acceptance, owing to the uncertainty inherent in separating the first group from the second.

Valeton (1972) pointed out that some deposits fit neither into the laterite nor into the karst bauxite group. In view of this fact, she developed the following classification based on bedrock lithology:

(1) Bauxite deposits overlying igneous and metamorphic rocks.
  (1.1) Slope type.
  (1.2) Plateau type on basic igneous rocks.
  (1.3) Plateau type on variable rock types.
(2) Bauxite deposits on sedimentary rocks.
  (2.1) Bauxites on clastic sediments.
  (2.2) Bauxites on carbonate rocks.
  (2.3) Bauxites on phosphate rocks.

Although consistent, Valeton's classification is one-sided, paying attention neither to the abundance of the types of deposits, nor to their depositological features.

Concentrating on depositological features, Patterson in his turn (1967) proposed the following classification:

(1) Blanket deposits: flat-lying layers of variable thickness.
(2) Interlayered deposits: discontinuous layers and lenses interstratified in sedimentary and/or volcanic rocks. Some are residual; some are redeposited.
(3) Pocket deposits: fillings of depressions, most often in dolomite and limestone, but sometimes in other rocks too. This group includes a subgroup of 'interlayered pocket deposits' (pockets intercalated in a sequence of sedimentary deposits).

Patterson's groups offer no clear-cut distinction either in the depositological or in the genetic sense. The use of his classification would therefore lead to numerous ambiguities.

The classification proposed by Grubb (1973) is based on the altitude at which bauxitisation has taken place:

(1) High-level or upland deposits.
(2) Low-level or peneplain-type deposits.

Grubb placed into the first group the residual deposits formed above the ground water table under fair drainage conditions, including among others the lateritic bauxite deposits of Guinea, Ghana, India and Hawaii.

He placed into the second group the residual, redeposited and allochthonous deposits characterised, according to him, by repeated fluctuations of the ground water table, citing as typical examples the deposits of Gove and Weipa in Australia, and those in Guyana and Surinam. Deprecating the distinction of a separate karst bauxite group

as superfluous, he placed the bauxites of Jamaica and Hispaniola into the high-level group and all other karst bauxites into the low-level group.

The distinction of high- and low-level deposits among the lateritic bauxites is doubtless justified and is not a novel idea either. The rejection out of hand of the karst bauxite group, however, is entirely unjustified. The Jamaican bauxites have incomparably more in common with the European karst bauxites than with the high-level lateritic bauxites. It is extremely difficult to decide in the case of a buried deposit of a past period whether bauxitisation in its time took place on an elevated or on a low-lying topography. Last but not least, there are deposits which, in the course of their long history of evolution, alternated between high and low positions as a result of vertical (epeirogenic) movements in the region. Grubb himself cites some such deposits, including those of Malaysia and of the Darling Range in Australia. All in all, such a classification cannot possibly be sufficiently unambiguous. The present writer accordingly sees no point in adopting it, particularly in extending it to the karst bauxites.

In the USSR, the classification of bauxite deposits took a course different in many respects from the above. Deposits were classified on a purely genetic basis early on. Malyavkin (1937), for example, distinguished lateritic, metasomatic, organogenic and metamorphic deposits. Likewise on a genetic basis, A. D. Arkhangelsky (1937)—who, remarkably enough, denied the existence of lateritisation and of lateritic bauxite deposits—classified bauxite deposits into groups of (1) marine, (2) lacustrine and/or marshy origin. He assumed all the deposits of the Mediterranean region to be of marine origin.

Vikulova (1946) distinguished five types of deposits according to their sites of accumulation: (1) deposits on plateaux forming watersheds, (2) valley-floor deposits, (3) deposits filling karst depressions, (4) lacustrine deposits, (5) lagoon and seashore types.

It was Peive (1947) who first recognised the importance of geotectonic position and, on this basis, distinguished two types of deposits: (1) those of the geosynclinal zones, and (2) those of the platform regions. According to him, the former deposits are all marine, and all overlie carbonate rocks; the latter ones are all residual or lacustrine or marsh-deposited. He placed all the deposits of the Urals and of the Mediterranean region into the first group. Peive's classification was adopted by Goretsky *et al* (1949), Rukhin (1958) and by many others. Sapozhnikov (1970, 1971) identified karst bauxites with the deposits of the geosynclinal zones.

The geotectonic classification was in fact based on the recognition of objectively existing, important relationships. In the form cited, however, it represents something of an oversimplification: in addition to geosynclinal zones and platforms, there also exist other types of geotectonic units, which do not lack bauxite deposits, either. Nor is the relationship postulated by Peive of a general validity: bauxite deposits overlying carbonate rocks do exist also on platforms, and laterite bauxite deposits are known also from geosynclinal zones (e.g. in Oregon State, US).

Aiming to eliminate these contradictions, Goretsky (1960) developed the following geotectonic classification, which is more detailed than those cited above:

(1) Deposits in stable zones of platforms.
  (1.1) On anteclises.
  (1.2) On syneclises.
(2) Deposits on mobile parts of platforms.
  (2.1) In the borderlands of syneclines.
  (2.2) In intramontane and foreland depressions.
(3) Deposits of geosynclinal belts.
  (3.1) In zones of transition between platforms and geosynclines.
  (3.2) In the outer part of geosynclinal zones.
  (3.3) In median masses.

Goretsky subdivided each of these groups into two or three subgroups on the basis of bedrock morphology, obtaining a total of 16 subgroups. Goretsky's classification does account for almost every geostructural type, and is self-consistent but its application in practice is fraught with a great deal of uncertainty. Different authors tend to hold different views concerning the structural position of this or that region, particularly the question as to what parts of a platform are to be regarded as stable or mobile. Goretsky himself, for example, characterised the South Ferghana bauxite region as a geosynclinal zone, whereas on the geotectonic map of the USSR, published in 1962, it figures as a gently folded platform. Another deficiency of Goretsky's classification is that—following A. D. Arckhangelsky—he pays hardly any attention to lateritic bauxite deposits. He places them in the same group with the allochthonous lacustrine and marshy bauxite deposits, confining them to the stable platform regions.

Kirpal' (1972a) distinguished residual and sedimentary deposits. His further subdivisions of both groups are based on tectonic position distinguishing platform and geosynclinal zones. Within these main groups, he distinguished a total of 14 subgroups on the basis of deposit and bedrock morphology. However, even his primary criterion of classification involves an ambiguity as in addition to pure autochthonous and pure allochthonous deposits there are mixed (allo–autochthonous) ones, the position of which cannot be decided unequivocally in his system. In his tectonically-based subgrouping, all the contradictions inherent in Peive's classification emerge again.

With a view to getting rid of these difficulties, Kirpal' (1977), leaving aside the issue of tectonic position, proposed a new classification in terms of the following four groups and 16 subgroups.

(1) Lateric bauxite (residual) deposits.
  (1.1) Deposits of linear form.
  (1.2) Deposits of lenticular form.
  (1.3) Deposits of bowal type.
(2) Polygenetic bauxite deposits (with local reworking).
  (2.1) Lenticular karst deposits.
  (2.2) Lenticular deposits.
  (2.3) Stratiform deposits.
(3) Sedimentary bauxite deposits in terrigenous sequences.
  (3.1) Valley-floor deposits.
  (3.2) Stratiform deposits.
  (3.3) Karst deposits.

(3.4) Karst-contact deposits.
(3.5) Kettle-shaped karst deposits.
(3.6) Karst-contact ('contact-follower') kettle-shaped deposits.
(4) Sedimentary bauxite deposits in carbonate sequences.
(4.1) Stratiform karst deposits.
(4.2) Kettle-shaped karst deposits.
(4.3) Lenticular karst deposits.
(4.4) Funnel-shaped karst deposits.

This classification is more unequivocal than the foregoing one, but the distinction of subgroups, especially those within Group 3, is not without ambiguity.

Bushinsky (1971), rejecting references to tectonic position, developed a classification based on genetic and depositological factors:

(1) Lateritic bauxite deposits (residual ones on alumosilicate rocks).
(1.1) Extralateritic bauxite deposits (overlying unweathered rocks).
(1.2) Sublateritic deposits (with a clayey 'lithomarge' between bauxite and bedrock).
(2) Complex deposits (lateritic and sedimentary bauxite deposits within one and the same deposit group).
(3) Sedimentary (clastogenic) bauxite deposits.
(3.1) Deposits overlying silicate rocks.
(3.2) Deposits overlying carbonate rocks ('karst deposits').
(4) Sedimentary-chemogenic bauxite deposits.
(4.1) Volcanogenic-sedimentary deposits.
(4.2) Deposits due to weathering of sulphide bearing rocks.
(5) Metamorphic deposits.

The existence of volcanogenic-sedimentary deposits is stipulated on a theoretical basis: no such deposit has been discovered so far, but Bushinsky deems it possible for such deposits to exist somewhere.

The present writer admits Bushinsky's classification to be better than those preceding it in that it encompasses practically all existing types of deposits. It nevertheless retains some degree of uncertainty in choosing genetics as the principal criterion of classification, as it is precisely in the assessment of genetic issues that the opinions of specialists diverge most widely.

In the Explanatory Note to the bauxite geology map of the USSR (1973), a 16-member authors' collective subdivided the bauxite deposits into three basic types corresponding to three bauxitic formations, distinguishing within those types some further groups on a morphogenetic basis.

(1) Bauxite deposits of the lateritic formations.
(1.1) 'Bowal type' with residual bauxite.
(1.2) 'Slope type', with clastic-reworked bauxite.
(2) Bauxite deposits in carbonate formations.
(2.1) Karst type.
(2.2) Infiltrated karst type.

(3) Bauxite deposits in terrigenous formations.
(3.1) Karst type.
(3.2) Stratiform type.

This threefold basic grouping has the advantage of taking into account the geological evolution as a whole of the region considered. The transitions between pure carbonate and pure terrigenous formations being gradual in Nature, however, the position of several deposits is thus left uncertain. From a bauxite geological viewpoint, the fact that the karst type is divided up between the carbonate and the terrigenous subgroup is not fortunate, as it is desirable to have all the karst bauxites under the same heading. The distinction of the karst, infiltrated-karst and stratiform types is not unequivocal, either.

In an emphatically genetic classification, Sapozhnikov (1975) distinguished residual (lateritic), sedimentary, polygenetic, infiltrated and metamorphic bauxites, which he further subdivided into a total of 15 subgroups. Karst bauxites constitute a separate class within his sedimentary group; some of the infiltrated and metamorphic bauxite deposits, however, are in fact also karst bauxites in essence.

In developing the classification to be used in this book, the writer has aimed at eliminating the uncertainties referred to above. He has therefore placed the main emphasis on viewpoints that are fully unequivocal and readily accessible to direct observation in the deposits. Of these, he deemed bedrock lithology to be the most important. As distinct from Valeton, however, the writer has not opted for a simple petrographic classification, based on the rock underlying the deposits. On the contrary profiting by the depositological experience accumulated, he has attributed a special importance to carbonate rocks:

(1) Bauxite deposits overlying alumosilicate rocks.
(2) Bauxite deposits overlying carbonate rocks.

In keeping with common usage, the deposits of the second group will be called 'karst bauxite deposits', regardless of whether the bedrock surface is strongly or gently karsted. The present book is concerned with the karst bauxite group, a more detailed classification of which is given in § 3.1.

The deposits overlying alumosilicate rocks are further subdivided depending on whether they were formed by decidedly residual processes or by processes including redeposition and transportation:

(1.1) *Laterite bauxite deposits.* Deposits overlying alumosilicate rocks, and formed by the *in situ* lateritisation of those. This includes bauxites whose redeposition did not take them beyond the outline of a bauxite deposit group (this latter is the 'slope type' of the Soviet classification of 1973, and the 'polygenetic type' of Saposhnikov 1975 and Kirpal' 1977).
(1.1.1) Residual deposits.
(1.1.2) Deposits of short-range (local) redeposition.

(1.2) *Tikhvin-type deposits.* Transported (allochthonous) deposits overlying alumosilicate rocks, divorced from any residual laterite profiles.
(1.2.1) Stratiform deposits.
(1.2.2) Valley-floor deposits.

Although the writer admits the importance of geotectonic position, he has rejected it as a criterion of classification. Today, structural geologists tend to diverge in the geotectonic assessment of many a region, all the more so since, over the past decades, geotectonics has been well and truly revolutionised by the adherents of the 'plate tectonic' theory who have developed an altogether novel geotectonic classification. At the same time, however, the adherents of the conservative fixistic school have also retained their well-established classifications. In such a situation, a classification of bauxites on a structural basis must of necessity involve numerous uncertainties, and reassessments would be needed almost year by year. The writer feels that the introduction of a classification in a structural basis will not be justified until some progress is made in geotectonical knowledge and until more of a consensus is attained concerning geotectonic issues.

The classification adopted in the final analysis thus reflects some of the features of Vadász', Valeton's, De Weisse's, Kirpal's and Bushinsky's classifications and of the Soviet classification of 1973. From each, the writer has gleaned the factors that he deemed to be the most realistic. The principal difference is that all bauxites overlying carbonate rocks have been united into a single main group, as in the writer's opinion the carbonate bedrock plays a decisive role in the process of formation, and greatly affects the material composition of these deposits. The principal depositological groups cannot, however, be sharply isolated from one another. Although rather few and far between, there do exist occurrences transitional between karst and lateritic bauxites, on the one hand, and karst and Tikhvin-type deposits, on the other. There is a separate discussion of them in the chapter on depositology.

The author's calculations show 85% of the bauxites currently known on the Earth to be lateritic; 14% is karst bauxites, and 1% is Tikhvin-type bauxite.

According to a calculation by Wedepohl (1969), the total bulk of sedimentary rocks on the globe, for an average thickness of 1200 m, is $1{\cdot}5 \times 10^2$ tons*. Average sedimentary sequence thickness was assumed to be 900 m by Goldschmidt (1933) and 1850 m by Ronov (1967). The writer has calculated the total mass of all bauxites, clayey bauxite and bauxitic clay currently known at round 45×10^9 tons. This is less than 0·001% of all sedimentary rocks if Wedepohl's figure is accepted. Bauxite, then, is a *rara avis* in sedimentary sequences: its formation and preservation require the combination of some unusual circumstances.

* All tons in this book are metric unless stated otherwise.

2. GEOGRAPHIC AND STRATIGRAPHIC DISTRIBUTION

2.1. GEOGRAPHIC DISTRIBUTION

The writer has, on the evidence currently available distinguished 88 karst bauxite regions, and has assigned a geographical term and a serial number to each (Appendix 1). The geographic positions of these regions are shown in figures 1 to 4. The number of deposit groups within the bauxite regions ranges from one to eleven. A total of 191 bauxite deposit groups is known.

In Europe, the great majority of all karst bauxite regions are rather close to the northern shore of the Mediterranean: their totality is accordingly called the *Mediterranean bauxite belt* (fig. 1, separate, at the and of the book). The northern border of the Mediterranean belt is marked by the bauxite regions of Hungary, Austria, Slovakia, the Carpatho-Ukraine, Rumania and the Crimea. The bulk of karst bauxites is in the mid-zone area, in Greece and Yugoslavia. There are, in addition, important bauxite regions in Southern France. No bauxites are known from the southern shore of the Mediterranean, if the occurrence of Maktesh Ramon in Southern Israel is not accepted as such.

The zone ends in Eastern Turkey. Farther east, there follows a zone of some 700 km which is barren of bauxites on currently available evidence. Bauxites reappear in Northwestern Iran, whence they can be traced through Afghanistan and Pakistan to Kashmir. They constitute the *Irano–Himalayan bauxite belt* which, in the geological respect, can be considered the eastern continuation of the Mediterranean belt (fig. 1).

Karst bauxites on the islands of the Western Pacific are united into a *Pacific bauxite belt* (fig. 2). The bauxite regions of Eastern Asia from Campuchea and Vietnam to Northeastern China constitute the *East Asian bauxite belt*. The *Uralo–Sibero–Central Asian bauxite belt* comprises the important bauxite regions of the Asian and East European parts of the USSR (fig. 3). Of all the karst bauxite belts, this is the one which reaches farthest north, beyond the 76th parallel.

For the karst bauxite regions of the Caribbean islands, Jamaica, Cuba and Hispaniola, the author has coined the term *Caribbean bauxite belt*. Finally, some minor karst bauxite deposits, known from the southeastern US, constitute the *North American belt* (fig. 4).

Separate from the above-listed belts is the Tulsk–Ryazansk group of deposits to the south of Moscow. As the deposits of this group are genetically rather different from the other karst bauxites, the writer has deemed it right to separate them geographically as well.

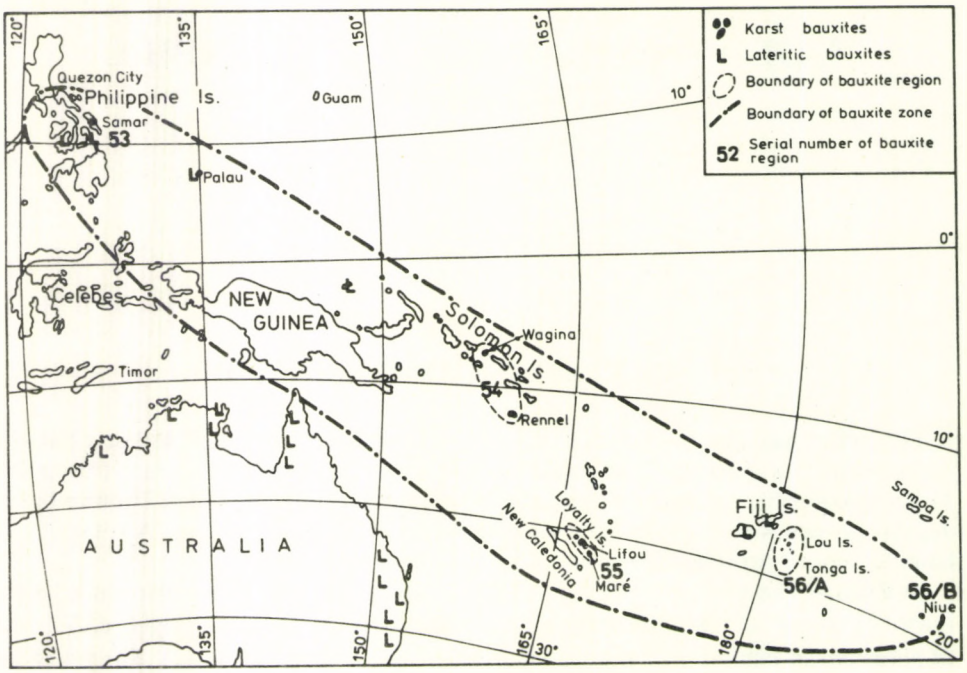


Figure 2. Bauxite deposit groups of the Pacific bauxite belt

The karst bauxite belts occupy rather limited portions of the globe (fig. 5, separate, at the end of the book). Their great majority is found on the northern hemisphere, between the 10th and 70th parallels. The only belt on the southern hemisphere (between the 10th and 25th parallels) is the Pacific belt. The map also shows the Tikhvin-type and laterite bauxite occurrences. The most extensive lateritic bauxite regions (Guyana Shield, Southeastern Brazil, West Africa, India and Australia) are situated south of the karst bauxite zones, on either side of the equator.

There are minor lateritic bauxite deposits also in the karst bauxite belts or in their vicinity. The Mediterranean belt, for example, includes the deposit of Locris in Greece and a group of deposits in Southern Turkey. There are quite a few minor laterite bauxite deposit groups to the north of the belt: Decazeville in France, Antrim in Northern Ireland, Ayrshire in Scotland, Vogelsberg in Federal Germany, Northern Moravia in Czechoslovakia, Nova Ruda and Lublin in Poland, and several deposit groups in the Ukraine. Of the latter, only the one at Kursk is of a significant size (Bielgorod).

No lateritic bauxite is known from the Irano–Himalayan karst bauxite belt. In the Pacific belt, lateritic bauxite has been identified on Fiji, and lateritic nickel ore on New Caledonia. No lateritic bauxite is known from the East Asian belt, either, but minor lateritic bauxite deposits have been identified to the southeast of it, in Fukien province, China and in South-Vietnam. In the Uralo–Sibero–Central Asian karst bauxite belt, some lateritic bauxite deposits are known along the Yenisei, in the Gissarsk Mountains, at Mugodjar, in the Southern Urals, and in the central Timan Mountains.

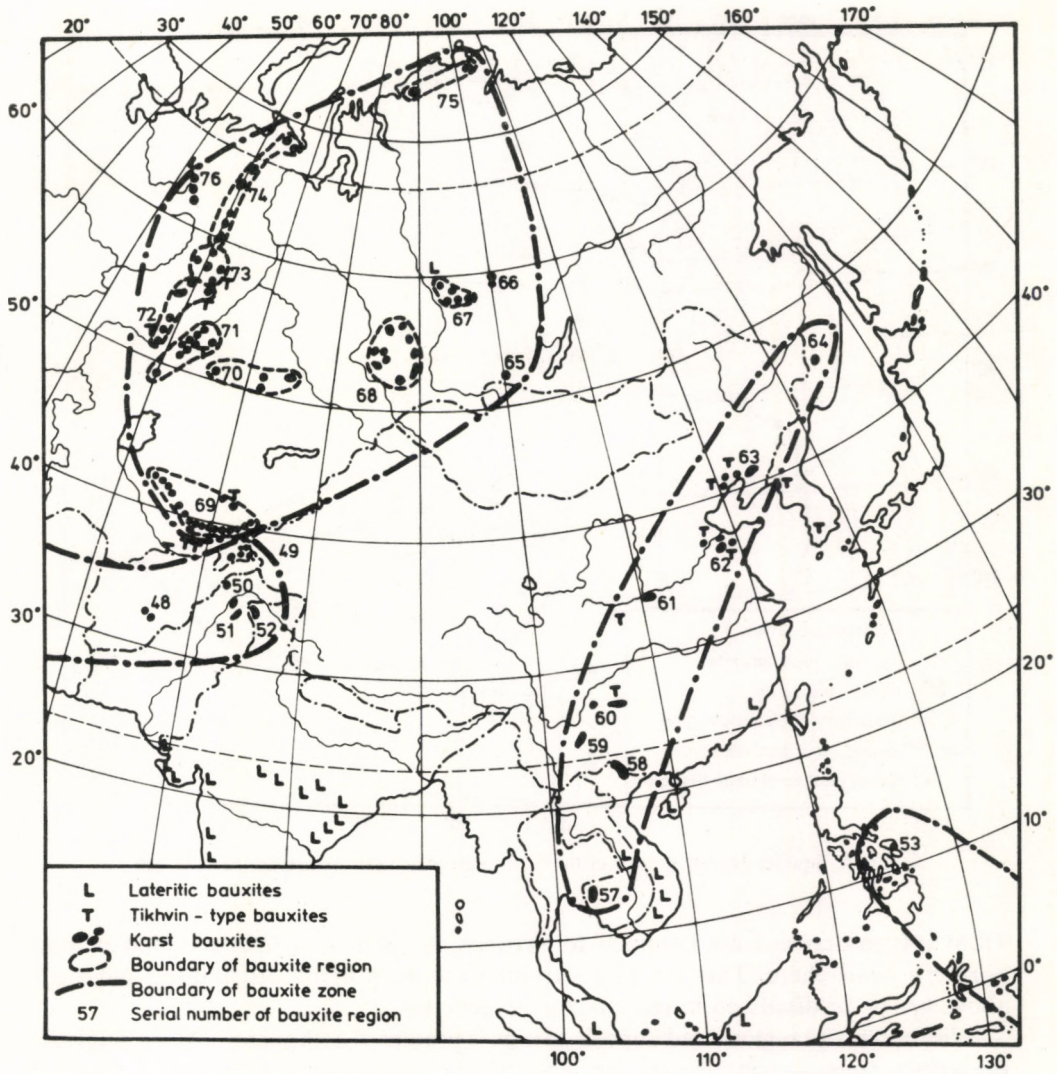


Figure 3. Bauxite deposit groups of the East Asian and Uralo–Sibero–Central Asian bauxite belt

No lateritic bauxite is known from the Caribbean belt, but there are laterites rich in iron and nickel in Cuba. Nor is there any lateritic bauxite in the Northern American karst bauxite belt, but the comparatively large lateritic bauxite occurrence of Arkansas is situated immediately to the west of it.

The most important deposits of *Tikhvin-type*, including the eponym are situated in the Tikhvin–Severoonezhsk region east of Leningrad and in the Ukraine. All other occurrences of Tikhvin-type bauxites are associated with karst bauxite zones. The most important of these are in the Southern and Central Urals, in the Gissarsk Mountains, and in the Southeastern US. One may summarise that the Tikhvin-type deposits are not so remote geographically from the karst bauxites as are most lateritic bauxites.

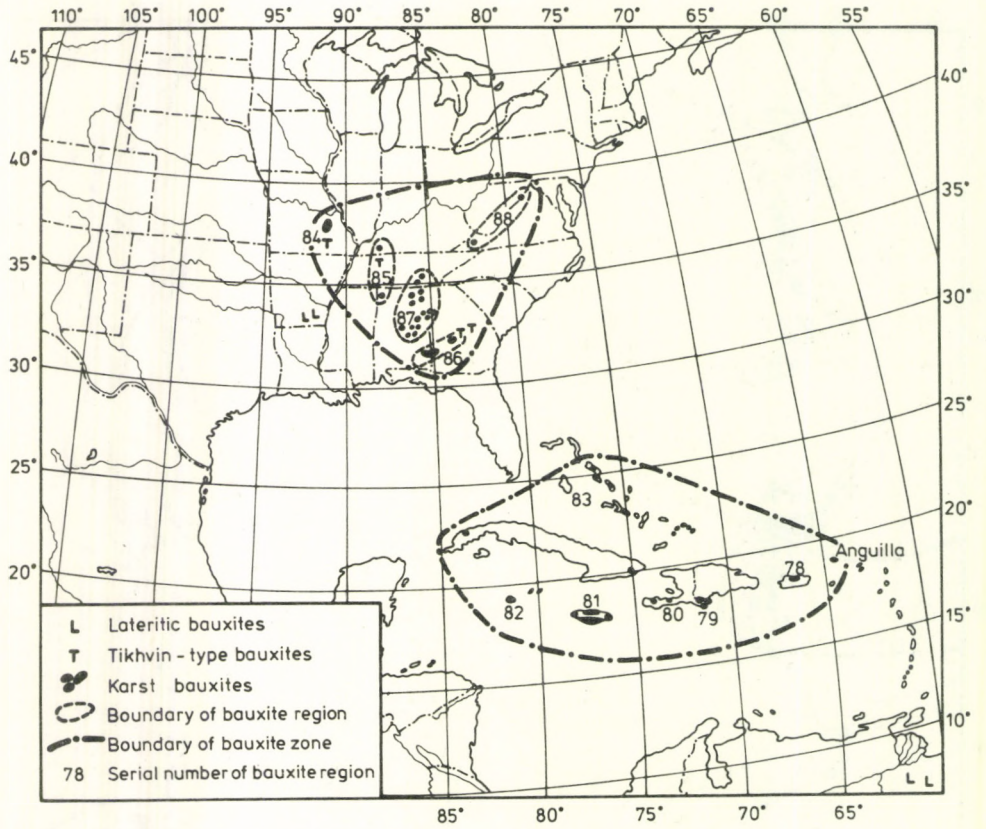


Figure 4. Bauxite deposit groups of the Caribbean and North American bauxite belt

Tikhvin-type deposits are confined to between the 30th and 65th parallels of the northern hemisphere. The factors controlling the geographic distribution outlined above are to be discussed in the chapter on genetics.

A list of bauxite regions and deposit groups is presented as Appendix 1. It also gives the bauxite tonnage contained in the deposits. The figures do not refer to the quantity of bauxite amenable to industrial processing ('ore reserves'), but to the total bulk of bauxitic rock contained in the deposits, including bauxite already extracted. These figures of an emphatically geological significance, then, express the total amount of more or less thoroughly bauxitised rock. The reliability of the tonnage estimate varies widely from one deposit group to the next, being a function of the degree in which the deposits have been prospected. These uncertainties notwithstanding a comparative assessment starting from bauxite tonnage permits a weighting and an estimation of the frequency of different features on a more realistic basis than heretofore. It has, notably, been found by the writer that the authors of most of the previous comprehensive treatises had paid much less attention to the bauxites, unfamiliar to them, of the remoter regions than to nearby deposits familiar to them, however small those latter may happen to be.

Table 1. Bauxite tonnage in the major world karst bauxite regions.

Serial number	Karst bauxite regions	Million tons
1	Jamaica	1200
2	Central Greece	415
3	Transdanubian Hills, Hungary	401
4	Provence, France	325
5	Kweichow, China	300
6	Dalmatia, Yugoslavia	295
7	Turgay Plateau, USSR	270
8	Shantung, China	250
9	Northern Urals, USSR	220
10	Crna Gora, Yugoslavia	290
11	Languedoc, France	185
12	Northern Vietnam	200
13	Central Kazakhstan, USSR	191
14	Timan Mountains, USSR	180
15	Yunnan, China	130
16	Dominican Republic	120
17	Liaoning, China	100
	Altogether	5072

Table 2. Karst bauxite tonnage distribution over the bauxite zones of the world.

Serial number	Bauxite zone	Million tons	Per cent
1	Mediterranean	2533	40
2	Irano–Himalayan	128	2
3	Pacific	122	2
4	East Asian	1014	16
5	Uralo–Sibero–Central Asian	1173	17
6	Caribbean	1390	22
7	North American	37	1
	Altogether	6397	100

The 17 principal karst bauxite regions of the globe are listed in table 1 in order of decreasing tonnage.

These 17 bauxite regions make up 79% of the total tonnage of karst bauxites (6397 million tons): the remaining 21% is made up by the other 71 regions. The figures in the table may strike a bauxite specialist as wildly wrong. It is therefore necessary to point out once again that they do not state the industrial-grade bauxite reserves, but the total quantity of bauxite in a region, including clayey and other sub-grade types.

Jamaican bauxites alone make up 19% of the known global karst bauxites. The distribution of bauxite tonnage over the seven bauxite belts is presented in table 2.

The belt with the greatest bauxite tonnage is the Mediterranean: it is regarded not without reason as the *locus classicus* of karst bauxite development. The runner-up is the Caribbean zone, dominated by the Jamaican bauxites. The East Asian and the Uralo–Sibero–Central Asian belts are of approximately equal weight. The remaining three belts contain but 5% of global karst bauxites.

The areal extent of a bauxite region is more or less independent of the bauxite tonnage it contains. In places, a tremendous bauxite reserve may concentrate in a pretty small area (e.g. Jamaica); in others, less bauxite may be spread rather thinly over a much vaster area (e.g. the Central Asian bauxite region).

2.2. STRATIGRAPHIC DISTRIBUTION

Appendix 1 also lists the age of the bauxite and of the rocks immediately under- and overlying it for each group of deposits. The reliability of this information is rather inhomogeneous, depending on the degree to which this or that region has been explored. Reliability is assigned by one of three degrees (proven, hypothetical, questionable) in Appendix 1.

Table 3 presents the distribution of karst bauxites currently known over the geological periods.

The bulk of karst bauxites is seen to be Tertiary and Quaternary; Mesozoic bauxites are scarcer, Paleozoic ones even more so. Absolute dating has revealed the Paleozoic to have lasted for 345 million years, the Mesozoic for 160, and the Tertiary and Quaternary taken together for 65 million years; it thus appears that the younger and shorter periods of Earth history include disproportionately more bauxite. These figures, however, express not the rate of formation of bauxite over the geohistorical periods, but the amounts preserved up to this day. The strata of the Paleozoic, together with the bauxites they may have contained, have been eroded to a much greater extent than the younger strata. This also is true, albeit to a less extent, of the Mesozoic and Paleogene bauxites. The erosion of Neogene and recent deposits has been much less severe. The geological evidence, currently available does not permit an estimate of any accuracy of the amount of bauxite thus eroded. It seems nevertheless likely that the rate of karst bauxite formation has increased rather substantially from the Proterozoic to Neogene.

The oldest karst bauxites currently known date from the *Cambrian* or, indeed, from the *late Proterozoic* according to some authors. The bauxites of this period include, in addition to the Boksonsk group of deposits in the Eastern Sayan Mountains, traces of ore in the Kozyrevsk region, and minor, partly metamorphosed deposits in the Alatau of Kusnetzk and in the Gornaya Soriya region (Mikhailov *et al* 1973).

The only known Ordovician occurrence (the one near Tselinograd in the Kazakhstan) is of insignificant size (Dolgopolov *et al* 1971).

No karst bauxite deposit is known from the *Silurian*.

Karst bauxite formation takes an abrupt increase in the *Devonian*. Significant groups of deposits are known at a number of places in and between the Southern and the Polar Urals, right up to the shores of the Arctic Sea. Further deposits are known at the Altai

Table 3. Age distribution of world karst bauxite tonnage.

Epoch	Geological period	Bauxite tonnage		Cumulative bauxite tonnage	
		million tons	per cent	million tons	per cent
Kainozoic	Quaternary	122	2	2418	38
	Miocene–Pliocene	1500	24		
	Oligocene	20	<1		
	Paleocene–Eocene	776	12		
Mesozoic	Upper Cretaceous	1501	23	2381	37
	Lower Cretaceous	313	5		
	Jurassic	315	5		
	Triassic	252	4		
Paleozoic	Permian	381	6	1598	25
	Carboniferous	780	12		
	Devonian	406	6		
	Silurian	–	–		
	Ordovician	1	<1		
Precambrian	Cambrian and Upper Proterozoic	30	<1		
	Altogether	6397	100	6397	100

Range near Berdsk and Obukhovo, in the Central Timan Mountains, and in Pakistan (in the Hindu Kush range).

Karst bauxites are even more abundant in the *Carboniferous* (Timan Mountains, the Taimyr Peninsula, the Aliysk and the Turkestanian ranges, and, in the western continuation of the latter, the Nuratau, Bukantau and Tamdytau ranges). Most of the Chinese karst bauxites are Carboniferous, and so is the occurrence in the State of Missouri in the US.

Permian karst bauxites are much less abundant. The most important occurrences are those of the Taurus Range in Turkey and some of the deposits in Vietnam and of Kweichow province in China. A minor deposit is known in Campuchea. Karst bauxites formed on the Permo–Triassic boundary are known in Northern and Northeastern Iran, and in the western part of the Taurus Range. The bauxites of the nearby islands, Samos and Naxos, are probably of the same age.

The known quantity of *Triassic* karst bauxites is even less. Bauxitisation is known to have taken place in Iran, extending also to the nearby Lesser Caucasus, Pamir and Southern Ferghana. There are deposit groups of significant size on the Balkan Peninsula, from Slovenia to the Albanian Alps. Finally, minor deposits of Upper Triassic age are known in the Northern Urals (from the vicinity of Bogoslovsk and Volchansk).

Jurassic karst bauxites are also comparatively scarce. The largest groups of deposits are those of Crna Gora (Montenegro) in Yugoslavia, and of the Parnassos, Kiona and Helicon ranges of Greece. There are minor occurrences on the islands of Euboea, Skopelos and Amorgos, on the Chalkidike Peninsula, on the Istrian Peninsula near Rovinj, in Slovenia (in the Hrušica Mountains), in the Swiss and French Alps, in the

Betic Cordillera of Spain, in the Northeastern Carpathians, in the Crimea, and in the Chimkent district near Tashkent (USSR).

In the *Lower Cretaceous*, bauxite formation expanded substantially, to cover the entire Mediterranean zone. Karst bauxites were formed in Portugal, in Northeastern Spain, in Southern France, on Sardinia. In Hungary, the bauxite deposit groups of Nagyharsány and Alsópere belong to this period. Most of the Rumanian deposits are Lower Cretaceous, and so are those of Bosanska Krupa and Vlasenica in Yugoslavia, and those around Eleusis and Mandra in Greece. Minor Lower Cretaceous bauxite deposit groups in the USSR include those of the Salair–Barzassk region, of the Chimkent and Kansaisk districts, of Mugodjar, and of the Southern and Central Urals.

The *Upper Cretaceous* was another peak period of karst bauxite formation. Most of the deposit groups of the Spanish Pyrenees, of the Languedoc and the Provence, and all deposits of Italy, Austria and Slovakia are of that age. In Hungary, the Halimba bauxite is Upper Cretaceous. Also belonging to this period are the Sohodol-Cîmpeni group of deposits in Rumania and, in Jugoslavia, the bauxites of the Grmeć Mountains, of Jajce and of the Grebnička Planina and some of those in Crna Gora (Montenegro). Sizeable deposits came to exist in the Parnassos, Kiona and Helicon ranges of Greece. There is a major Upper Cretaceous group of deposits in Turkey. Minor ones are known from the Zagros Range in Iran and from Israel (Maktesh Ramon). In the USSR, there are minor Upper Cretaceous deposits on the Turgay tableland, along the Angara River, and in the Chadobetsk district.

In the *Plaeocene* and *Eocene*, significant quantities of bauxite came to be formed in the Transdanubian Hills of Hungary, on the Istrian Peninsula, and on the Dinaric karst. Minor occurrences are known from Catalonia (from the Ebro Massif), from the Peloponnese Peninsula in Greece, from Albania, from near Sütlegen in Turkey, from Pakistan and from Kashmir. There are large groups of deposits in the Kazakhstan, in district of Arkalyk and in the region along the Yenisei. This is when the comparatively minor deposits in the Southeastern US were formed.

Hardly any karst bauxites are known from the *Oligocene*. Those that came to be accumulated in this period include the minor deposits of Northern Slovenia and some quite small ones near Jajce (Yugoslavia). In Hungary, it is the redeposition of bauxites in secondary deposits (Óbarok–Újbarok, Vázsonypuszta) that took place in the Oligocene. Finally, an Oligocene age is conjectured for the small deposits of the Salento Peninsula in Italy.

The *Miocene* and *Pliocene* are once again important periods of karst bauxite formation. This is when the major deposits of Jamaica, Haiti and the Dominican Republic came to exist. Some Vietnamese bauxites were also redeposited during the Neogene into secondary deposits.

The bauxites of the Pacific islands are post-*Quaternary*. The formation of karst bauxite has not ceased there up to this day.

The seven karst bauxite belts are distinct not only geographically but also as to the age of the bauxite formation (table 4).

In the Mediterranean belt, the oldest karst bauxite is Permian: the youngest one is Miocene. The largest quantity of bauxite formed during the Upper Cretaceous. In the Irano–Himalayan belt, karst bauxite formation started in the Devonian but continued to the middle Eocene only. Its bulk is older than in the Mediterranean zone: more than

Table 4. Age distribution of bauxite tonnage in the karst bauxite zones of the globe.

Geological age	Mediterranean		Irano–Himalayan		Pacific		East Asian		Uralo–Sibero Central Asian		Caribbean		North American	
	zone bauxite													
	10^6 ton	per cent	10^6 ton	per cent	10^6 ton	per cent	10^6 ton	per cent	10^6 ton	per cent	10^6 ton	per cent	10^6 ton	per cent
Quaternary	–	–	–	–	122	100	–	–	–	–	–	–	–	–
Miocene–Pliocene	30	<1	–	–	–	–	80	8	–	–	1390	100	–	–
Oligocene	20	<1	–	–	–	–	–	–	–	–	–	–	–	–
Paleocene–Eocene	491	18	30	23	–	–	–	–	238	20	–	–	17	46
Upper Cretaceous	1179	46	11	9	–	–	1	<1	310	26	–	–	–	–
Lower Cretaceous	237	10	5	4	–	–	–	–	71	6	–	–	–	–
Jurassic	300	13	10	8	–	–	–	–	5	<1	–	–	–	–
Triassic	193	8	46	36	–	–	–	–	13	1	–	–	–	–
Permian	83	3	25	20	–	–	273	26	–	–	–	–	–	–
Carboniferous	–	–	–	–	–	–	660	66	100	8	–	–	20	54
Devonian	–	–	<1	1	–	–	–	–	405	35	–	–	–	–
Silurian	–	–	–	–	–	–	–	–	–	–	–	–	–	–
Ordovician	–	–	–	–	–	–	–	–	1	<1	–	–	–	–
Cambrian	–	–	–	–	–	–	–	–	30	3	–	–	–	–
Altogether	2533	100	128	100	122	100	1014	100	1173	100	1390	100	37	100

half of the Irano–Himalayan bauxites are Permian and Triassic; Upper Cretaceous bauxites are subordinate.

In the Pacific belt, bauxites are exclusively post-Quaternary. In the East Asian belt, on the other hand, bauxite formed predominantly during the Carboniferous, and to a lesser extent during the Upper Cretaceous. The Miocene and Pliocene merely saw the secondary redeposition of Paleozoic bauxites (Vietnam).

The Uralo–Sibero–Central Asian belt exhibits a great variety of bauxite ages. This is the only belt featuring Cambrian and Ordovician bauxites. Bauxite formation gained momentum in the Mesozoic, to attain a maximum in the Upper Cretaceous, and to cease in the Eocene.

The bauxites of the Caribbean belt are Miocene and Pliocene. In the North American belt, karst bauxites formed in two isolated episodes: during the Carboniferous and during the Paleocene–Lower Eocene.

In summary, the geohistorical distribution of bauxite formation is similar between the Mediterranean and the Irano–Himalayan belts only; the other belts differ essentially from these two. In some zones, there was only one period of bauxite formation, whereas in others bauxite was a recurrent feature over millions of years. All this was controlled by the geological–geotectonic evolution of the belts in question, to be expounded in more detail in the chapter on genetics.

3. DEPOSITOLOGY*

3.1. CLASSIFICATION OF KARST BAUXITE DEPOSITS

Karst bauxites used to be regarded as a uniform group by most authors: for a long time, no attempts were made at subdividing them into types according to the nature of the deposits. It was Bushinsky (1964) who first subdivided karst bauxites into 'proximal' ('near-karst') or 'contact' types on the one hand and 'distal' ('remote-karst') or 'areal' types on the other. According to Bushinsky, the deposits of the first group occur close to where an alumosilicate-type rock is in contact with a karst area. The karst depressions are filled with clayey-sandy sediments; it is in these sediments that the bauxite deposit is embedded. Bushinsky's second type occurs in extensive karst areas, where karst depressions are largely or entirely filled with bauxite. Bushinsky retained this twofold division also in his book published in 1971, but he occassionally called the areal type the 'Mediterranean type' there. In a paper published in 1970, he distinguished a third type of karst bauxites, represented by the deposits of Tulsk and Ryazansk, formed by infiltrating ground water, in connection with the oxidation of pyrites. In his book of 1971, on the other hand, he had already moved this third type into his main group of sedimentary–chemogenic deposits. Sapozhnikov (1975) distinguished monocyclic and polycyclic karst bauxites: these correspond by and large to Bushinsky's areal and proximal type, respectively. Pastukhova (1975), subdivided bauxites overlying carbonate rocks into a sedimentary and a sedimentary–lateritic group, according to the sites of bauxitisation. In her opinion the former deposits are due to the redeposition of bauxite previously formed, whereas the latter are due to *in situ* bauxitisation. Within the second group she distinguished subgroups of 'mature' or 'homogeneous' and 'immature' or 'motley' composition.

According to the writer's own experience, karst bauxite deposits are so manifold that their subdivision (by depositological criteria) is entirely justified. Retaining Bushinsky's three principal types, he has added two further ones (Bárdossy 1970b), replacing at the same time Bushinsky's descriptive terminology by a geographical one, of the sort employed in stratigraphic nomenclature. Thus Bushinsky's proximal type became the '*Kazakhstan type*'; his distal type kept the name '*Mediterranean type*' coined by Bushinsky himself. The present writer also retained Bushinsky's term '*Tulsk type*', and named the two types newly introduced by himself the '*Ariège type*' and the '*Salento type*', respectively. These latter types are distinguished by different de-

* Here and in the following, I am using this term as an English equivalent to the useful German term 'Lagerstättenkunde' (science of mineral deposits).

positological features, which permit a more unequivocal distinction than any classification on a pure genetic basis.

After examination of the most recent literature and consultation with a number of bauxite specialists, the writer has found it necessary to distinguish one more category by the term '*Timan type*', a group to which he has transferred some of the deposits of the Kazakhstan type. The six types of bauxite deposits enumerated above are not sharply distinct: there are numerous deposits intermediate between the pure types. Furthermore, there are also deposits transitional between karst and lateritic bauxites, on the one hand, and karst and Tikhvin-type bauxites on the other.

In Appendix 1, the individual bauxite regions are characterised either as depositologically pure types or as transitions: the symbol 'Med > Kaz', for example, means a Mediterranean-type deposit with some features transitional towards the Kazakhstan type. A calculation of the bauxite tonnage contained in each of the above six types of deposits gave the results stated in table 5.

An overwhelming majority of all karst bauxite deposits, then, belongs to the Mediterranean type. This is why specialised literature in the West used to be concerned almost exclusively with the description of this type, which it considered, as a matter of fact, to be the only existing form of karst bauxite.

Table 5. Bauxite tonnage distribution over deposit types.

Serial number	Deposit type	10^6 tons	Per cent	Number of deposit groups
1	Mediterranean	4457	70	145
2	Timan	1032	16	14
3	Kazakhstan	636	9	20
4	Ariège	142	2	4
5	Salento	130	2	7
6	Tulsk	1	<1	1
	Altogether	6398	100	191

Table 6. Distribution of deposit types over the bauxite zones with respect to tonnage.

Serial number	Karst bauxite zones	Mediterranean		Timan		Kazakhstan		Ariège		Salento	
		number of deposit	10^6 ton	number of deposit	10^6 ton	number of deposit	10^6 ton	number of deposit	10^6 ton	number of deposit	10^6 ton
I.	Mediterranean	95	2341	–	–	–	–	4	142	5	50
II.	Irano–Himalayan	15	128	–	–	–	–	–	–	–	–
III.	Pacific	7	122	–	–	–	–	–	–	–	–
IV.	East Asian	3	123	6	810	1	1	–	–	2	80
V.	Uralo–Sibero–Central Asian	14	353	7	212	15	608	–	–	–	–
VI.	Caribbean	11	1390	–	–	–	–	–	–	–	–
VII.	North American	–	–	1	10	4	27	–	–	–	–
	Altogether	145	4457	14	1032	20	636	4	142	7	130

The reserves contained in the last three types of deposit are insignificant, but their distinction is amply justified nevertheless by their peculiar depositological features.

The distribution of the above-named types of bauxite (the Tulsk type excluded) over the karst bauxite zones of the Earth is presented in table 6.

The Tulsk type is omitted from this table because, separate spatially from the first five types, it fits into none of the geographic bauxite belts.

The Caribbean, Pacific and Irano–Himalayan belts contain exclusively Mediterranean-type deposits. Most deposits of the Mediterranean belt are, of course, also of Mediterranean type. These four belts, then, are highly similar among them as to depositological features. The remaining three belts are also similar among them in that they include all the deposits of the Timan and Kazakhstan types, with a preponderance of Timan-type deposits in the East Asian belt and of Kazakhstan-type deposits in the Uralo–Sibero–Central Asian belt. It is not fortuitous that Tikhvin-type deposits are also found in these last three belts.

Comparing the geographic distribution of the above-defined two principal belt-groups, the four belts where Mediterranean-type bauxites are predominant are seen to lie to the south of the other three (fig. 5) ones.

The principal depositological features of each of the six main types of bauxite deposits are outlined in succession in the sections below.

3.2. MEDITERRANEAN-TYPE DEPOSITS

These deposits are fairly homogeneous as to lithology. In extreme cases, the entire deposit may consist of high-grade bauxite. Most often, however, high-grade bauxite passes into clayey bauxite or bauxitic clay towards the top, the bottom and the edges of the deposit. More seldom, these deposits contain interbeddings of authigenic limestone and/or layers of more or less coarse detritus, derived from the carbonate rocks of the immediate surroundings. Any other alien material in the deposits is exceptional and insignificant in volume.

These are the karst bauxite deposits in the strict sense: all other types of deposit are, in the final analysis, but transitions of one sort or another towards the other two main depositological groups (the Tikhvin-type and the lateritic-bauxite deposits).

Table 6 reveals deposits of Mediterranean type to be known from all karst bauxite belts except the North American. The largest numbers of Mediterranean-type deposits, however, occurs in the Mediterranean belt.

3.2.1. GEOMETRY

Mediterranean-type deposits are clear-cut as to outline, but their sizes and shapes exhibit great variety. In the writer's experience, eight basic forms can be distinguished, but there are all sorts of transitions between any two of these. Transitions from the stratiform to the lenticular and from the lenticular to the sinkhole-type form are especially abundant. In Appendix 1, the deposit forms encountered in each of the groups distinguished are indicated in a separate column, in order of the relative abundances of forms.

In the following, we shall be concerned with the morphologic features of the bauxite deposits as a whole: their internal petrographic constitution is to be discussed in the chapter on petrography. The eight basic types of deposits will be reviewed in decreasing order of their typical sizes and bauxite tonnages.

3.2.1.1. STRATIFORM DEPOSITS

These are the karst bauxite deposits of the largest size. Their length and width is greater than, or only slightly less, than one kilometre; their area tends to exceed 1/2 km^2. The bedrock of the deposit is karsted, non-uniform, with undulations of 1 to 2 m amplitude, interspersed with dolines 5 to 10 m deep and with rather steep prominences. The bauxite may be missing altogether from above these latter. The bauxite top is smooth and comparatively plane. The thickness of the bauxite is 1 to 40 m, depending on bedrock morphology. Such a deposit may contain 10 to 70 million tons of bauxite and bauxitic rock.

Deposits of this type are few and far between; they are confined to the richest bauxite regions. The largest stratiform deposit of Hungary is in the Halimba Basin. It covers an area of about 7 km^2 (fig. 6), with a north–south extent of 3·5 km and an average east–west extent of 2 km. The bauxite layer is interrupted in places by smallish barren spots, where prominences of the bedrock with no bauxite above them pierce the deposit. In most places, the deposit is 5 to 10 m thick: greater thicknesses, up to 35 m, correspond

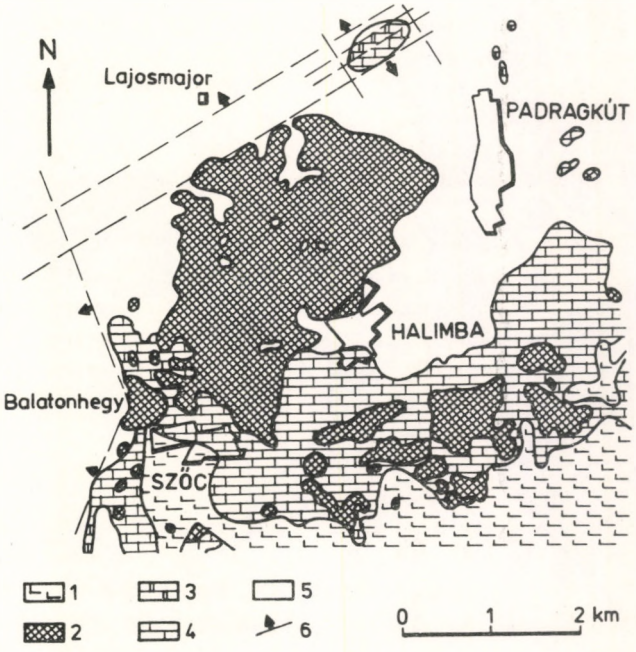


Figure 6. Stratiform bauxite deposit, Halimba, Hungary

(1) dolostone and limestone, Upper Triassic; (2) bauxite; (3) limestone, Senonian; (4) limestone, Eocene; (5) post-Eocene deposits; (6) principal faults

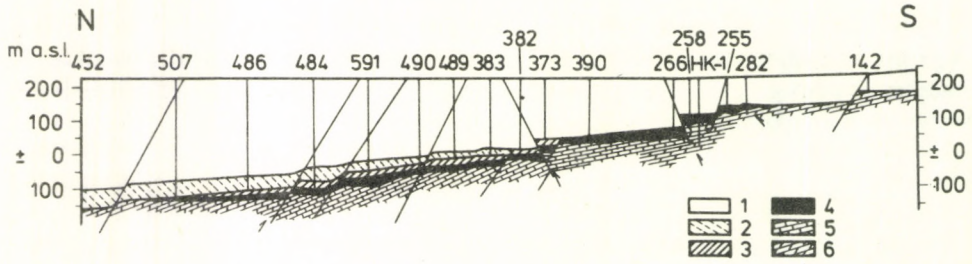


Figure 7. Dip profile of Halimba Basin, Hungary. After Erdélyi (1965)

(1) Eocene beds; (2) Senonian strata with coal beds; (3) multicoloured marl with interbedded conglomerates, Senonian; (4) bauxite, clayey bauxite and bauxitic clay; (5) 'Dachsteinkalk', Upper Triassic; (6) Hauptdolomite, Upper Triassic

to depressions in the bedrock surface, as illustrated by a geological profile (fig. 7). The deposit is dissected by a number of small normal and thrust faults. Even so, the almost totally plane deposit top is conspicuous, smoothing out as it does all the ups and downs of the bedrock relief.

At Iszkaszentgyörgy, the Kincses, József and Rákhegy ore bodies in effect constitute a single deposit despite their highly irregular outline. Their aggregate area is 2·6 km^2. The Bitó deposit nearby covers 1·5 km^2 (fig. 8). At Gánt, the Bagolyhegy deposit has an area of 1·3 km^2; the Harasztos–Újfeltárás deposit, of 1·0 km^2. Bauxite used to be

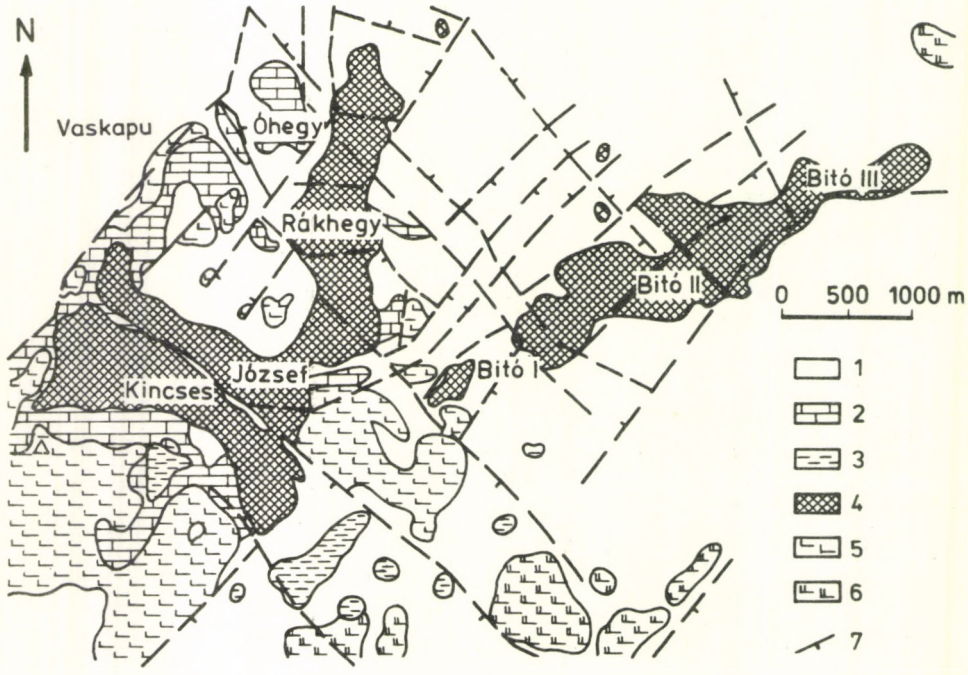


Figure 8. Stratiform bauxite deposits, Iszkaszentgyörgy, Hungary

(1) post-Eocene strata; (2) limestone, Eocene; (3) bauxitic and bauxite-like clay; (4) bauxite and clayey bauxite; (5) Hauptdolomite, Carnian–Norian; (6) dolostone, Ladinian; (7) principal faults

thickest in this latter (10 to 25 m on average; up to 40 m locally). At both Halimba and Iszkaszentgyörgy, the bauxite deposits taper away gradually towards the edges. At Gánt, on the other hand, they are delimited on some sides by abrupt fault scarps (fig. 9) of the dolostone bedrock.

In France, the Villeveyrac Basin to the southwest of Montpellier contains a stratiform bauxite deposit of some 5 km^2 area, mostly 4 to 8 m thick. The southern part of the Bédarieux Basin contains an uninterrupted layer of bauxite about 3 km long and 500 to 600 m wide, of a thickness of 5 to 10 m in most places and 35 m as a maximum (Combes 1973). In Yugoslavia, the Nazda–Podbračan deposit in the Vlasenica region has an area of about 0·7 km^2: its thickness is 10 m on average, 40 m as a maximum.

Stratiform deposits much thinner than the above ones (2 to 4 m thick on average) are known from Northwest Iran (from the Bukan region) and from Northern Iran (near Jajarm). Their most characteristic morphologic feature is a hardly karsted bedrock surface.

Devonian bauxites in the Northern Urals also form stratiform deposits in places. Typically, these are 1 to 5 m thick, attaining 10 to 25 m above the occasional karst depressions only. Most of these depressions are of 30 to 80 m diameter; their slopes

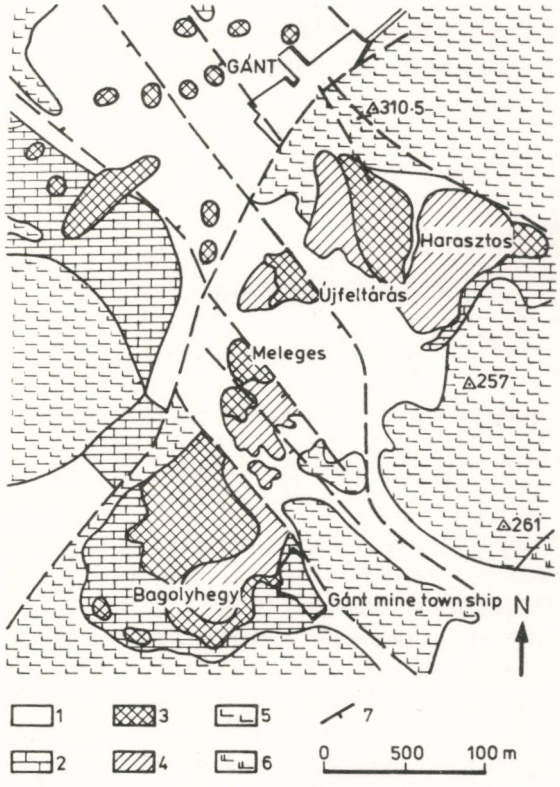


Figure 9. Stratiform bauxite deposits, Gánt, Hungary

(1) post-Eocene strata; (2) limestone, Eocene; (3) bauxite, clayey bauxite, bauxitic clay; (4) worked-out bauxite; (5) dolostone, Upper Triassic; (6) dolostone, Middle Triassic; (7) principal faults

rarely exceed 15° (Bushinsky 1971). The deposits themselves are of 1 to 5 km^2 area. Insular prominences of bedrock with no bauxite above them are fairly frequent. The full extent of the deposit group is unknown as yet: the entire complex dips east at angles from 20 to 30°; bauxite was traversed even in marginal boreholes 1200 to 1400 m deep (Kirpal' and Tenyakov 1974). The deposits are often delimited in the striking direction by pre-bauxitisation ('preforming') fault scarps (Gutkin and Rodchenko 1965).

3.2.1.2. BLANKET DEPOSITS

In areal extent, these deposits are equal to, or may indeed be larger than, the stratiform ones. This group includes some quite young. Pliocene and Quaternary deposits with no sedimentary cover. The bauxite overlies a limestone bedrock blanket-like, covering its gentle hummock-shaped rises as well as its depressions. Accordingly, the deposit-top surface copies the bedrock relief rather than being more or less plane. Another important difference is the much smaller thickness of which the bauxite tonnages contained in any one deposit do not as a rule exceed 5 to 20 million tons.

In Jamaica, extensive blanket deposits occur on the St Ann Plateau, in the St Elizabeth Plain and within the Mile Gully graben. There are some blanket deposits also in the southern part of the Manchester Plateau. Bauxite thickness is from 4 to 10 m typically, and up to 20 m over the karst depressions of the bedrock. Of 1 to 5 km^2 size as a rule, these deposits have highly irregular outlines.

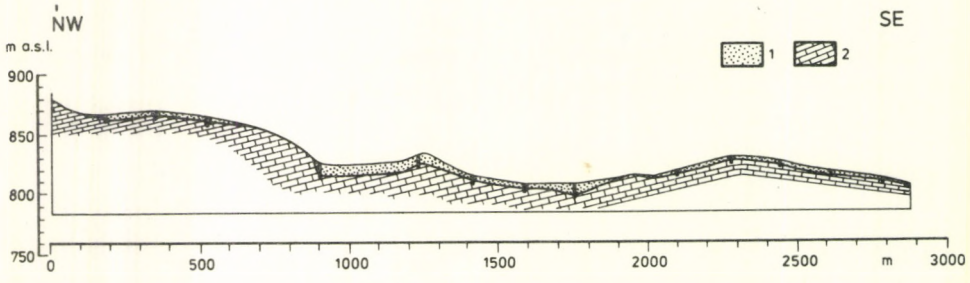


Figure 10. Profile of a blanket-like bauxite deposit. Haiti, Rochelois Plateau. After Goldich and Bergquist (1948)

(1) bauxite; (2) limestone, Eocene

The Rochelois Plateu in Haiti has blanket deposits of complicated outline, of 1 to 3 km^2 area (fig. 10). Not thicker than 1·5 to 2 m in most places, the bauxite attains 10 m only above some dolines in the limestone bedrock. Barren spots of more or less the same size, corresponding to prominences of the bedrock, are also fairly frequent here* (Goldich and Bergquist 1948).

There are also blanket deposits in the Pacific bauxite belt. The northwestern part of Rennell Island is covered by a blanket 1 to 3 m thick over an area of several square kilometres (De Weisse 1970; Taylor and Hughes 1975). Lifou, Maré and Ouvea islands

* The irregular outline of these deposits is shown in figure 11. Note the small areal extent of the dolines below bauxite.

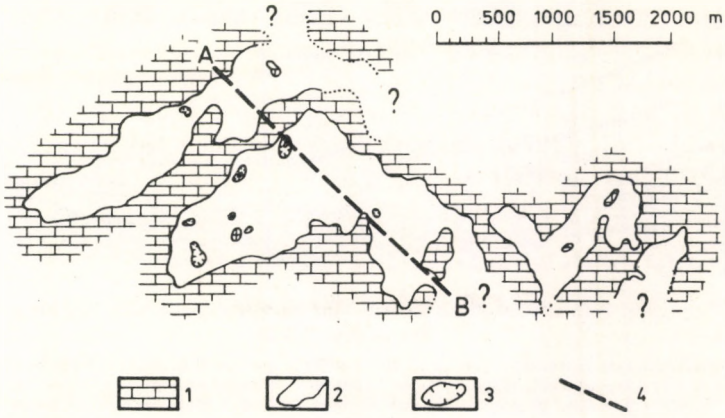


Figure 11. Map sketch of blanket-like deposits. Haiti, Rochelois Plateau. After Goldich and Bergquist (1948)
(1) limestone bedrock, Eocene; (2) bauxite; (3) sinkhole-like depressions of the bedrock; (4) trace of the profile shown as figure 10 of this book

in the Loyalty Group are covered almost everywhere (over an aggregate area of some 30 km^2) by a bauxite blanket 1/2m thick on average, exceeding 1 m only above the gentle karst depressions of the underlying Pleistocene limestone (Gsell 1971; oral communication). Similar blankets were described by Schofield (1959) from Niué Island, lying to the east of the Tonga Group, and by Carroll and Hathaway from Northeast Guam (1963). On the latter island, the bauxite blanket is even less than 1/2 m thick, and the total bauxite tonnage is so little that the present writer has preferred not to include it in his comprehensive list of bauxite deposits.

3.2.1.3. STRIP-LIKE DEPOSITS

Often several kilometres long for a width of less than 800 m or, sometimes, than 500 m, these deposits are not necessarily smaller than the stratiform ones. Their thickness is 5 to 10 m typically, and up to 20 m above the occasional depression of the bedrock relief. The deposit-top surface is more or less plane. Total bauxite tonnages also attain those of the stratiform deposits.

A comparatively rare type, the strip-like deposit is most abundantly represented in the Brignoles region of France. The Mazaugues deposit is uninterrupted over a length of some 10 km; the Le Val-Vins–St Cristophe deposit measures 12 km; the Combecave –Peygros deposit, 11 km; the Rigoulier–Pas de Recous deposit, 6 km. Figure 12 is an outcrop map, figure 13 is a north–south transverse profile of the Le Val-Vins deposit. In the latter, the great thrust fault that shoved the pre-bauxitic beds above the post-bauxitic ones, is clearly visible. The deposits are 5 to 10 m thick in most places, 20 m thick at most. The bedrock relief is but gently karsted; the deposit-top surface is gently depressed over the central parts of the deposits (Nicolas 1970). The bauxite deposit gradually taper away and then wedge out at the edges. The strip is not as a rule composed of bauxite in its entire length: industrial-grade ore usually forms lenses embedded in bauxitic clay. The aggregate bauxite tonnage in one strip is 20 to 50 million tons.

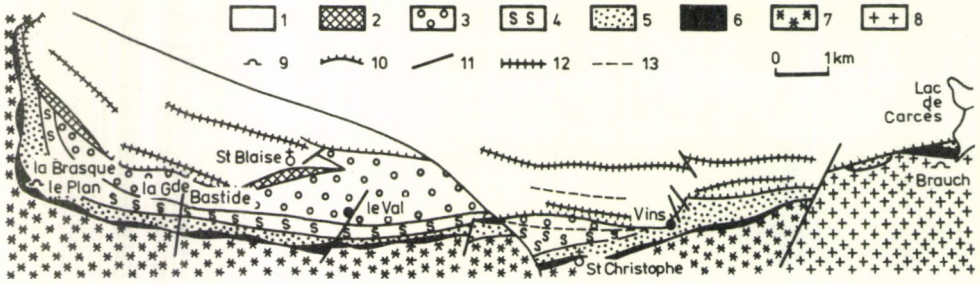


Figure 12. Geological map sketch of the Le Val-Vins bauxite deposit area according to Nicolas and Hieronymus (1969)

(1) overthrust bedrock succession, pre-bauxitic; (2) rags of overthrust bauxite; (3) gravel and sand, Bégudian (Maestrichtian); (4) sand, Fuvelian (Upper Campanian); (5) sandy-clay strata, Valdonian (Lower Campanian); (6) bauxite outcrops; (7) dolostone, Malm; (8) limestone and dolostone, Bathonian; (9) access to mine; (10) overthrust; (11) fault; (12) anticline axis; (13) syncline axis

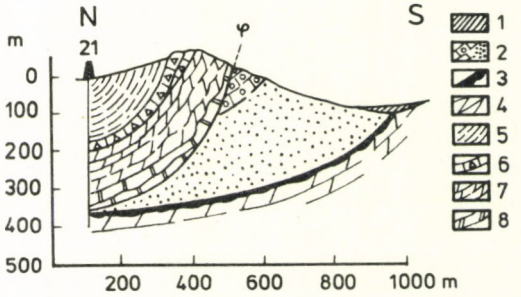


Figure 13. Geological profile through Vins village. After Lemperière (1973)

(1) alluvium; (2) limestone and sandstone, Upper Cretaceous; (3) bauxite deposits; (4) dolostone, Malm; (5) limestone and marl, Bajocian; (6) limestone, Upper Liassic; (7) dolomite and limestone, Hettangian; (8) limestone, Rhaetic

The deposits of the Nikšička Župa in Crna Gora (Montenegro) State, Yugoslavia are somewhat smaller (1 to 2 km long; 100 to 300 m wide). The bauxite is 10 to 20 m thick on average, but may exceed 50 m locally. Bauxite tonnage per strip ranges from 10 to 20 million tons.

3.2.1.4. LENTICULAR DEPOSITS

Typical extent of these deposits is in the 1 to 40 hectare (0·01 to 0·4 km^2) range. Outlines are most often irregular, more seldom round or oval. Diameters range from 200 to 600 m. Bauxite is thickest in the central part of the lens, gradually tapering away towards the edges. Typical bauxite thickness is 5 to 20 m; maximum thickness is about 30 m. Total bauxite tonnage per lens is 0·1 to 3 million tons. This is the most widespread form among Mediterranean-type deposits.

On the Manchester Plateau of Jamaica, several hundred bauxite lenses occur close together, often separated only by limestone rises not more than 20 to 50 m wide Outlines are highly irregular. To the south, these lenses gradually merge into blanket deposits. Lenticular deposits are frequent also in the other Jamaican bauxite regions, except in the Cockpit Country. They have the remarkable feature that, in the majority of cases, their surfaces are not plane: they copy the bedrock relief instead, much as in a

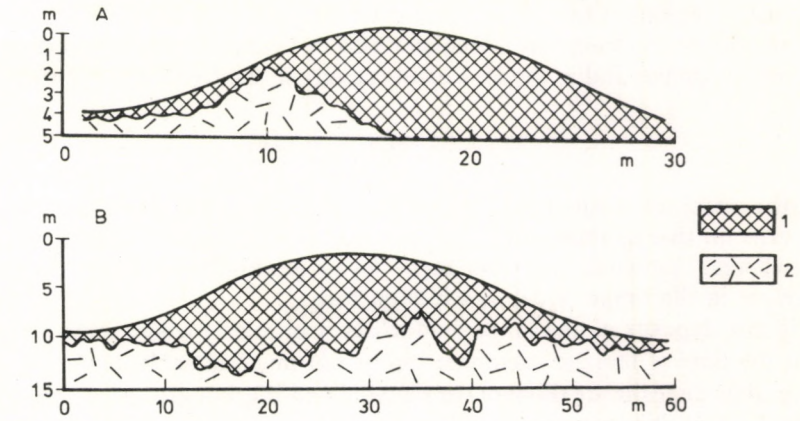


Figure 14. Profiles of lenticular bauxite deposits on Manchester Plateau, Jamaica
(A) road cut; (B) bauxite hill prospected by drilling; (1) underlying limestone; (2) bauxite

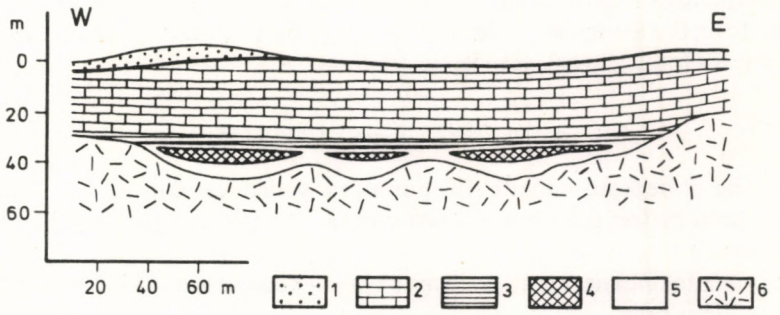


Figure 15. Lenticular bauxite deposit. Szőc, Hungary
(1) sandy loess, Pleistocene; (2) limestone, Eocene; (3) multicoloured and carbonaceous clay, Eocene; (4) bauxite; (5) clayey bauxite and bauxitic clay; (6) dolostone, Upper Triassic

blanket deposit. On the Manchester Plateau, however, some of the lens surfaces are convex upward, as revealed by open pits, road cuts and numerous boreholes (fig. 14). These comparatively flat, smooth-topped and rounded bauxite hummocks differ sharply from the nearby limestone hillocks of similar size, but having steeper and more dissected flanks.

The washing together of local dissolution residues could hardly have given rise to such a surface morphology. No ravines have formed in the area since the deposition of bauxite; the forms cannot therefore be of erosional origin. The most probable hypothesis of their origin invokes formation from wind-blown volcanic tuffs, to be discussed in detail in § 8.2.8.

Some lenticular deposits are comparatively thin (1 to 4 m) for an area of 20 to 30 hectares. Examples of this type occur in Slovenia, Yugoslavia, and also in the Yazd and Kerman regions of the Central Iranian Plateau.

In Hungary, lenticular deposits are most frequent in the Nyirád–Szőc region (fig. 15), on the southern rim of the Halimba Basin, and also in the Kislőd, Alsópere and

Nagyegyháza regions. Most Yugoslavian, Greek and Turkish bauxite deposits are lenticular. This form of deposit is frequent also in France (in the Languedoc bauxite region, in the Alpilles and in the Le Revest region), in the Urals and in Soviet Central Asia.

3.2.1.5. GRABEN DEPOSITS

In areal extent, they equal the lenticular types more or less, but they are cut off by fault scarps on two or more sides. Bauxite tends to be thickest next to these border faults, whereas lenticular deposits taper away gradually outward. Total bauxite tonnage are in the range of 0·2 to 4 million tons.

Fairly rare deposits of this shape tend to occur primarily where some faulting took place at the time of bauxite accumulation or not much later. In Hungary, the most typical graben deposits are those of the Fenyőfő and Iharkút groups of deposits. Given their areal extent, they contain very thick bauxite, up to 100 m (fig. 16). This type also includes the Angerrét and Meleges I and II deposits of Gánt, Hungary.

In the Nyirád region, deposits structurally delimited on one or two sides but gradually tapering in all other directions are fairly common (fig. 17). These represent transitions from the lenticular to the graben type. Transitional deposits of this sort also occur in Crna Gora, Yugoslavia (Burić 1966).

3.2.1.6. CANYON-LIKE DEPOSITS

In this type of deposit, bauxite fills out steep-walled, canyon-shaped depressions, several hundred metres long but not wider than 10 to 50 m, either nearly straight or quite tortuous.

Typical bauxite thickness is 10 to 20 m; a typical bauxite tonnage should be between 0·1 to 0·8 million tons.

Deposits of this form are also rare. In the Dalmatian bauxite region, the writer had the occasion to visit a largely worked-out canyon-like deposit near Blizanci village (fig. 18), where the tortuous trace of the canyon and the near-vertical walls of the bedrock limestone were clearly apparent. The deposit was not dissected by any faults. A similar deposit was described by Marić (1970) at Kruševo near Obrovac. The open pit mining of that deposit had laid bare a highly tortuous canyon of several hundred metres length. In Italy, the writer has seen some canyon-like deposits near Spinazzola. These are 200 to 400 m long by 20 to 30 m wide, dotted with funnel-shaped expansions and depressions in which the bauxite is up to 40 m thick (its usual thickness is 10 to 20 m). These canyons are less tortuous than the Yugoslavian ones. Their sub-parallel east–west arrangement suggests structural (fault) control.

3.2.1.7. SINKHOLE DEPOSITS

These deposits tend to be of less extent (0·1 to 1·0 hectare) than the previous ones, and to have round or oval, more seldom irregular outlines. Their diameters are 50 to 200 m typically. The sinkholes are very steep-walled, sometimes vertical, with funnel-shaped bottom parts. In places, several steep-walled funnels packed close together may constitute a larger composite doline. At the centre of a sinkhole feature, the bauxite

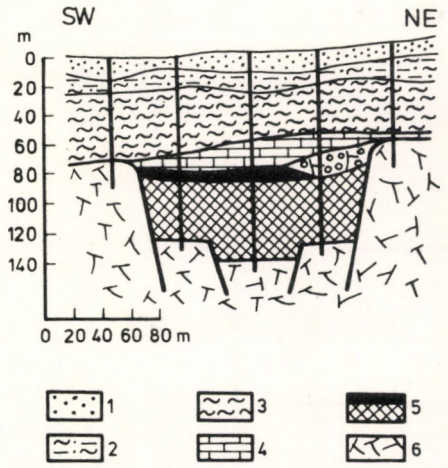


Figure 16. Graben type bauxite deposit. Fenyőfő, Hungary. After Szantner and Szabó (1970)

(1) sand, Pleistocene; (2) marl and clayey sand, Upper Pliocene; (3) marl, Miocene; (4) limestone, locally underlain by limestone debris, Middle Eocene; (5) sand with bauxite debris, underlain by the bauxite deposit; (6) dolostone, Upper Triassic

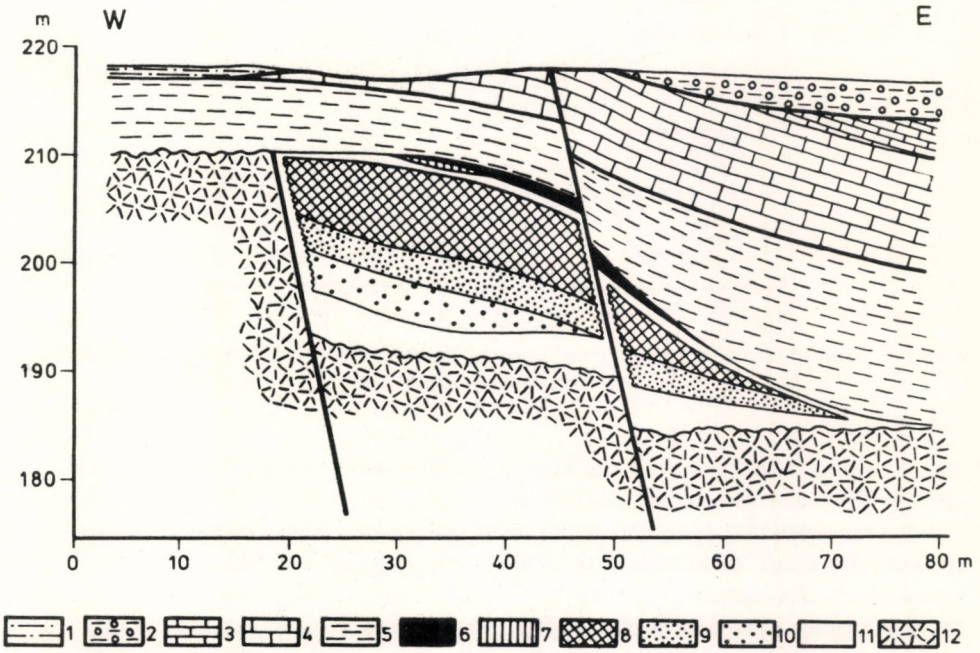


Figure 17. Bauxite deposit leaning on a fault. Nyírád–Izamajor, Hungary

(1) sand, clay, Upper Pliocene; (2) clayey gravel, Upper Pliocene; (3) limestone, Tortonian; (4) nummulites limestone, Middle Eocene; (5) miliolinid limestone, Middle Eocene; (6) carbonaceous clay, Lower Eocene; (7) multicoloured clay, Lower Eocene; (8) fair grade bauxite; (9) bauxite; (10) clayey bauxite; (11) bauxitic clay; (12) dolostone, Upper Triassic

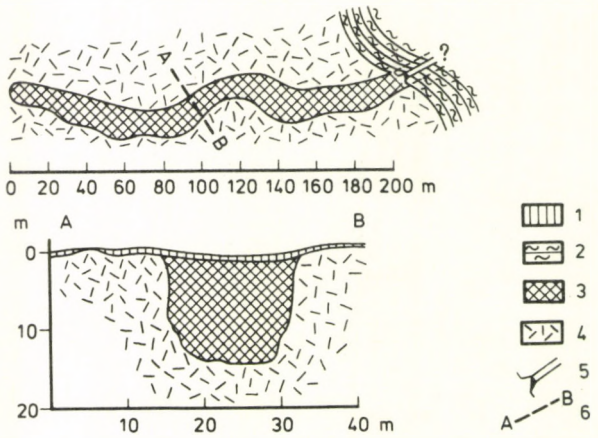


Figure 18. Canyon-like bauxite deposit, Blizanci near Mostar, Yugoslavia
(1) soil; (2) marl, marly sandstone, Upper Eocene; (3) bauxite; (4) alveolina limestone, Lower and Middle Eocene; (5) incline; (6) trace of geological profile

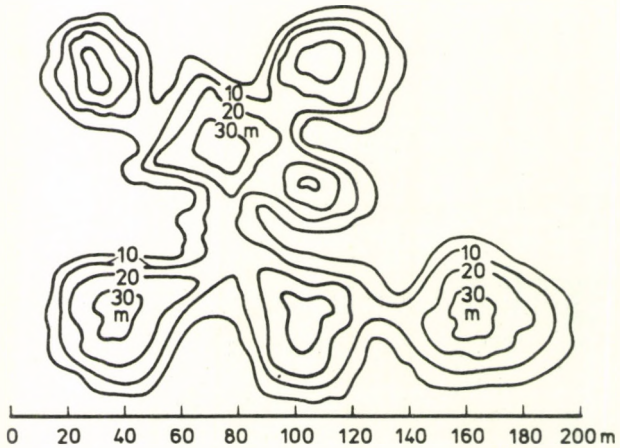


Figure 19. Ground plan of a composite sinkhole-type deposit cluster near Tavernes, Haut-Var, France. Contours indicate doline depth

may be up to 50 m thick, but the lateral termination of the deposit tends to be very abrupt. The bauxite tonnage per doline is in the range of 0·05 to 1 million tons.

After the lenticular type, this is the most widespread form of deposit. Sinkholes of considerable depth, filled with 20 to 50 m of refractory high-alumina (bauxitic) clay, are found near Cserszegtomaj in Hungary. The more frequent sinkholes of the Nyirád and Iharkút regions have depths of only 10 to 40 m as a rule, but exceptionally up to 80 m. In France, this is the typical deposit form of the Haut-Var region. The isolated, funnel-shaped sinkholes there are most often of 20 to 60 m diameter and 20 to 40 m depth. Composite dolines of complicated outline, probably due to the coalescence of adjacent single funnels, are more rare (fig. 19).

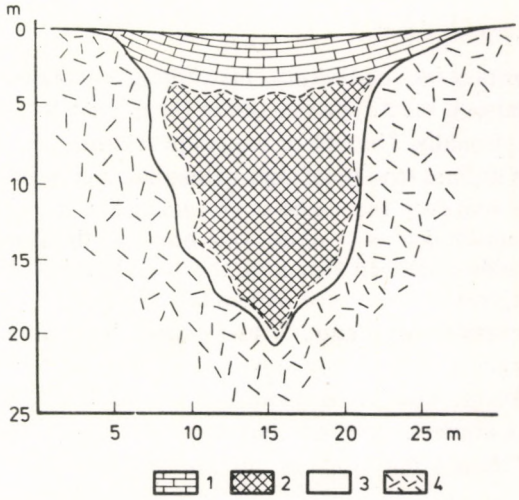


Figure 20. Bauxite deposit in a conical doline near Višnjan, Istria, Yugoslavia
(1) limestone, Lower Eocene; (2) bauxite; (3) clayey bauxite and bauxitic clay; (4) rudist limestone, Senonian

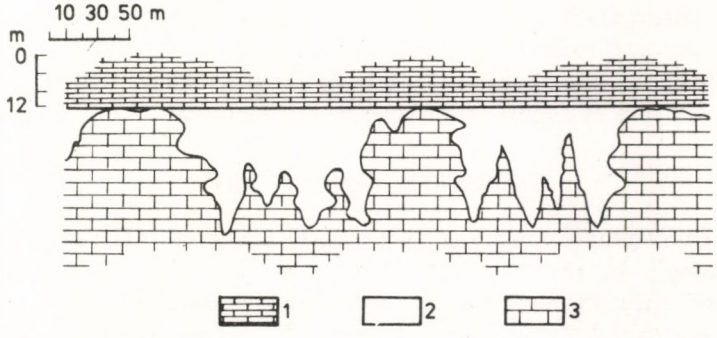


Figure 21. Composite sinkhole-type bauxite deposit cluster, Kiona Range, Greece. After Nia (1968)
(1) limestone, Turonian–Senonian; (2) bauxite; (3) limestone, Cenomanian

Sinkhole deposits are especially frequent on the Istrian Peninsula. Most of the deposits there have the conspicuous feature that the overlying beds are somewhat subsided over the central part of the filling (fig. 20). Such forms are frequent also in the Mostar and Imotski regions (Yugoslavia), where dolines of the usual size (of 20 to 100 m diameter) are accompanied by some quite small ones. The bauxite contained in these latter ones is of high enough grade to be economically workable. In the mined-out holes, the bedrock relief is clearly visible. These funnel-shaped small forms are of 2 to 10 m diameter and of 5 to 10 m depth. In the Kiona Mountains of Greece, the composite doline is the most frequent form (fig. 21). A few such forms, far less abundant than some others, occur also in Jamaica, especially in the intensely karsted Cockpit Country.

3.2.1.8. BAUXITE 'NESTS' AND 'BAGS'

These are the deposits of the smallest size, mere fillings of small depressions of 5 to 40 m diameter in the bedrock relief. The bauxite in them is 0·5 to 2 m thick as a rule; it attains 5 m only exceptionally. The nest or bag shape is controlled by the relief forms of the bedrock surface. Outlines tend to be highly irregular, zigzagging. The bauxite tends to be missing from above even the slightest prominence of the bedrock. These forms contain from a few hundred tons to about one hundred thousand tons of bauxite.

The type is fairly widespread. In Hungary, the Nagyharsány deposits are a case in point: their capricious forms have been adequately disclosed by mining (fig. 22). The nests are seen to be connected by tenuous branches and channels of bauxite. The form is frequent in Yugoslavia, Greece, Italy, Spain, and also in the Palaeozoic bauxite horizons of the Turkestan and Aliysk mountains of the USSR (Abdullayev 1970).

The eight sections above encompass but the most typical forms of deposits. In addition to these, there is a multitude of intermediate forms.

3.2.2. ARRANGEMENT OF DEPOSITS

A lone bauxite deposit in the middle of a large area containing no other bauxite is a rarity. Deposits tend to occur in groups. The term 'group of deposits' is used with a specific meaning in this book (cf Chapter 1). By the manner that the deposits are arranged, two types of deposit group can be distinguished.

3.2.2.1. RANGES OF DEPOSITS

The deposits of a range constitute more or less linear strings. The length of the group may accordingly be several times as great as its width. This arrangement is comparatively rare. Its most typical representatives occur in Soviet Central Asia (Turkestan and Aliysk ranges; also, Bukantau, Tamdytau and Nuratau mountains). These two regions have an aggregate length exceeding 1000 km, for a width of 20 to 50 km. They contain exclusively deposits of lenticular, nest and bag shape (fig. 23, separate, at the end of the book).

The Devonian region in the Northern Urals constitutes a north–south range of more than 300 km length and 10 to 30 km width, made up mostly of stratiform deposits and a few lenticular ones (fig. 24, separate, at the end of the book). The Brignoles region of France includes two more or less parallel ranges of interconnected strip-like deposits. The length of the southern range is 34 km; that of the northern range is 35 km; the width of each is 1 to 3 km (fig. 25).

3.2.2.2. FIELDS OF DEPOSITS

There is no significant difference between the length and width of such a group. The outlines of bauxite fields, as defined by the outermost deposits, tend to be rather intricate. Most of the Mediterranean-type deposits occur in groups of this type.

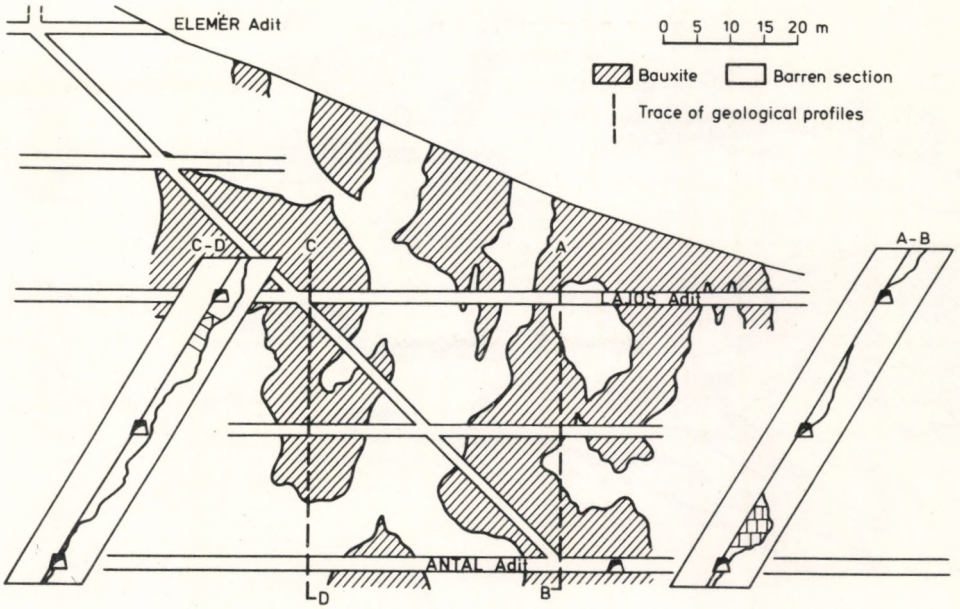


Figure 22. Bauxite nests and bags, Nagyharsány, Hungary. After a manuscript sketch by Vadász (1944). Map and profiles of the mine

Most deposit groups include deposits of different form types. Large stratiform deposits tend to be surrounded by lenticular ones. These latter tend to grow smaller towards the edges of the group, as exemplified by the Gánt region (fig. 9).

In Jamaica, Anderson (1971) stated deposits to be of the sinkhole type in the interior of the island; towards the seashore, lenticular deposits appear. These are surrounded by limestone plateaux of comparatively gently sculptured relief, bearing blanket deposits. Finally, near the coast, blanket and lenticular deposits dominate, largely made up of bauxitic clay. It was Hill and Davis (1971) who recognised that this arrangement is a result of structural control. The writer's own observation was that the limestone plateaux uplifted to the greatest height in the interior of the island are most intensely karsted, and are accordingly dominated by sinkhole deposits (e. g. Cockpit Country). All the limestone plateaux of appreciable extent surrounding these heights are tilted, gradually sloping in this or that direction. High up on these tilted plateaux there are mostly lenticular deposits; these pass into blanket deposits down the slope. In the larger grabens such as Mile Gully and on the low limestone plains, blanket deposits occur almost exclusively. The chemical and mineralogical composition of the bauxites duly reflects this arrangement, too.

In Central Greece, Nia (1968) recognised the following arrangement in the Turonian group of bauxite deposits: the deposits are mainly lenticular with a gentle bedrock relief in the southeastern part, whereas the more strongly karsted bedrock of the northwestern part carries deposits of the composite-doline and intermediate lens-to-doline types. The transition between the two subgroups is a gradual one.

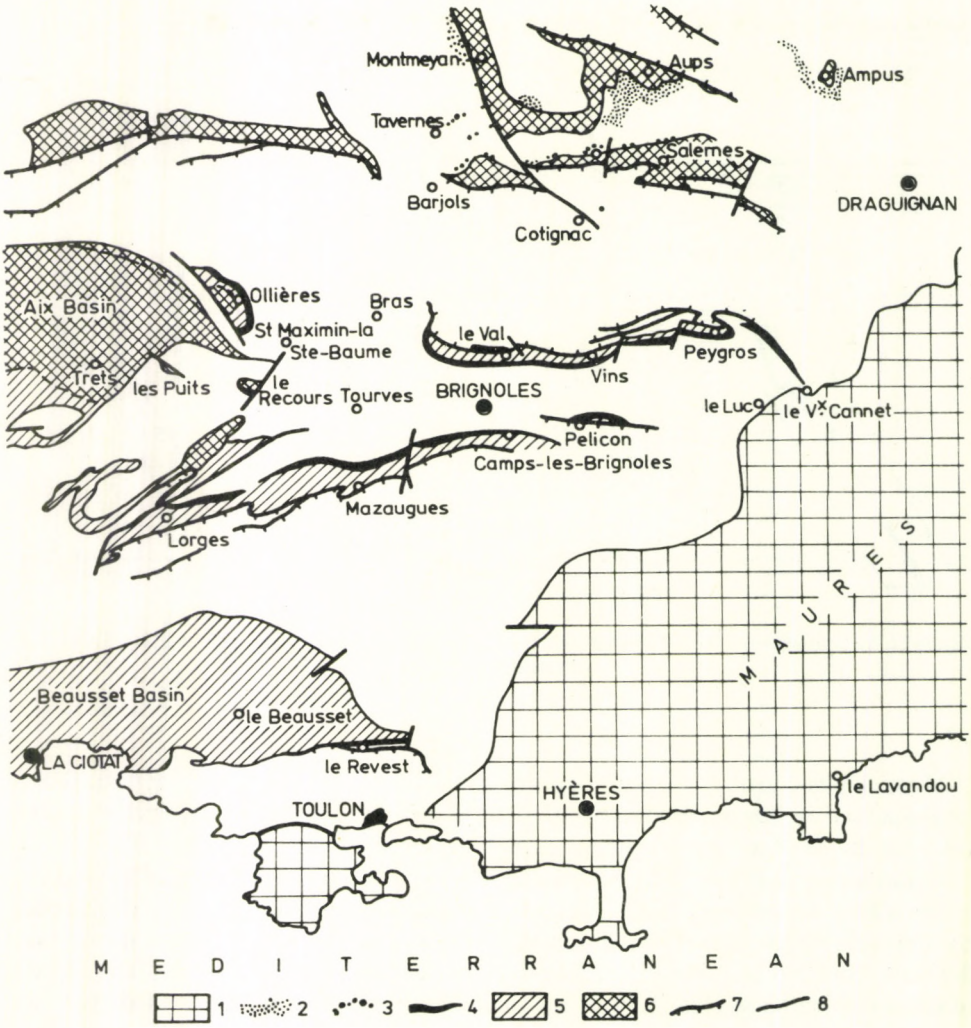


Figure 25. Bauxite geological sketch map of the bauxite deposit groups of Brignoles and Le Revest, France. After Nicolas (1970)

(1) crystalline basement; (2) bauxitic clay and bauxite-like clay in sinkholes; (3) bauxite in sinkholes; (4) banded bauxite deposits; (5) synclines containing Santonian and Campanian strata; (6) synclines containing Paleocene and Eocene strata; (7) overthrusts; (8) principal faults

The areal density and the size distribution of deposits within a group should be an important depositological criterion. For a Hungarian example, let us cite Barnabás (1968): over the 21·7 km^2 of the Nyirád region prospected up to that time, 102 bauxite lenses and sinkholes had been identified. Of these, forty were in the 0·1 to 1·0 hectare size range; 53 in the 1·1 to 5·0 hectare range; and 9 in the 5·1 to 10·0 hectare range. Similar numerical data were published by Marušić (1960) for a region named Belićko Polje on the Istrian Peninsula. There, 106 sinkholes, very small ones of necessity, had been identified within a single square kilometre (fig. 26).

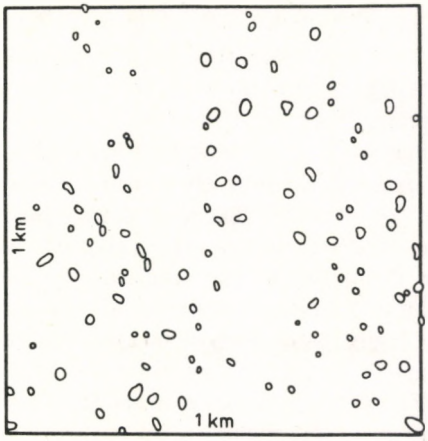


Figure 26. Distribution of sinkhole-type bauxite deposits in the Belićko Polje area, Istrian Peninsula, Yugoslavia. After Marušić (1960)

For a quantitative characterisation of deposit density, it would be necessary to calculate relative areas of bauxite-covered and barren areas per square kilometre for each group of deposits. The variation of this ratio as a function of the azimuth would be a useful tool of prospecting management. Unfortunately, no data of this kind are available at the present time.

3.2.3. RELATIONSHIP BETWEEN THE BAUXITE AND ITS BEDROCK

3.2.3.1. BEDROCK LITHOLOGY

Most deposits of the Mediterranean-type overlie carbonate-rock series several hundred metres thick. Appendix 1 indicates among others bedrock lithology for each group of bauxite deposits. Of the 145 deposit groups of Mediterranean type, 106 have a limestone bedrock: 4 more overlie marbles (that is, metamorphosed limestones). In 6 groups, the bedrock is predominantly limestone accompanied by some dolomitic limestone and dolostone. In 9 groups, the bedrock is largely dolostone, with some dolomitic limestone and limestone. In Hungary, this group includes the Halimba deposit, where the bedrock is largely Upper Triassic 'Hauptdolomite', and to a less extent 'Dachsteinkalk' limestone. The Fenyőfő–Bakonyoszlop region is of a similar nature. Eighteen groups have an exclusively dolostone bedrock: of these Nyirád, Szőc, Kislőd, Iszkaszentgyörgy, Gánt and Nagyegyháza in Hungary, Bédarieux in France, Northwest and Northern Iran, the Central Iranian Plateau and the Pamirs are the most important. The only place where the rock immediately underlying the bauxite deposit includes marly dolostone in addition to pure dolostone is Yazd in Iran. In one region (the Salt Range in Pakistan), limestone is accompanied by some calcareous sandstone.

In the great majority of cases, then, the rock immediately underlying the bauxite deposit is a high-purity carbonate rock that rarely contains more than one percent non-carbonate matter. Its organic content is also low, so that the bedrock is typically a very light grey or pale yellow rock.

Remarkably enough, marls will carry no bauxite even in the immediate proximity of a bauxite region. For example, in the Fenyőfő group of Hungary, the bauxite overlies Norian 'Hauptdolomite' or Rhaetic 'Dachsteinkalk'. Bauxite disappears to the northwest, where these bedrock types are replaced by Carnian marl. The Iszkaszentgyörgy group is limited to the southeast by an interface of Ladinian dolostone against Anisian and Lower Triassic marls and sandy marls.

In Jamaica, the bauxite region is cut off by a line marking the contact of pure bedrock limestone against the marly 'Montpelier' limestone with chert nodules and bentonite intercalations. This cutoff is especially striking on the St Ann Plateau (Robinson 1971).

3.2.3.2. BEDROCK FACIES

The rock formation immediately underlying the bauxite deposit is neritic in most groups of bauxite deposits. Within this category, biogenic reef facies are especially frequent. Shoal water does not invariably imply the proximity of a shoreline: often enough, a shallow bank remote from any shore may be inferred. In the Transdanubian Hills of Hungary, for instance, the immediate bedrock is of the 'carbonate-plateau' facies. This means water slightly deeper than that which is indicated by a true reef. Within this range, 'Hauptdolomite' indicates shoal water and 'Dachsteinkalk' a slightly greater depth. These facies do not suggest the proximity of a shoreline, either, but rather an extensive archipelago shelf (Végh-Neubrandt, oral communication, 1974).

Lagoon facies and facies of reduced salinity immediately underlying a bauxite deposit have been identified in just a few cases, including the Helicon Mountains (Papastamatiou 1965) and the Maktesh Ramon region of Israel (Würzburger 1960). Not a single instance is known of bauxite being directly underlain by fresh-water sediments.

3.2.3.3. CONTINUOUS VERSUS INTERRUPTED DEPOSITION

One single instance is known where bauxite emerges in full conformity from a limestone of lagoon facies: the Kephanedes deposit of the Helicon Mountains in Greece (Papastamatiou 1965). First of all, some scattered bauxite ooids and pisoids appear in the limestone. These grow more abundant upward, in step with a reduction in the share of the carbonate matrix. Higher up, this matrix disappears entirely, giving way to an ooidic–pisoidic bauxite. The zone of transition is 5 to 15 cm thick (fig. 27).

The passage may be continuous also in the small bauxite deposit near Sisophon in Campuchea, as implied by the scattered appearance and upward-growing abundance of minute red bauxite fragments in the Upper Permian limestone underlying the bauxite. The bedrock-to-limestone interface, on the other hand, is a sharp one, contradicting the assumption of conformity. According to Saurin (1965), this bauxite had accumulated in the sea, without a prior immersion. In the writer's opinion, information about this deposit is not as yet detailed enough to permit a satisfactory analysis of this problem.

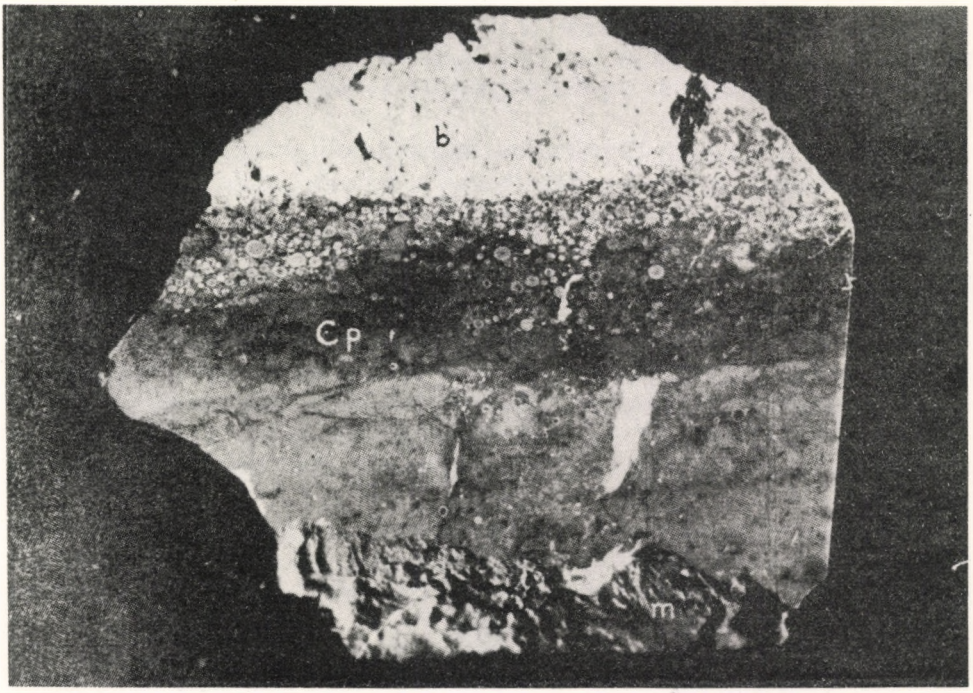


Figure 27. A lagoon-facies limestone grading into bauxite; continuity of deposition. Kephanedes, Helicon Range, Greece. After Papastamatiou (1965)
(m) bedrock limestone; (Cp) calcareous-bauxitic transitional section; (b) bauxite

All other deposits overlie their respective bedrocks *unconformably*, with a sharp boundary representing a marked hiatus. If this hiatus is a short one, there may be just an erosional disconformity, whereas longer hiatuses are often accompanied by an angular unconformity. The latter is the case in the bauxite region of the Transdanubian Hills of Hungary.

3.2.3.4. KARSTIFICATION OF THE BEDROCK RELIEF

The surface of the bedrock under the bauxite deposit is more or less karsted in most cases. The degree of karstification is highly varied, as regards both the depth and the intricacy of the karst relief. The slopes are the steeper and the relief is the more intricate, the closer is the spacing of prominent and hollow karst forms. Unfortunately, the information available does not permit the quantifying of these relations. The writer has accordingly introduced a five-degree scale of karstification (very gentle, weak, moderate, strong, very strong). This degree of karstification is indicated for each of the groups of deposits listed in Appendix 1. Of the 145 Mediterranean-type deposit groups, no relevant information on karstification was available for eight. For the rest, the following picture has emerged.

In none of the deposit groups is the bedrock surface entirely plane with no karstification at all. It is not entirely plane even in the above-mentioned Campuchean group; the description suggests that some slight karstification must have taken place.

The bedrock is *very gently karsted* in four Iranian deposit groups (Bukan; upper and lower bauxite horizons of Jajarm; Behbehan). The bedrock undulates very gently; the rises are very flat and seldom higher than one metre.

Two deposit groups are known in which parts of the bedrock surface are very gently karsted, the rest being weakly karsted. These are Podlipa–Borovnica in Slovenia, Yugoslavia, and Boksonsk in the Eastern Sayan Mountains, USSR. In the latter case, the rock immediately underlying the bauxite deposits has been shown to be penetrated by karst cavities 1 to 1·5 m wide. Tapering away downwards, these disappear at a certain depth (more than 50 m below the deposit in some of the boreholes: Ilina 1958).

The bedrock is *weakly karsted* in 19 deposit groups. The amplitude of the relief attains 3 m; slopes are steeper. This group includes the Nagyharsány group of Hungary.

In 17 groups, bedrock–surface karstification is weak in part of the area and moderate in the rest. This category includes the Triassic deposit group of Lika, the Senonian one of the Grmeć Mountains, the Cenomanian one of Bijele Poljane (all in Yugoslavia), and the Lower Malm one of Central Greece.

The bedrock exhibits *moderate karstification* in 54 deposit groups. Here, the amplitude of the karst relief may attain 10 m, with locally some small steep-walled funnels and pinnacles. In the immediate bedrock, bauxite-filled karst passages and cavities are frequent. Most of the Hungarian bauxite occurrences belong to this category.

In 26 of the deposit groups, the bedrock relief is moderately karsted in parts of the area and strongly karsted in the rest. These groups include mostly lenticular and more rarely stratiform deposits, with alternating larger, comparatively gentle depressions and deep, steep-walled sinkholes. The karst relief tends to be most rugged next to the outcrops of bauxite. In Hungary, the Gánt and Iszkaszentgyörgy occurrences belong to this category.

Strong karstification has been observed in 12 deposit groups. The amplitude of the karst relief exceeds 10 m and may attain even 50 m. Very steep-walled karst forms are frequent, in places including overhanging walls and toadstool shapes. The relief tends to be roughest under groups of sinkhole deposits, where each deposit is essentially the filling of a karst sinkhole. The most typical examples include Cserszegtomaj in Hungary, Istria, Obrovac, Mostar (Yugoslavia), Spinazzola (Italy) and the Haut-Var (France).

One deposit group exhibits *very strong karstification:* this is the Jamaican Cockpit Country, where the relief as a whole is a complex surface of close-spaced karst towers and sinkholes. Bauxite occurs as sinkhole fillings.

Finally, a single region is known where weakly, moderately and strongly karsted reliefs combine in the bedrock surface of the bauxite. This is the Brignoles group of extensive strip-like deposits in France. Also here, the karst relief is most rugged in the proximity of the outcrops. This fact is most conspicuous in the worked-out open-cast pits (fig. 28).

To sum up, in a large majority of all Mediterranean-type bauxite deposit groups, the bedrock surface is moderately to strongly karsted. Weakly karsted bedrock reliefs are



Figure 28. A karsted bedrock surface under a worked-out bauxite deposit. Combecave–Pins-Bâtards deposit near Brignoles, France. Photo by A. F. De Lapparent (1956)

subordinate; very gently karsted ones are insignificant. Characteristically, the deposits of this last type are small, with an aggregate world tonnage of 43 million tons. The most substantial reserves of Mediterranean karst bauxites are connected with moderately and strongly karsted bedrock reliefs.

The degree of karstification and the nature of the resulting relief is the resultant of a number of factors. In several deposit groups, the bedrock tends to be more intensely karsted under the higher-lying than under the lower-lying deposits. This is the case e.g. in Jamaica, as pointed out by Anderson (1971).

A number of authors attribute a controlling role to bedrock composition. De Weisse (1964), for example, points out that limestone tends to be more intensely karsted than dolostone, for the reason that dolomite is less soluble than calcite. As a matter of fact, limestone is karsted more readily as a rule than dolostone, but there are some other factors that may exert an even stronger influence on karstification. As a result, a dolostone bedrock in one area may be more intensely karsted than a limestone bedrock in the next area. For a striking example, one may cite the neighbouring Halimba and Nyirád deposit groups of Hungary: the 'Dachsteinkalk' at Halimba is less intensely karsted than the 'Hauptdolomite' at Nyirád. This in the writer's opinion is due to the circumstance that the dolostone at Nyirád has been made more amenable to karstification by intense fault fracturing.

Tectonic fracturing of the bedrock may thus become an essential factor in bauxite accumulation. There are many deposit groups whose deposits formed along fault lines where structural fracturing was strongest. The deepest dolines tend to lie at the crossing of faults. This is the case in the bauxite regions of Istria, Dalmatia and Central Greece. The canyon-like deposits of Spinazzola (Italy) also follow ancient faults. According to Combes (1973), structural control can be clearly recognised also in the Bédarieux group of deposits (France). The dolomite underlying the northern part of the region is intensely fractured: the sinkholes in that area are 20 to 40 m deep. In the southern part, the dolostone is much less fractured: the karst depressions there are seldom deeper than 10 m.

Clayey impurities in the bedrock and the consequent reduced permeability also seem to be an important factor. Meleshchenko (1945) observed in the Southern Urals that, wherever bauxite overlies a stratified argillaceous limestone, karstification is very slight, and the bedrock relief is just gently undulating (fig. 29, A); on the other hand, the bedrock is intensely karsted with a very uneven relief wherever the bauxite is underlain by a high-purity massive limestone (fig. 29, B).

Early on in bauxite geology, it was a widely held view that the karst-surface relief had been sculptured prior to the deposition of bauxite. Vadász (1951), however, proposed the idea that karstification could continue also during the process of bauxite accumulation. A similar conclusion was drawn by Bonte (1958), who stated the bulk of karst sculpture to have taken place in the course of bauxite deposition, in step, so to speak, with that process. Finally, in the last few years, conclusive evidence was found in some localities for the continuation of karstification even after the cessation of bauxite deposition. Under the structurally tilted deposits of Yugoslavia and Greece, for example, upright karst forms are often observed. These steep-walled towers and funnels are perpendicular to the actual horizontal plane rather than to a prior horizontal as defined by the bauxite deposits, which means that their formation postdates the tilting (fig. 30).

As a decisive proof for subsequent karstification, Sakać (1966) described the Radinovići deposit near Bosanska Krupa, embedded in an overturned fold so that the Campanian–Maestrichtian cover had come to lie beneath the deposit and the Turonian bedrock to lie above it. The interface between the deposit and the original bedrock is a gently sculptured, undulating karst relief, whereas the interface towards the original cover is an intensely sculptured one—obviously posterior to deposition and, indeed, to folding. Sakać inferred from this observation the necessity of distinguishing a paleo-relief and a pseudo-paleo-relief, the former being the original karst morphology of the bedrock prior to the accumulation of bauxite, and the latter being the relief formed after deposition, under the influence of downward-filtrating karst water.

According to Bonte (1970), sub-recent karstification takes place under deposits exposed at the ground surface or close to it; this is a process that may result in particularly rugged karst forms. In the open-cast pit Combecave–Pins-Bâtard near Brignoles he observed the karst forms to be most rugged near today's ground surface. The karst relief is the gentler, the more deeply it had been buried prior to its excavation (fig. 31). Bonte also attributed this phenomenon to the action of infiltrating ground water.

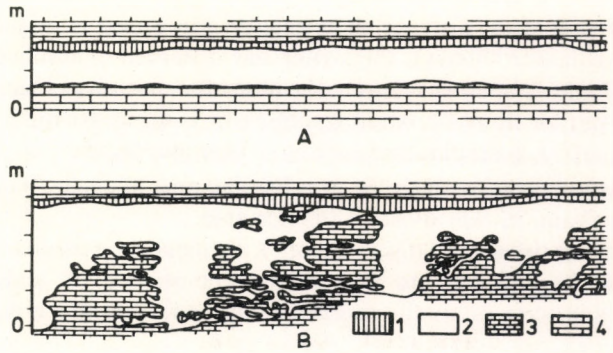


Figure 29. Relationship between bedrock composition and karstification. Novo-Pristansk deposit group, Southern Urals, USSR. After Meleshchenko (1945)

(1) green and greenish-grey bauxite; (2) red bauxite; (3) white massive limestone, Upper Devonian; (4) grey, well-bedded clayey limestone, Upper Devonian

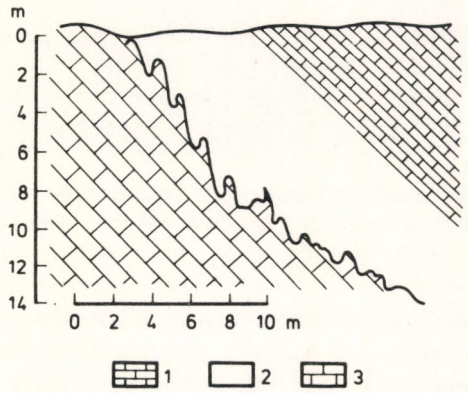


Figure 30. Upright karst forms beneath a steeply dipping bauxite deposit. Anthimos open pit, Distomon, Central Greece

(1) limestone, Tithonian; (2) bauxite; (3) limestone, Lower Malm

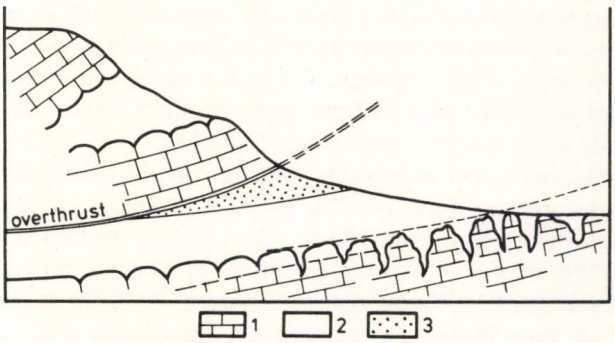


Figure 31. Relationship between dissection of karst relief and overburden thickness. Combecave–Pins-Bâtards, open pit near Brignoles, France. After Bonte (1970)

(1) bedrock limestone, Bathonian; (2) bauxite; (3) sandstone, Senonian

Sub-recent karstification younger than bauxite deposition is frequent in Jamaica, too. On the Manchester Plateau, the writer has observed a number of open karst funnels with sinkholes at their bottoms in the immediate vicinity of bauxite lenses. The open funnels do not, of course, contain any bauxite. Even under the bauxite deposits, however, open karst passages and cavities have been disclosed by mining. The writer could observe some examples on the St Ann Plateau near Walkers Wood, as well as in the St Elizabeth Plain, in the so-called Pepper area.

The writer is of the opinion that subsequent karstification is more or less significant especially where the deposits are exposed on the surface or underlie a shallow overburden. This process gradually ceases to function at depths greater than 50 m.

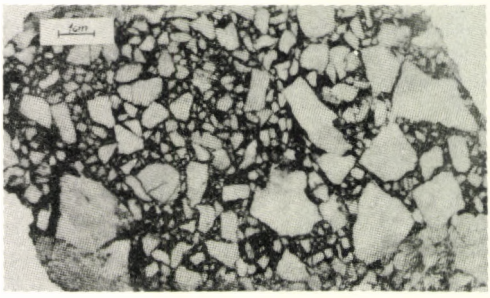


Figure 32. A limestone pseudo-breccia in a bauxitic matrix ('underore limestone breccia') at the bottom of a bauxite deposit. Middle Devonian, Northern Urals, USSR

A product of karstification simultaneous with or subsequent to bauxite accumulation is the so-called '*under-ore breccia*', recognised by Soviet geologists under Palaeozoic bauxite deposits in the Urals and in Central Asia. It consists of sharp angular limestone fragments embedded in a bauxite matrix (fig. 32). The size of the limestone fragments gradually increases downward and the share of the bauxitic matrix gradually decreases, giving a gradual passage into the underlying limestone, which in turn is traversed by bauxite-filled karst passages. According to Bushinsky (1971), this is a pseudo-breccia resulting from the unusually strong karstification of the limestone bedrock. His ideas were confirmed by Knyazeva (1966), who found that under the Ivdel deposits of the Northern Urals this 'under-ore breccia' is confined to where the bedrock is a pure reef limestone highly prone to karstification. It is missing wherever the bedrock is a siliceous or marly limestone.

Özlü (1977) observed a similar breccia, which he called residual, under the deposits at Mortas, Dogankuzu and Kiziltas of the Upper Cretaceous bauxite horizon in the Taurus Range. Limestone fragments of 1 to 50 cm size are cemented by bauxitic clay. The thickness of the breccia varies capriciously from 20 cm to 5 m. Özlü attributes a special role to karst cavity collapse in the formation of this breccia.

The bedrock in immediate contact with the bauxite deposit is often altered, pulverulent. In a limestone bedrock, this pulverulent zone is just a few centimetres thick, but in dolostone it may attain a thickness of 2 to 3 m. In Hungary pulverisation of the dolomite in direct contact with the bauxite deposit is most intense at Gánt and at

Iszkaszentgyörgy. In France, the writer has had the occasion to observe especially strong pulverulency in the Bédarieux bauxite occurrence.

Most marked under the central parts of the deposits pulverulency gradually disappears under their tapering edges. It is most obvious under deposits cropping out to the ground surface or lying close to it. For example, on the southern rim of the Halimba basin, under the lenses of the Malomvölgy group, dolostone underlying the bauxite deposit is intensely altered, loosely pulverulent to a depth of 10 to 100 cm. This layer's depth of burial used to be 1 to 20 m. In the basin, on the other hand, where mine workings are now 150 to 300 m deep, the surface of the 'Dachsteinkalk' is quite fresh, and even on the dolostone bedrock the writer could observe only a slightly pulverulent layer not thicker than 1 to 2 cm.

As far as the writer could see, limestone is most loose and pulverulent immediately under the bauxite deposits also in Jamaica. Even though full of holes and pores of dissolution, the outcrops surrounding the deposits consist of comparatively hard limestone.

All these observations imply that pulverisation is a process due to infiltrating ground water, presumably rendered 'aggressive', suitable for pulverising the carbonate rock, during its percolation through the bauxite. We shall return to this process in the chapter on Mineralogy.

3.2.3.5. ALIEN INTERCALATIONS BETWEEN THE BAUXITE DEPOSIT AND ITS BEDROCK

In most bauxite regions, the bedrock is directly overlain by bauxite or bauxitic clay. At a few places, however, intercalations of alien matter have been observed. In view of their depositological and genetic significance, these merit discussion in some detail.

In one of the deposits at Ollières, France, Portlandian limestone is overlain by a limestone breccia of Upper Barremian age. The calcareous cement of the breccia has yielded a number of Chara and calcareous algae including Munieria baconica. According to Nicolas and Esterle (1965), this breccia had formed in a lagoon prior to bauxite accumulation above the Portlandian Limestone surface. They see this as a decisive argument for the allochthony of this deposit.

In some of the bauxite deposits at Ivdel in the Northern Urals, the limestone bedrock is overlain in places by a well-bedded siltstone ('aleurolite') 0·5 to 1 m thick. Bauxite overlies the uneven surface of the siltstone with an erosional disconformity. According to microscopic observations by Knyazeva (1966), the siltstone is a highly altered tuffaceous deposit containing debris of effusive rocks, altered plagioclases and quartz grains typical of tuffs.

According to Bušer and Lukács (1970), there are in Slovenia thin lenses of tuff, tuffite and clay shale intercalated in places between the Triassic bauxite deposits and the underlying Upper Ladinian limestone or dolostone: a more detailed lithological and faciological study of these sediments, however, is lacking as yet.

In one of the open-cast pits at Campo Felice in Central Italy, the writer has found a green clay and clay marl in lenses of 1 to 4 cm thickness and a few square metres area, intercalated between the bauxite and its limestone bedrock. The writer's mineralogical and petrographic examinations revealed this material to be likewise a strongly altered

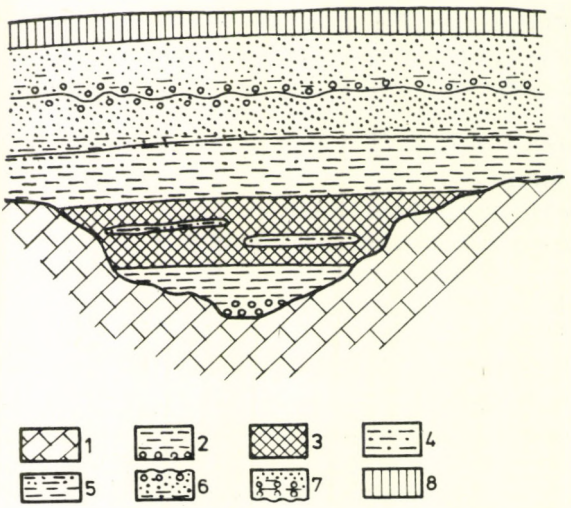


Figure 33. Constitution of a bauxite deposit in the Haţeg deposit group, Rumania. After Papiu *et al* (1971)
(1) limestone, Malm–Lower Cretaceous; (2) clayey–detritic basal layer; (3) bauxite; (4) clayey–detritic intercalations in the bauxite; (5) clayey–sandy cover beds; (6) sandstone and conglomerate; (7) transgressive beds, Cenomanian; (8) Quaternary

argillaceous tuffite. Its gradual passage into the bauxite implies that this tuffite was itself bauxitised in the course of bauxite accumulation (Bárdossy *et al* 1977).

Eyles (1958) found a phosphate-rich layer of 2 to 30 cm thickness, interrupted in places, inserted between the bauxite and its limestone bedrock in the Shooters Hill deposits near Mandeville in Jamaica. This layer abounds in fish teeth (the teeth, above all, of sharks, Chaetodonts and Sparids) smaller than 10 mm. Eyles regards this layer, which is much more radioactive than the overlying bauxite, as the vestige of a former bed of marine rock phosphate, although the evidence for this is not quite conclusive. Remarkably, the bauxite itself has an above-average phosphate content in these places. Anderson (1971), on the other hand, prefers the explanation that the phosphate-rich layer is a dissolution product of the limestone formed at a time when the limestone was already covered after the deposition of the bauxite. This assumption is confirmed by the circumstance that the limestone directly underlying the 'phosphate rock' contains the same kinds of fish teeth, although more scattered of course.

There are deposits where the detritic intercalation between the bauxite and the carbonate bedrock is even thicker and possibly uninterrupted. These deposits do not, properly speaking, belong to the pure Mediterranean type, but rather represent a transition towards the Kazakhstan type. Such deposits are marked with the symbol Med > Kaz in Appendix 1. In all their other depositological features, however, they agree with the Mediterranean type; the writer has accordingly placed them in that group in his statistical calculations.

The deposits of the Sohodol-Cîmpeni and Haţeg groups in Rumania are of this sort. Figure 33 shows the structure of a deposit at Haţeg according to Papiu *et al* (1971). In addition to a clayey-sand layer of some meters thickness right above the bedrock, there are thin clayey-sand intercalations also inside some of the bauxite lenses. Further cases

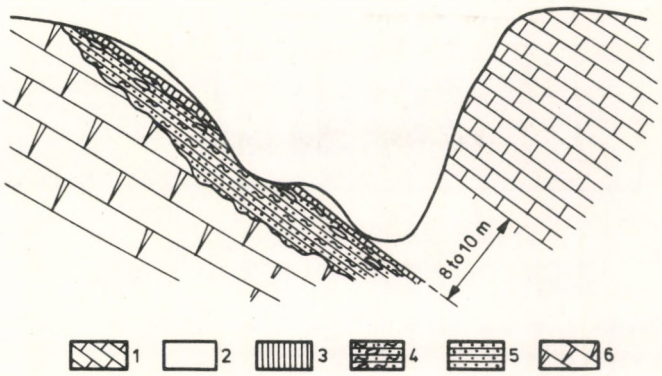


Figure 34. Profile of Mandra II bauxite deposit, Greece. After Papastamatiou and Maksimović (1970)
(1) clayey limestone, Cenomanian; (2) red bauxite; (3) red clay; (4) clayey sandstone with intercalations of red and green clays; (5) green sandstone; (6) limestone, Upper Triassic

are some of the deposits in the Eleusis–Mandra group in Greece (fig. 34). Lastly, some of the deposit groups in Northern and Southern Ferghana, USSR, also belong to this type (Krivtsov 1969).

The above-mentioned intermediate deposits taken together contain 25 to 30 million tons of bauxite, that is, less than one percent of the aggregate tonnage of all Mediterranean-type deposits. They do not, accordingly, weaken our depositological statements concerning the Mediterranean type of bauxite deposits. It would nevertheless be wrong to ignore them altogether, as they are an important evidence for the statement that there are transitions rather than sharp limits between the principal types of deposits.

3.2.4. RELATIONSHIP OF THE BAUXITE TO ITS COVER

Everywhere except in Jamaica, Hispaniola and the other Caribbean and Pacific islands, where bauxite is overlain by local regolith scree, and torrential clastics at best, bauxite formation was followed by the deposition of a sedimentary cover. (The loose texture and high porosity as well as some other mineralogical and petrographical properties of those island bauxites are due precisely to the absence of a cover.)

In Jamaica, the Pepper area is worthy of mention. Bauxite there is overlain by one to two metres of clay plus limestone and andesite pebbles spread by the flash floods of torrents rushing down the nearby Spur Tree Fault scarp. The torrents have in places cut elongate gullies into the bauxite surface, and even dug a few kettle-like pits (potholes), also filled with clay and gravel (Lee 1964).

According to Anderson (1971), boreholes near Porus traversed one to ten metres of limestone gravel above the bauxite, probably spread onto a valley-floor bauxite deposit by torrents emerging from the nearby limestone heights. In Harmons Valley, the bauxite is overlain by one to six metres of lacustrine marl, formed according to Anderson during repeated temporary inundations or water table rises. Apart from these local sedimentary covers, the Jamaican bauxites lie exposed everywhere.

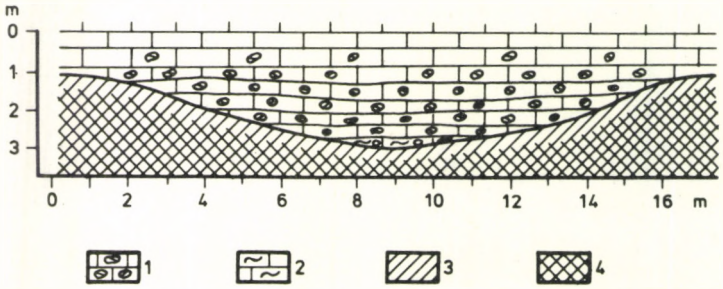


Figure 35. Contact of a bauxite deposit with its cover, Les Cannonettes open pit, Alpilles, France
(1) cover limestone with bauxite pebbles, Senonian; (2) calcareous marl, Senonian; (3) yellow and multicoloured bauxite; (4) brick-red bauxite

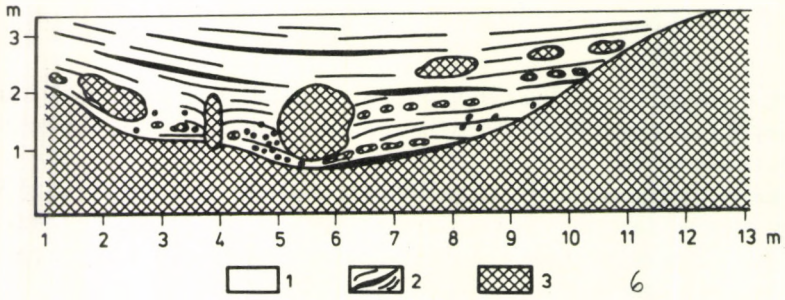


Figure 36. Contact between a bauxite deposit and its cover, Peygros Mine near Brignoles, France; Roadway profile
(1) grey carbonaceous clay, Senonian; (2) lenses of coal, Senonian; (3) dark brown and brownish red bauxite, Turonian



Figure 37. Eroded relief of a bauxite deposit top, covered with Miocene sandy gravel. Darvastó VI open pit, Nyirád, Hungary

The bauxites of all other regions were overlain by an uninterrupted sedimentary cover at one time or another. Wherever the deposits are exposed today, this means that their sedimentary cover has been removed by subsequent erosion. In most deposit groups, there is a longer or shorter hiatus between the bauxite and its sedimentary cover. There are on the other hand, a few deposit groups where conformity of deposition is observed.

Wherever there is a hiatus, the deposit-top surface is plane or gently undulating. In both cases, the interface towards the sedimentary cover is very sharply defined. The deposit-top surfaces of some strongly karsted lenticular and doline deposits exhibit saucer-shaped depressions; these are deepest above the central part of the deposit, where the interface may lie deeper by 2 to 3 m than at the edges of the deposit (fig. 20). The subsidence may be imputed partly to the compaction of the bauxite, and partly to ongoing karstification during and even after bauxite deposition.

The surfaces of some deposits exhibit ravine-like incisions, e.g. the deposit Les Canonettes in the Franch Alpilles. In the open pit there, the overburden has been completely removed prior to the winning of the bauxite, laying bare a ravine 1·5 to 2 m deep and 15 to 20 m wide. It traversed the open pit over a length of some 100 metres and disappeared under the overburden beyond the pit outline. According to Lemperière (1973, oral communication), the limestone beds filling the ravine were seen to form an overlapping transgressive arrangement on both sides of the ravine, indicating the gradual swamping of the gully (fig. 35). The writer had the occasion to observe an even more typical ravine in 1973 in the underground Mine de Peygros near Brignoles. A roadway there exposed the section of a ravine 2 to 3 m deep and 13 to 16 m wide. Each layer of the sedimentary filling overlapped the one below it on both sides of the ravine. Large bauxite pebbles and boulders were embedded near the ravine bottom (fig. 36).

In another type of situation, the original sedimentary cover of the bauxite may have been eroded and a younger sedimentary cover deposited subsequently. The result is as a rule an uneven surface of erosion, e.g. in some of the deposits at Darvastó in the Nyirád group of deposits (Hungary), where the bauxite came to be covered by clayey–sandy–gravelly Miocene beds after the erosion of the original Eocene cover (fig. 37).

Wherever the sedimentary cover develops conformably on the bauxite, there is no sharp interface between the two. There is instead a zone of peculiar transitional lithology, representing the gradual passage from bauxite to cover. This zone may be from 0·1 to 15 m thick.

Regardless of whether a hiatus exists, the top of most bauxite deposits is parallel to the bedding of the sedimentary cover: even the gentlest of angular unconformities are rare.

3.2.4.1. LITHOLOGY OF THE IMMEDIATE COVER

The immediate cover is much more varied as to lithology than the immediate bedrock: it may exhibit rather substantial differences even within one and the same group of deposits. Appendix 1 states for each deposit group the most frequent lithology of the immediate cover. Starting from the data given there, the writer has produced a list of the aggregate frequency of occurrence of cover lithologies.

Multicoloured, brown and yellow clay and claystone	22	deposit groups
Grey carbonaceous clay and claystone	22	deposit groups
Alternating claystone and sandtone beds	5	deposit groups
Sandstone and siltstone ('aleurolite')	16	deposit groups
Sandy marl	1	deposit groups
Clayey gravel	1	deposit groups
Clayey conglomerate	1	deposit groups
Calcareous and dolomitic conglomerate, breccia	14	deposit groups
Marl, calcareous marl	16	deposit groups
Limestone, agrillaceous limestone	67	deposit groups
Dolostone	2	deposit groups
Bone (carbonaceous shale) and dolostone, alternating	1	deposit groups
Weathered dolostone detrius and dolostone flour	2	deposit groups

(As deposit groups often feature more than one type of cover lithology, it would make no sense to calculate sums or percentages.)

The most frequent immediate cover, then, is *limestone*, less pure as a rule than the limestones of the immediate bedrock. In numerous deposit groups, it is high in organics and/or clay minerals. It frequently contains bauxite fragments and pebbles of some centimetres size, derived from the erosion of nearby bauxite deposits. It may be tinted pink or a pale red by totally elutriated fine bauxitic material. In the deposits of Campuchea, the bauxite is overlain by a siliceous limestone.

Dolostone plays much less of a role in the immediate cover than in the bedrock. Multicoloured, yellow, brown, grey and carbonaceous clays and claystone are, on the other hand, rather frequent. Carbonaceous clay often includes thin interbeddings or stringers of coal or lignite. As distinct from the inbedded bauxite, the clays overlying it are well-layered. The clayey cover is usually only 10 to 50 cm thick, a few metres as a maximum: it passes upward into marl and then into limestone. Sandstones and siltstones are remarkably frequent in the immediate cover; conglomerates and breccias are fairly so.

In the deposits where the cover develops from the bauxite in continuity, there are some peculiar, transitional rock types. The grey bauxite of the Eifelian bauxite horizon in the Northern Urals, for example, gradually passes upward into dark grey shale, marl and limestone. The transitional bauxitic marl and bauxitic limestone is 10 to 30 cm thick as a rule but attains 2 m in a few places (Bushinsky 1971). The zone of transition in the Ivdel deposits of this same region also includes black bituminous marl, bauxitic marl and black bauxitic limestone (Knyazeva 1966). In the Salair group of deposits, too, Eifelian bauxite passes gradually into the marine limestone overlying it. The zone of transition consists of a bauxitic limestone, 11 m thick on average, which locally includes some thin interbeddings of bauxite (Nagorsky 1958, Kharin 1969).

In some deposits of the Brignoles group, a continuous passage between bauxite and the overlying succession of clay, lignite, sandstone and limestone was observed by Nicolas (1968, 1970). The zone of transition is typically 10 to 50 cm thick.

The cover most widespread in the Hungarian deposit groups is Eocene clay. At Gánt, the most frequent cover sediment is a multicoloured clay; at Nyirád and at Szőc in the Cseres and Malomvölgy deposits, at Kislőd, Fenyőfő, Bakonyoszlop and Iszkaszentgyörgy, it is grey clay and dark grey carbonaceous clay rich in pyrite. The carbonaceous clay overlying the Nyirád and Malomvölgy deposits locally contains finely elutriated bauxite particles. These get scarcer upward, and disappear at a height of 2 to 3 m above the deposit. In the Alsópere and Tés regions, bauxite is overlain by brown to brownish grey Upper Aptian clay and clay marl containing Munieria baconica.

A carbonate cover is more rare. A Lower Barremian clayey–nodular limestone constitutes the cover of the Nagyharsány deposits. Upward, this passes into a pure thick-bedded limestone within a few metres. At Nézsa, Middle Eocene limestone follows above a yellow clay 5 to 20 cm thick. Occasionally, the bauxite is overlain directly by Eocene Miliolinid–Alveolinid limestone also in the Nyirád and Szőc regions.

In the Nagyegyháza–Csordakút–Mány basin, where bauxite is underlain in places by dolostone detritus, the immediate cover is also a redeposited dolostone detritus, 30 to 83 m thick (Sóki 1973). It is probably the material of a proluvial fanglomerate (oral communication by Tóth, 1975).

The Santonian cover of the central and northern parts of the Halimba Basin emerges from the bauxite with continuity. In these parts, the bauxite deposit itself contains interbeddings of dolostone and limestone gravel, as well as irregularly scattered dolostone and limestone pebbles. Both grow more abundant in the top portion of the deposit; beside them, finely dispersed authigenic dolomite particles appear in the matrix. This zone of transition, 1 to 5 thick, thus consists of dolomitic bauxite and bauxitic dolostone. It is overlain without a sharp limit by 10 to 30 m of dolomitic clay marl and marl, a limnic–lacustrine deposit, which in places contains interbeddings of dolostone and limestone conglomerate. No upward passage of a similar nature is known from any other bauxite deposit.

The writer uses the term '*secondary cover*' to denote sediments deposited on top of the bauxite after the erosion of an original ('primary') cover. For example, in parts of the Nyirád group of deposits, the primary Eocene cover was eroded, and bauxite is now directly overlain by Badenian (Miocene) gravel, clayey gravel and sand. Parts of the Fenyőfő–Bakonyszentlászló region, where the Eocene cover was similarly eroded, are now covered by Upper Oligocene clays, sands and marls.

3.2.4.2. FACIES OF THE IMMEDIATE COVER

The bauxite in most deposit groups is overlain by a transgressive sequence. The first term of this sequence—that is, the immediate cover—is most often of a subaerial, marsh or freshwater facies; it is successively covered by brackwater, lagoon and, farther up, by neritic deposits. More seldom, the immediate cover itself is of brackwater facies. Deposit groups where bauxite is directly overlain by marine beds are few and far between.

There are some bauxite regions which have never been inundated by the sea: hence, no transgressive cover could form above them in the first place. This situation is most

frequent in the French deposit groups. The deposits of the Alpilles, for example, are overlain by lacustrine limestone; those of the Haut-Var, by river-laid sand and sandstone. In the Brignoles region, bauxite is covered by a limnic clayey conglomerate, overlain in its turn by a lacustrine limestone.

In a few instances, the cover facies are regressive: in the Saint Chinian and Cazouls regions of France, the bauxite is overlain by brackwater clays, sands and marls which gradually pass into limnic and river-laid deposits. In the regions of Villeveyrac and Loupian, the brackish sandy clay and marl of the immediate cover is overlain by a lacustrine limestone.

3.3. TIMAN-TYPE DEPOSITS

This type of deposit has been distinguished by the writer on the basis of the constitution of the bauxite complex and of deposit morphology.

The bauxitic complex of this deposit type includes in addition to bauxite and bauxitic clay also significant amounts of clayey and detritic sediments. Claystones, siltstones, and sandstone are the most frequent accessory sediments. Conglomerates, breccias, iron-rich claystones, sedimentary iron ores and lignites are more rare. The share of bauxitic rocks within the bauxite complex ranges from 10 to 70%.

The most frequent deposit morphology is stratiform; large flat lenses are more rare. The bedrock surface is weakly to very gently karsted, with flat rises and gentle depressions. Within a deposit, the bauxitic rocks usually concentrate into a single horizon, more seldom into several ones, each horizon being made up of comparatively thin layers or smallish lenses. From the Kazakhstan-type deposits, which are of rather similar constitution, the Timan type is distinguished by much weaker bedrock karstification.

Timan-type deposits constitute a mere fourteen deposit groups, their aggregate tonnage nevertheless makes up 16% of world total karst bauxite tonnage. Of the deposit groups in question, six are in the East Asian, seven in the Uralo–Sibero–Central Asian bauxite belt; one is in the North American belt. No deposit of this type is known from the remaining belts (table 6). They were given the name 'Timan type' because it was on the deposits of the Timan Mountains that the most copious literature and the most detailed geological descriptions were available to the writer.

Some Timan-type deposits which overlie a more intensely karsted bedrock constitute a transition towards the Kazakhstan type of deposits. Such transitions are rare; passage into the Tikhvin type are much more common. The two most typical examples of the latter are the Naurzumsk group of deposits in Central Kazakhstan and the southeastern margin of the Coastal Plain in the US. Tikhvin-type deposits occur in the vicinity of Timan-type deposits in a number of other places, too. The Devonian bauxite deposits discovered in the Central Timan region (Demina 1977, Kirpal' 1977) are transitional towards the Ariège type.

3.3.1. DEPOSIT MORPHOLOGY

These bauxite deposits being composite in nature, it is necessary to consider the form of the *deposit as a whole*, as constituted by the bauxitic complex in its entirety, separately from the form of the *bauxite ore bodies*, as constituted by the bauxite and bauxitic rocks only.

In the eponymic South Timan deposit group, two deposit forms can be distinguished: (a) *stratiform deposits* 3 to 5 km long, (b) *flat lenses* of a few hundred metres diameter (Gulyanitsky 1969). The deposits are situated near flat rises of the bedrock surface, at a distance of 2 to 5 km. The stratiform deposits are quite close to the rises as a rule; the lenses are somewhat farther, closer to the deeper parts of the ancient relief. There are all manner of transitions between the two deposit forms.

The bauxitic complex is from 5 to 30 m thick above the flat rises of the bedrock; in the depressions on the other hand, its thickness may exceed 100 m. It falls into two parts: a lower sub-ore zone and an upper ore zone. The sub-ore zone usually has at its base a fine conglomerate in a sandy-clay matrix, overlain by arkosic sandstone, siltstone and sandy claystone. There are some rare lenticular interbeddings of bauxite in the upper part of this zone. The sub-ore zone fills out the depressions of the bedrock surface, where their thickness may attain 75 m. In that case, they tend to constitute the bulk of the deposit (Kirpal' and Tenyakov 1974).

The ore zone is a single stratiform body which in addition to bauxite contains clayey bauxite and bauxitic clay also. Its thickness ranges from 1 to 12 m; it is 4 to 6 m typically. Towards the edges of the deposits, the bauxite ore bodies gradually wedge out, or pass into a claystone. Above the rises of the bedrock surface, it overlies the bedrock limestone directly, whereas in the depressions it overlies the sub-ore sediments. This is clearly visible on Demina's (1971) geological profiles (fig. 38).

There are some vast stratiform deposits of 10 to 100 km^2 surface area in the Chinese bauxite regions. Their total thickness is 5 to 20 m, of which bauxite makes up 1 to 8 m. The bedrock is overlain by a few metres of shale, sandy or iron-rich shale or ooidic iron ore. Upward, these pass gradually into a bauxitic shale and thence into bauxite. The bauxite ore body is an uninterrupted layer within the deposit, just as in the Timan region. There are some carbonaceous interbeddings from 10 to 40 cm thick and from 300 to 400 m long in its upper part (Beneslavsky 1961). At the top of the deposit, there is a bauxitic shale.

The Chinese bauxite deposits contain less sandy–clayey accessory sediments than the Timanian ones. Microscopy has revealed the bauxite proper to contain a surprising amount of detritic quartz, mica and chlorite (Schüller 1957). This is another essential feature that distinguishes these deposits from the Mediterranean-type bauxites.

There are extensive stratiform deposits near Eufaula and Springvale on the southeastern margin of the Coastal Plain in the US, the bulk of which is made up of micaceous clayey sandstone and sandy clay, in a thickness of 20 to 25 m. Lenses of kaolinitic clay, of 100 to 250 m length and 20 to 25 m thickness, are embedded in it. It is in the cores of these lenses that the ore-grade bauxite occurs: there is a gradual passage from the bauxite into the kaolinitic clay, through clayey bauxite and bauxitic clay. The

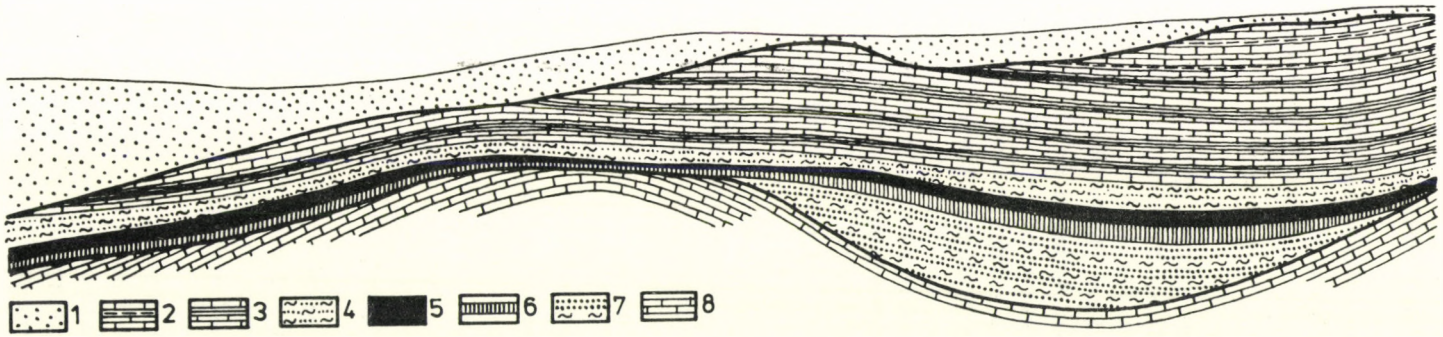


Figure 38. Profile of a bauxite deposit in the Southern Timan Mountains deposit group, USSR. After Demina (1971)

(1) Quaternary; (2) clayey–carbonatic beds, Middle Carboniferous; (3) limestone and clay beds, Lower Carboniferous; (4) multicoloured sandy–clayey beds, Lower Carboniferous; (5) carbonaceous clay with lenses of coal; (6) ore body part of bauxite deposit; (7) under-ore part of bauxite deposit; (8) limestone, Upper Devonian

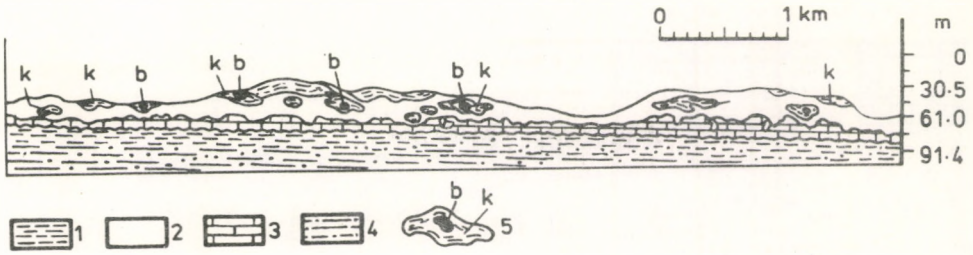


Figure 39. A stratiform bauxite deposit near Springvale, Georgia, US. After Clark (1965)
(1) clay, sand, Eocene; (2) Nanafalia formation (bauxitic complex), Lower Eocene; (3) bedrock limestone, Paleocene; (4) sand, clay, Upper Cretaceous; (5) b – bauxite, bauxitic clay, k – kaolinitic clay

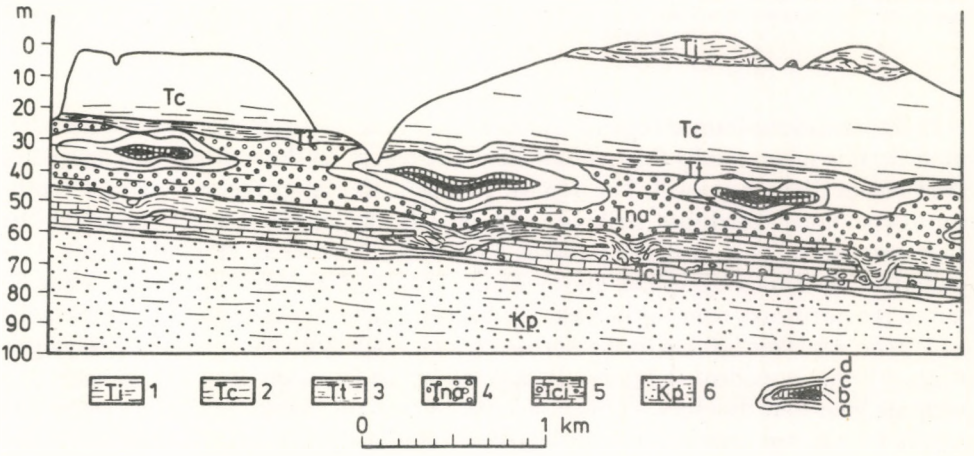


Figure 40. A stratiform bauxite deposit near Andersonville, Georgia, US. After Zapp (1965)
(1) clay, Eocene; (2) clayey-sand, Eocene; (3) sandy-clay, Eocene; (4) Nanafalia formation, Lower Eocene; (5) bedrock limestone grading upward into terra rossa, Palaeocene; (6) micaceous sand, clayey in places, Upper Cretaceous; (7) a – bauxite, clayey bauxite, b – bauxitic clay, c – kaolinitic clay, d – sandy-clay surrounded by clayey-sand

bauxite ore bodies within the lenses of clay are just 2 to 8 metres thick and at most 50 m long, roundish or elongate in outline. All this is apparent in the geological profiles of the Andersonville and Springvale groups of deposits (figs 39 and 40). Each of these ore bodies contains just 5000 to 10 000 tons of bauxite and somewhat more bauxitice clay (Clark 1965; Clarke 1966).

3.3.2. RELATION OF THE BAUXITE TO ITS BEDROCK

In the Southern Timan region, the bedrock of the bauxite is a neritic clayey limestone of Upper Devonian (Famennian) age. In the Chinese bauxite regions there are marine limestones of Cambrian, Ordovician or Devonian age, more seldom dolostones or dolomitic limestones.

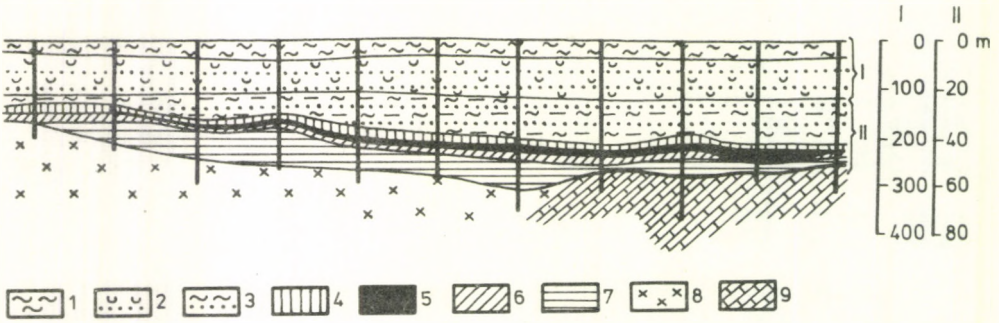


Figure 41. Geological profile of a bauxite deposit in the Naurzumsk group, USSR. After Kirpal' (1972b)
(1) clayey sand and sand; (2) sand, sandstone and diatomaceous earth; (3) kaolinitic clay, carbonaceous clay and sand; (4) bauxitic and bauxite-like clay; (5) bauxite, clayey bauxite; (6) bauxite-like clay and red clayey sandstone; (7) claystone, aleurolite and sandstone; (8) basalt, Upper and Middle Triassic; (9) limestone, Lower Carboniferous

In the two above-named regions, the carbonate bedrock succession is more than 100 m thick. It is substantially thinner in the bauxite region on the southeastern margin of the Coastal Plain in the US. The deposit groups of this latter region occupy a south-southwest–north-northeast striking zone of 270 km length. Near Eufaula, at the southwest end of this zone, the Paleocene bedrock limestone is 25 to 40 m thick. It is underlain in its turn by a thick clayey–sandy Upper Cretaceous succession. Some 50 km to the northeast, near Springvale, the bedrock limestone is reduced to between 20 and 30 m; a further 100 km to the northeast, near Andersonville, it is just a few metres thick. All this is evident in figures 39 and 40. In the Irwinton group of deposits, the bedrock limestone disappears altogether, and the bauxite complex directly overlies the Upper Cretaceous sandy clay. The deposits of this last group belong definitely to the Tikhvin type. This, then, is a case of gradual lateral passage between Timan- and Tikhvin-type deposits. The transition does not affect the thickness and lithologic constitution of the bauxite complex, but, in step with the gradual disappearance of the bedrock limestone, the ore bodies grow smaller and more clayey (Clark 1965; Clarke 1966; Lang *et al* 1965).

Another form of transition between Timan- and Tikhvin-type deposits can be observed in the Naurzumsk group of deposits in Central Kazakhstan. In the central part of this group, the bedrock is a Lower Visean limestone; it is surrounded by Lower and Middle Triassic basalt, dolerite and quartz porphyry. These formations, truncated by a common plane of erosion, are overlain by stratiform bauxite deposits of the Albian–Cenomanian age. This deposit has a single uninterrupted bauxite layer in its central part. The deposit overlies the limestone in its bulk, but overlaps onto the basalt also. This is clearly visible on a geological profile (fig. 41) by Kirpal' (1972b). The writer feels it not to be fortuitous that the bauxite body is thickest above the limestone bedrock.

In every group of deposits of the Timan type, the bauxite overlies the bedrock complex unconformably. The unconformity is mostly erosional, more seldom angular, as in the Timan Mountains proper.

The bedrock surface is weakly to very gently karsted—this is one of the principal distinctive features of the Timan-type deposits. Some karstification at least has been

described from every group of deposits, however. Subsequent karstification is not unknown, either; it is implied according to Clarke (1966) by slump and landslide features frequently uncovered in the open-cast pits working the Eufaula deposits.

At several deposit groups (e.g. Andersonville), the authors describe *in situ* accumulations of the bedrock's weathering products (fig. 40). According to Demina (1971), residual weathering products of the underlying limestone are often encountered in the bedrock depressions of the South Timan group of deposits too, in a thickness of 0·1 to 2·5 m.

3.3.3. RELATION OF THE BAUXITE TO ITS COVER

In every Timan-type group of deposits, the accumulation of the bauxite complex was followed directly by the deposition of some cover sediments. The cover is a conformable one in most cases, overlying the bauxite with a very brief hiatus.

In the South Timan group of deposits, the bauxite complex is overlain as a rule by a dark grey or black carbonaceous clay and by coal. These are generally 3 to 5 m thick, thinner above the rises of the ancient relief than above its depressions (fig. 38). Some authors lump this layer together with the bauxite complex, regarding the multi-coloured sandstone and siltstone layers overlying it as the first term of the cover sequence.

In the eastern part of the Central Timan, the bauxite complex is overlain by sandy and clayey beds: the carbonaceous clay and the coal are missing (Demina *et al* 1970). In both groups of deposits, the accumulation of bauxite was followed by a slow transgression.

In the Chinese bauxite regions, too, the immediate cover is typically a coal or bone (carbonaceous shale): the bauxite complex can in fact be regarded as the initial term of the coal measures. Sandy–clayey layers or limestone overlying the bauxite are much less frequent. According to Beneslavsky (1961), bauxite grade in the Gun-syan group of deposits is best where the immediate cover is a limestone, mediocre under the coal and bone, and poorest where the immediate cover is a sandy shale.

Near its bottom, the coal measure complex contains interbeddings of marsh and lacustrine limestone; at its top, it passes into a marine limestone as the result of a slow transgression. In Shantung and Liaoning provinces, the coal measures extend into the Upper Carboniferous and even the Permian, and contain bauxite layers in several horizons. Each of these is generally underlain by bone and overlain by sandstone (Beneslavsky 1961). These deposits, denoted A to G in the relevant literature, do not as a rule exceed one to three metres in thickness; they are nevertheless uninterrupted, traceable over 5 to 30 km. Their depositological features assign them incontrovertibly to the Tikhvin type of deposits. In these bauxite regions, then, a peculiar temporal succession from Timan- to Tikhvin-type deposits is observed: the Timan-type deposits at the bottom are overlain by seven horizons of Tikhvin-type ones, embedded in the overlying coal measures.

This is the succession observed on the edges of large sedimentary basins. Towards the basin interior, the coal measures pass into marine limestone; in step with that process, the bauxite layers also disappear gradually (Goretsky 1960; Wang Chung-Hsing 1965).

On the southeastern edge of the Coastal Plain of the US, near Eufaula, the beds of minor watercourses, filled with cross-bedded sand, were discovered on the top of the bauxitic complex. In places, the immediate cover is a lignite (Clarke 1966). In the Springvale area, the bauxite is overlain disconformably by a marine glauconitic sand (Clark 1965). The transgression was more rapid here than in the Eufaula area.

3.4. KAZAKHSTAN-TYPE DEPOSITS

In this depositological type, too, the carbonate bedrock is covered by a lithologically manifold but genetically coherent bauxitic complex. The bauxitic complex proper is made up largely of clays, sandy–silty clays, carbonaceous clays, lignites and clayey sands, more seldom of sands, sandstones, conglomerates and ooidic iron ores. Bauxite and bauxitic clay constitute several horizons of lenticular or nest-shaped ore bodies within this complex, most often its upper part. As distinct from the Timan-type deposits, the bedrock surface is karsted at least moderately, but most often strongly.

This type of deposit is unknown from the Mediterranean zone; this is why it is omitted from the comprehensive treaties written in Hungary and in Western Europe (e.g. Vadász 1951; De Weisse 1964; Valeton 1972). Its most typical occurrences are in the Kazakhstan region of the Soviet Union; hence the name given by the writer.

The type includes twenty groups of deposits altogether, fifteen in the Uralo–Sibero–Central Asian bauxite zone, four in the North American, and one in the East Asian zone. Their aggregate bauxite tonnage has been estimated at 636 million tons by the writer; this is 10% of world total karst bauxite tonnage. They are marked 'Kaz' in Appendix 1.

According to both the relevant literature and the writer's personal observations, the bauxite deposits in Missouri (US) constitute a transition towards the Mediterranean type, although they are closer to the Kazakhstan group in their basic features. In Appendix 1, they have been marked 'Med < Kaz'.

3.4.1. DEPOSIT MORPHOLOGY

The basis of classification, here as before, is the form of the entire bauxite complex; the bauxite ore bodies included within those may be of a variety of forms. In the Kazakhstan group of deposits proper, Kirpal' (1972a) distinguished four 'morphogenetic' deposit shapes. His classification is applicable with some complements to all Kazakhstan-type deposits, and the present writer has accordingly adopted it for use in this book. Of course, these four forms are merely the basic types: there is a considerable variety of transitional shapes.

3.4.1.1. SINKHOLE DEPOSITS

These deposits usually fill out oval, funnel- or cone-shaped karst depressions, usually round, more seldom of irregular outline. Their walls are very steep in places, vertical or even overhanging, in which case they are called *funnels* or *wells*. Kettle-shaped dolines of more gently sloping walls are also known. The bottom parts of these are not tapering as in a funnel, but rounded as in a kettle.

Their typical diameter at the top is from 30 to 300 m; their depth is from 30 to 200 m, much greater than in the Mediterranean-type sinkhole deposits. Some occasional sinkholes are up to 260 m deep, as described e.g. by Krivtsov (1968, 1969) from the Tatarsky group of deposits of the Yenisei River region.

Sinkholes are the most frequent deposit-form type. They are known in the bauxite regions of Chadobetsk and of the Yenisei River, and in the groups of deposits of Tselinograd, the Western Turgay and Kamensk–Uralsk. In the US, they are known in Missouri and from the Valley and Ridge Province.

A typical profile of the bauxite complex is as follows. At the sinkhole bottom there is usually a bed of sandy–silty clay, more seldom of sand, with lumps of igneous and metamorphic rock, locally of a hard, iron-rich bauxite or a bauxitic iron ore, embedded in it. This layer is overlain by multicoloured or grey clay composed of kaolinite and hydromicas, which in places contains lenses of bauxitic clay or clayey bauxite 1 to 2 m thick. In the top part of the sinkhole filling, the bauxite ore bodies proper are embedded in kaolinitic clay. Frequent interbeddings of carbonaceous clay and stringers of lignite tend to occur directly above the bauxite ore bodies or near the edges of the sinkholes. This is the succession seen on the geological profile of the Rock Run deposit of Alabama (fig. 42). In this same region, bauxite laterally passes into carbonaceous clay in some of the sinkholes (fig. 43).

The bauxite ore bodies are irregular masses as a rule, digitating towards the edges of the sinkhole. Their bottom surfaces copy the bedrock relief, more or less; their top surfaces are often gently depressed. Thickness is thus greatest above the central, most depressed part of the deposit; cf a geological profile of one of the deposits

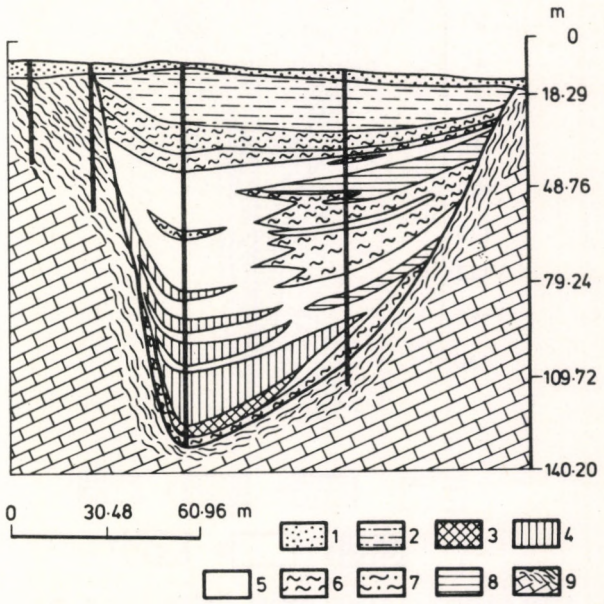


Figure 42. Bauxite deposit in a kettle-shaped doline near Rock Run, Alabama, US. After Cloud (1967)
(1) sandy, clayey soil; (2) sandy clay; (3) bauxite; (4) clayey bauxite; (5) bauxitic clay; (6) kaolinitic clay; (7) sandy kaolinitic clay; (8) carbonaceous clay with lignite stringers; (9) limestone, Lower Ordovician, with terra rossa on top

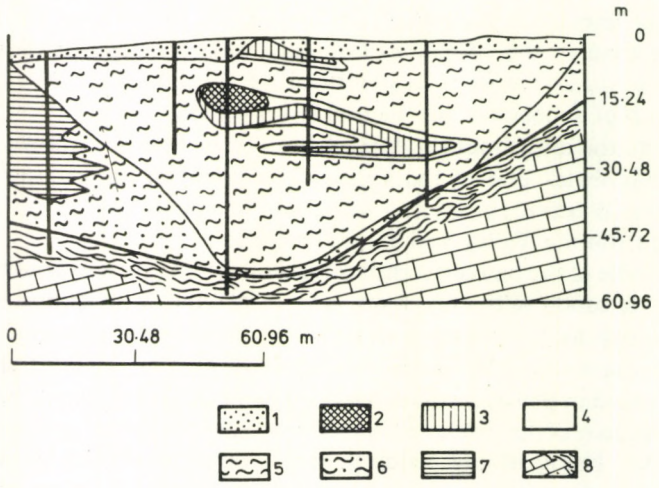


Figure 43. Bauxite deposit filling a kettle-shaped doline near Rock Run, Alabama, US. After Cloud (1967)
(1) sandy, clayey soil; (2) bauxite; (3) clayey bauxite; (4) bauxitic clay; (5) kaolinitic clay; (6) sandy kaolinitic clay; (7) carbonaceous clay with lignite stringers; (8) limestone, Lower Ordovician, with terra rossa on top

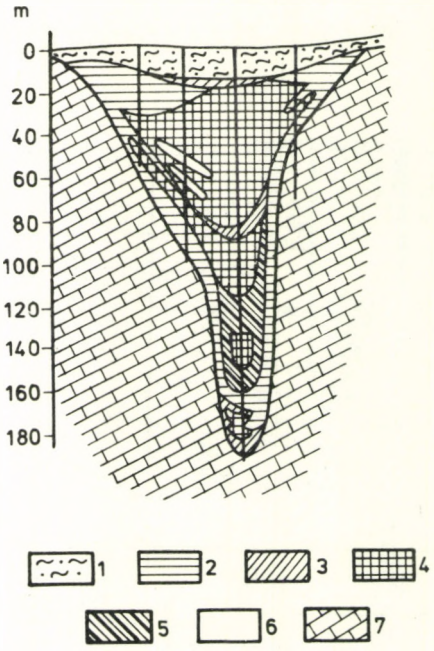


Figure 44. A pit-like doline containing a bauxite deposit; Srednye-Tatarsk deposit group of the bauxite region along the Yenisei River, USSR. After Krivtsov (1968, 1969)
(1) clay, sandy clay, Quaternary; (2) yellow and brown clay, Paleocene; (3) grey kaolinitic clay; (4) bauxite; (5) carbonaceous clayey bauxite and carbonaceous bauxitic clay; (6) bauxitic iron ore; (7) limestone, Proterozoic

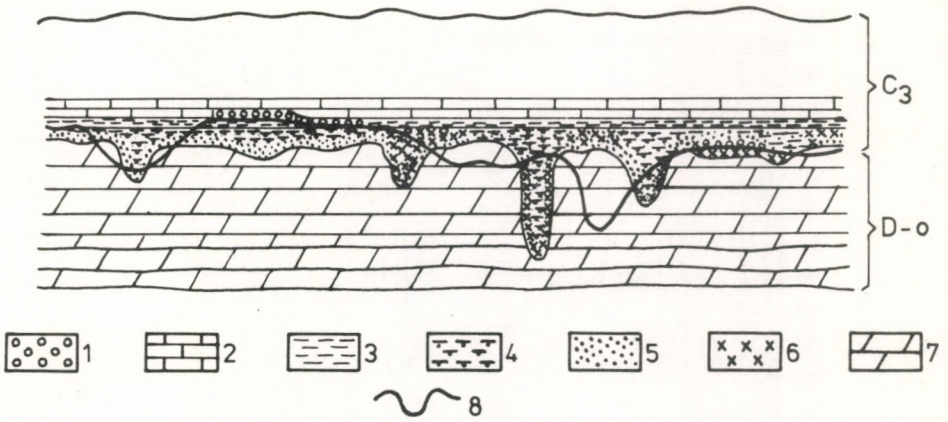


Figure 45. Geological profile of the bauxite-and-refractory-clay region in Missouri State, US. After Keller *et al* (1954)

(1) gravel; (2) limestone, Upper Carboniferous; (3) claystone cover; (4) bauxite and bauxitic clay; (5) sandstone; (6) clayey sandstone; (7) bedrock dolostone, Ordovician–Devonian; (8) present level of erosion

of the Srednye-Tatarsk group (fig. 44). The bauxite bodies are typically 5 to 10 m thick, but may attain 60 m exceptionally, e.g. in one of the deposits of the Ivanovsk group (Pel'tek 1971).

The bauxite ore bodies embedded in a sandy–clayey sinkhole filling in the sinkhole deposits of the Valley and Ridge Province are of regular funnel shape, meaning that their top surface also is markedly conical (Dunlap *et al* 1965).

In the Missouri State bauxite region, deep, funnel-shaped sinkholes alternate with flatter, kettle-shaped and lenticular ones (fig. 45). The bauxite ore bodies make up a comparatively large portion of the bauxite complex: clays, sandstones and conglomerates are confined to the bottom and to the margins of the complex. This is why the writer regards these as transitional towards the Mediterranean type of deposits.

3.4.1.2. GORGE OR CANYON DEPOSITS

These are from 100 to 600 m long and from 20 to 100 m wide. Their side walls, facing each other, are steep, almost vertical in places. They usually develop along structural planes. In numerous deposits of this form type, the gorge axis coincides with the tectonic contact between a carbonate rock and an igneous or metamorphic one: in that case, one of the side walls is sculptured in the carbonate rock, the other in the igneous or metamorphic rock. The writer calls them *contact-follower deposits* after Kirpal' (1972a).

Such deposits are most frequent in the Yenisei bauxite region and in the Western Turgay group of deposits (fig. 46). The irregular, rambling shape of the bauxite ore body is worthy of notice. In the Sokhatino group of deposits, especially vigorous karstification of the limestone bedrock sculptured a peculiar oblique gorge which followed the plane of contact with an adjacent amphibolite, so as to be overhung by the latter (fig. 47). This is somewhat of a freak case, however. A similar case was described

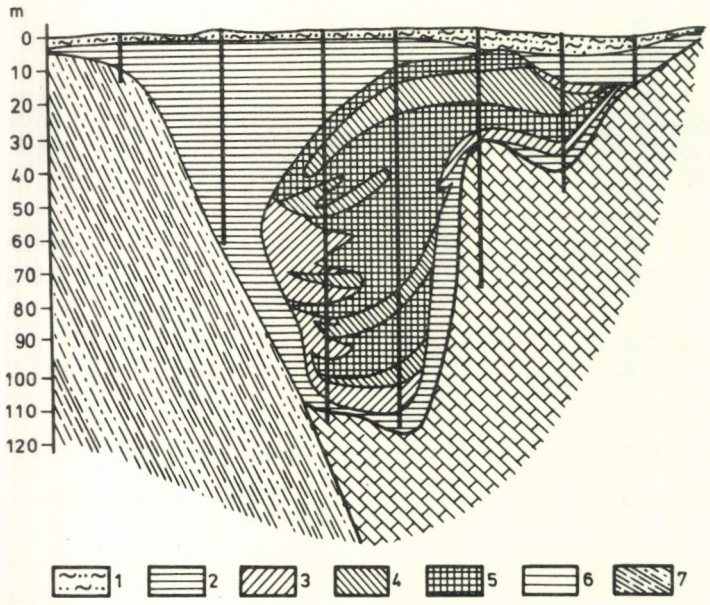


Figure 46. Cross section of a canyon-type bauxite deposit formed along a petrographic contact. Kirgiteysk deposit; bauxite region along the Yenisei River, USSR. After Krivtsov (1968, 1969)

(1) clay, sandy clay, Quaternary; (2) yellow and brown clay, Paleocene; (3) grey kaolinitic clay; (4) clayey bauxite; (5) bauxite; (6) limestone, Proterozoic; (7) phyllitised sandy shale, Proterozoic

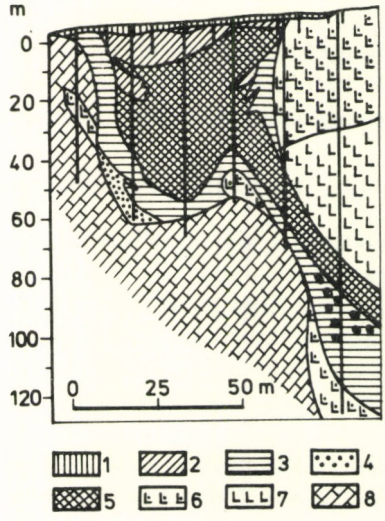


Figure 47. Bauxite deposit along a petrographic contact, with an amphibolite body overhanging it. Sokhatino deposit group, bauxite region along the Yenisei River, USSR. After Bobrov, Bushinsky (1971)

(1) sand, clay, Quaternary; (2) red bauxitic clay; (3) red clay; (4) quartz sand; (5) bauxite; (6) a clayey regolith formed in place; (7) amphibolite, Proterozoic; (8) marble, Proterozoic

by Kirpal' and Tenyakov (1974) from the Porozhnisk deposit of the Angara River group.

In structure, the bauxite complex is similar by and large to that in the sinkhole-type deposits. Highly irregular of form, the bauxite ore bodies are invariably elongate parallel to the gorge axis, with the line of maximum thickness also following that axis. The ore bodies are from 10 to 60 m thick; they wedge out rather steeply on both sides. In one of the deposits of the Western Turgay group, an uninterrupted bauxite profile 80 m thick was discovered (Venkov 1968, oral communication). This is the greatest continuous thickness of bauxite known so far in this group.

3.4.1.3. STRIP AND VALLEY DEPOSITS

These differ from the preceding type in their larger size and less vigorous karstification. One to five kilometres long and 100 to 700 m wide, they are typically between 10 and 50 m in depth. Local depressions deeper than 100 m have been encountered also in deposits of this type, however. They usually follow contacts between carbonate and alumosilicate rocks, and copy the meanderings of the contact line. More seldom, they follow fault planes, in which case they run more or less straight.

This deposit form is most frequent in the Arkalyk group of deposits. The surface of the dolomitic limestone bedrock is overlain in places by an *in situ* weathering residue 1 to 12 m thick, in which some kaolinitised brachiopods have been found. This layer in turn is overlain by 10 to 30 m of stratified, yellow and grey micaceous clay, in places with some thin interbeddings of carbonaceous clay. The ore bodies are embedded in the middle and upper part of the bauxite complex. They are made up of pure bauxite and bauxitic clay in repeated alternation, separated here and there by intercalations of carbonaceous, clayey bauxite and lignite. Typically 3 to 20 m thick, the ore bodies exceptionally attain 65 m in thickness (Krivtsov 1968, 1969). The largest uninterrupted ore body is 2 km long and 500 m wide. The ore bodies tend to be thickest above the contact line of the carbonate with the alumosilicate rock; their top surface usually slopes towards the dolomitic limestone. In parts of the deposit group, the bauxite passes laterally into a white, low-iron bauxitic clay. The ore bodies are overlain by 1 to 20 m of multicoloured, kaolinitic clay.

Valley-shaped deposits of this type are subordinate in number in some of the deposit groups of the Yenisei River bauxite region, and also in the Mugaysk deposit group of the Central Urals. At the latter locality, there is a deposit of 3 km long, 150 to 400 m wide and 100 to 150 m deep. The bauxite complex includes several lenticular ore bodies (Tkachenko and Bol'shun 1971).

3.4.1.4. DEPOSITS OF COMPOSITE KARST DEPRESSIONS

These are extensive deposits of irregular, oval or elongate outline. Typically, their depth is 10 to 60 m only, but their bottom relief is enlivened by numerous secondary depressions of kettle, sinkhole and gorge form the depth of which may attain 100 and even 150 m.

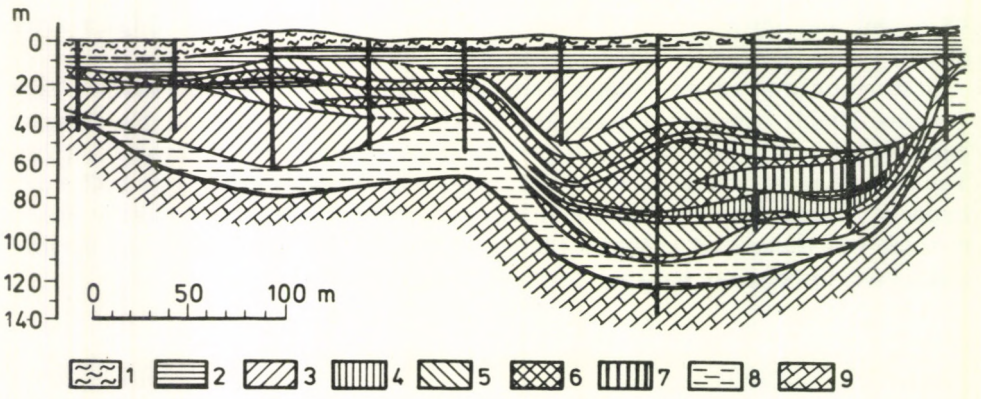


Figure 48. Valley-shaped bauxite deposit. Arkalyk deposit group, Central Kazakhstan bauxite region, USSR. After Krivtsov (1968, 1969)

(1) multicoloured clay, Neogene; (2) multicoloured clay, Paleocene; (3) multicoloured kaolinitic clay within the bauxitic complex; (4) bauxitic clay; (5) clayey bauxite; (6) bauxite; (7) carbonaceous clayey bauxite and lignite; (8) yellow kaolinitic clay; (9) limestone and dolostone, Devonian

This form type is most frequent in the Krasnooktyabrsk deposit group of the Western Turgay region. The deposit shown in figure 48 is 1900 m long and 1000 m wide. The ground plan shows the bedrock contours which reveal a number of local sinkholes and also an elongate depression on the left-hand side of the drawing.

The profile of the bauxite complex is similar to those of the foregoing form types. The lenticular ore bodies are embedded into the top part of the bauxite complex, they tend to be thicker above depressions in the bedrock relief. They are frequently overlain by carbonaceous clay and lignite. There are, in places, interbeddings of lignite and carbonaceous clay also within the ore bodies proper, as cf e.g. figure 49.

There are some similar bauxite deposits in the Ivanovsk deposits group of the Yenisei River region (fig. 50), where sandy and sandy–clayey rocks play a greater role in making up the bauxite complex.

3.4.2. DEPOSIT PATTERNS

The deposits tend to be arranged in ranges or strings 5 to 500 km long. The strings follow contacts between carbonate and alumosiliacete rocks, more or less; some are just close by, but others straddle the contact line. The long axes of the gorge, strip and valley deposits in a string are parallel by and large to the contact, or may even coincide with it.

Deposit groups of this type are most frequent in the Western Turgay. In figure 51, the map of the Temirsk group of deposits is presented as an example: most deposits are seen to follow the contact line. Almost the entire group of deposits at Arkalyk is constituted by a string of strip and valley deposits following a contact. There are no deposits at all on the side of the bedrock limestone (fig. 52).

Deposit groups whose major and minor axes are of about equal length ('equiaxed' groups) are unknown from the Kazakhstan-type deposits. The deposits of the Salair–

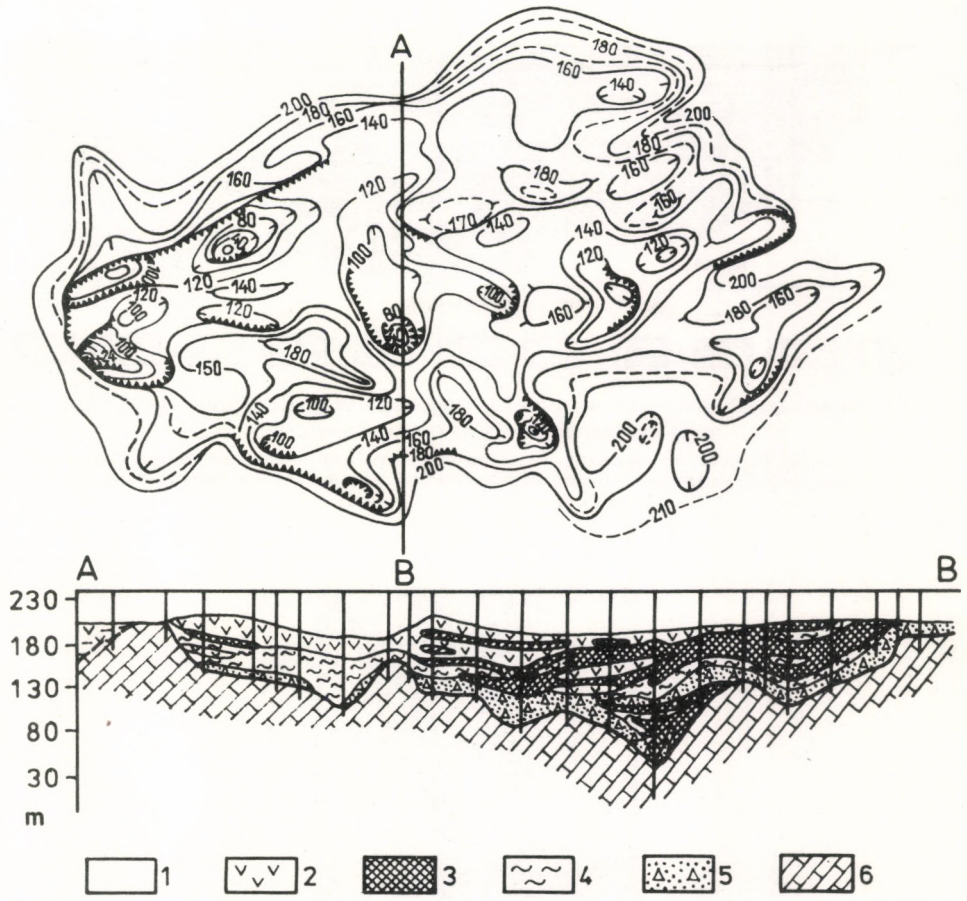


Figure 49. Bauxite deposit filling out a composite karst depression. Krasnooktyabrsk deposit group, Turgay Plateau, USSR. After Kim (1970)

(1) sandy-clay cover, Tertiary; (2) carbonaceous clay and lignite; (3) bauxite and clayey bauxite; (4) bauxitic clay; (5) kaolinite clay and kaolinitic–hydromica clay; (6) limestone, Carboniferous

Barzassk group cover an oblong area with no well-defined ranges or strings. In the Krasnooktyabrsk deposit group, the entire outcrop area of the bedrock limestone is pockmarked with bauxite deposits. This group is 25 km long from northeast to southwest and 1 to 6 km wide in the perpendicular direction. This is a fairly drawn-out arrangement, but no string by any means. Deposit density is characterised by the following figures: the group comprises 24 deposits including 527 ore bodies, 127 of which contain industrial-grade ore (Kirpal' 1972a; Kirpal' and Tenyakov 1974).

The frequency distribution of deposit forms is widely different from one bauxite region or deposit group to the next. In the Arkalyk group, for example, deposits are almost exclusively strip- and valley-shaped; sinkhole deposits are very rare. In the Western Turgay, on the other hand, the deposits are sinkhole-type, in their majority, whereas contact-follower gorge-like deposits are subordinate. Finally, in the central

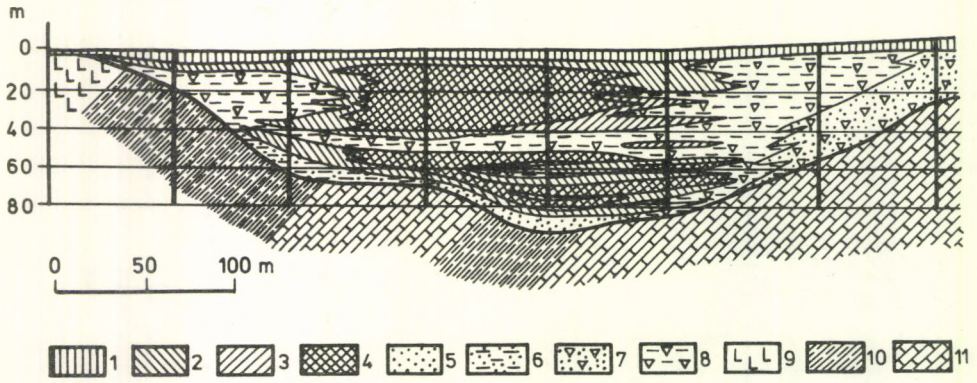


Figure 50. Bauxite deposit filling out a composite karst depression. Ivanovsk deposit group; bauxite region along the Yenisei River, USSR. After Bobrov (1964)

(1) sand, clay, Quaternary; (2) bauxitic clay; (3) clayey bauxite; (4) bauxite; (5) quartz sand; (6) clayey sand; (7) sand with quartz pebbles; (8) sandy clay; (9) amphibolite, Proterozoic; (10) shale, Proterozoic; (11) limestone and dolostone, Proterozoic

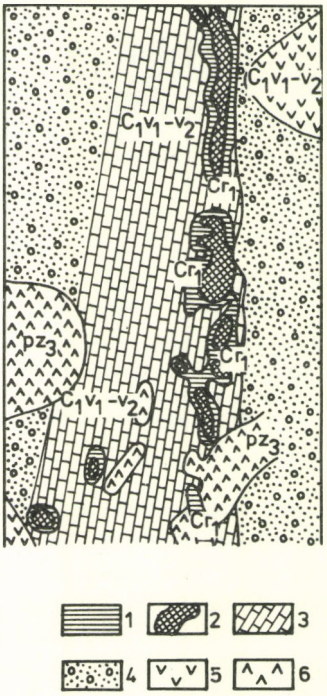


Figure 51. A string of bauxite deposits. Temirsk deposit group, Turgay Plateau, USSR. After Krivtsov (1968, 1969)

(1) clay; (2) bauxite; (3) limestone, Lower Carboniferous; (4) tuffaceous sandstone, Lower Carboniferous; (5) porphyrite, Visean; (6) diorite, Late Paleozoic

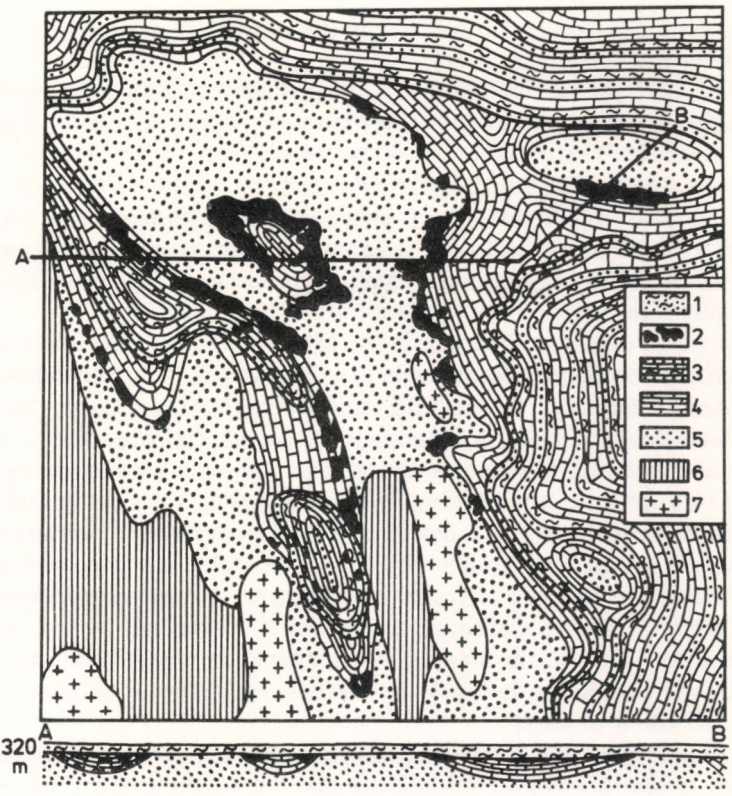


Figure 52. Arrangement of the bauxite deposits at Arkalyk, USSR. After Kripal' (1972a)
(1) Tertiary clays and sands (in profile only); (2) bauxite deposits; (3) limestone, claystone, aleurolite and sandstone, Lower Carboniferous; (4) limestone and dolomitic limestone, Lower Carboniferous–Upper Devonian; (5) sandy shale, Upper Devonian; (6) quartzite, shale, gneiss, gneissose granite, Precambrian; (7) granite

part of the region, there are the large composite karst depressions of the Krasnooktyabrsk group.

In the US, there are no contact-follower gorge, valley or strip deposits at all. Most deposits are of the sinkhole type, often of kettle and cauldron shape; that is, they fill out gentler karst depressions. There are also a few more extensive karst basins (poljes).

3.4.3. RELATION OF THE BAUXITE COMPLEX TO ITS BEDROCK

In the Kazakhstan type of deposits, limestone is not the exclusive bedrock of the deposits. This is most evident in the contact-follower deposits where, by definition, carbonate rock, on one side only; on the other, it is partly alumosilicate rock. They nevertheless are true karst bauxites, as the gorge- or valley-shaped depressions containing the deposits were sculptured by the karstification of the carbonate rock. The contacts involved are tectonic in their majority, and structural deformation is known to

enhance karstification: this is probably why the bedrock is most vigorously karsted precisely along the tectonic contacts.

Often, alien rock types are interbedded in the carbonate rocks underlying the bauxite deposit. In the Arkalyk deposit group, for example, a few metres of claystone and siltstone ('aleurolite') are intercalated between beds of on Upper Devonian (Famennian) dolomitic limestone. In the Yenisei River region, a Proterozoic bedrock limestones include interbeddings of slate and mica schist (fig. 50). In the Krasnooktyabrsk deposit group, there is an intercalation of andesite tuff and andesite porphyrite in the bedrock limestone which is, in addition, penetrated in places by granite porphyry intrusions (Venkov and Fatkhudinov 1966).

The Ordovician dolostone bedrock in the Missouri State region abounds in chert nodes, which in places aggregate into beds 10 to 20 cm thick. Thin intercalations of clay have also been encountered. In the Valley and Ridge Province region, the Cambrian and Ordovician dolostones and limestones underlying the bauxite are also cherty; they locally contain clayey impurities in addition. The high-purity limestone or dolostone bedrock so widespread in the Mediterranean type, then, is decidedly rare here.

The hiatus between the bedrock and the bauxite complex is very long as a rule. In the Arkalyk deposit group, for example, the bauxite complex is Palaeocene and Lower Eocene, whereas the bedrock is Upper Devonian to Lower Carboniferous. In the Western Turgay, a Lower Carboniferous bedrock is overlain by a Cenomanian–Turonian bauxite complex. The hiatus is even longer in the Chadobetsk bauxite region, where a Lower Cambrian bedrock is overlain by Albian–Cenomanian bauxite (Senonian according to some).

In most places, there is a marked *angular unconformity* between a steeply dipping bedrock and a near-horizontal bauxite complex. As a result, the above-mentioned alumosilicate interbeddings of the bedrock run up against the bauxite with their edges rather than with their bedding planes, a fact of great relevance to karstification.

In places, the *in situ* weathering residue of the bedrock is still present between the bedrock surface and the bauxite complex; it tends to be clayey or silty. In the Missouri State and the Valley and Ridge Province regions, it contains the chert nodes weathered out of the bedrock limestone. The residual layer is 1 to 5 m thick typically, and 10 m thick as a maximum; downward it passes with no sharp limit into the unaltered carbonate rock (figs 40, 42 and 43). According to Krivtsov (1969), such an *in situ* weathering residue, a few metres thick, is found in places also in the groups of deposits along the Angara and at Chadobetsk. In the Arkalyk deposit group, its thickness attains 12 m in places. The writer has had the occasion to inspect such a weathering profile, in which traces of original bedding in a Frasnian shale and clayey sandstone could still be recognised. It seems as though, in this deposit group, most of the weathering residue had been derived from the clayey and silty interbeddings in the limestone complex.

Karstification is comparatively strong in the Kazakhstan type of deposits. Of the 20 deposit groups, there is only one—the Chimkent group in Soviet Central Asia—where the scarce data of the relevant literature allows the interference the karstification of the bedrock relief to be just moderate to weak. One other, minute deposit group exhibits moderate karstification. In 9 groups, the bedrock is moderately to strongly karsted. These include the largest deposit groups—e.g. Arkalyk and the Western Turgay. In

three minor deposit groups, the bedrock is karsted strongly. Finally, there are six groups where the bedrock is strongly karsted in places and very strongly karsted elsewhere (Tselinograd, Chadobetsk, and four groups along the Yenisei River). The sinkholes, funnels, gorges and canyons there are much deeper (from 100 to 260 m) than on even the most vigorously karsted bedrocks of Mediterranian-type deposits.

The Soviet specialists are much intrigued by the problems of evolution posed by these karst forms of extraordinary depth. There seems to be a consensus in that they are unlikely to have fully developed prior to the deposition of the bauxite complex. More probably, karstification must have been comparatively slight early on in the deposition process, and the relief only gently undulous. It gained momentum after the filling up of the depressions, in structurally fractured zones above all, along faults (especially at their crossing), where the shattered rock provided better drainage conducive to stronger leaching. According to Kim (1970), there are some large composite karst depressions on the Turgay Plateau, each of which corresponds to a fault block.

The rates of karstification and of deposition were not uniform everywhere. The sinkholes ceased to grow deeper whenever the ground water table rose high enough to attain their bottom. When it rose even higher, the sinkhole bottom was swamped, and a marsh developed on top of the bauxite complex—this is how the above-mentioned interbeddings of carbonacous clay and lignite came to exist. Once an uplifting had again depressed the ground water table, the rate of accumulation could increase, and so could the rate of sinkhole sculpture. This is how Pel'tek (1971) accounts for the formation of two or more bauxite horizons and of the layers of carbonaceous clay overlying them in the Yenisei River bauxite region. In his opinion, karstification was again enhanced after a general emersion of the region in the Neogene and Quaternary, as proved by numerous traces of collapses and landslides visible within the bauxite complex. The writer attributes a special significance to Pel'tek observation that the karst dolines and gorges reach down to about the same bottom level (140 to 160 m above the sea), regardless of the present-day elevation of the ground surface. This level may correspond to the deepest ground water table, below which the karstification process could not penetrate.

Traces of subsequent karstification can be observed also in the Valley and Ridge Province. According to Warren *et al* (1965), in the deposits near Spottswood, for example, large angular blocks of bauxite came to be embedded into bauxitic clay and kaolinite as a result of karst cavity collapse and slumps.

The evolution of the funnel-shaped bodies mentioned in connection with the Valley and Ridge Province are explained by Knechtel (1963) in terms of the succession in Figure 53, by the collapse of karst caverns and cavities. According to him, Phase 1 of this evolution produced a residual weathering product (a terra rossa). Subsequent karstification gave rise to caverns (phase 2a), which collapsed and were filled with terra rossa as the process progressed. Leaching was strongest next to the side walls of the collapsed cavities; it converted the terra rossa into clayey bauxite (phase 2b). Bedrock leaching continued, lowering the bedrock relief. The former cavity thus gradually came to be surrounded by terra rossa on all sides, which helped to preserve the original funnel shape (phase 3). Further dissolution continued the lowering of the relief and of the ore body proper (phase 4). Although the explanation of the funnel-shaped ore bodies by cavity collapse seems correct, the interpretation in terms of the above four

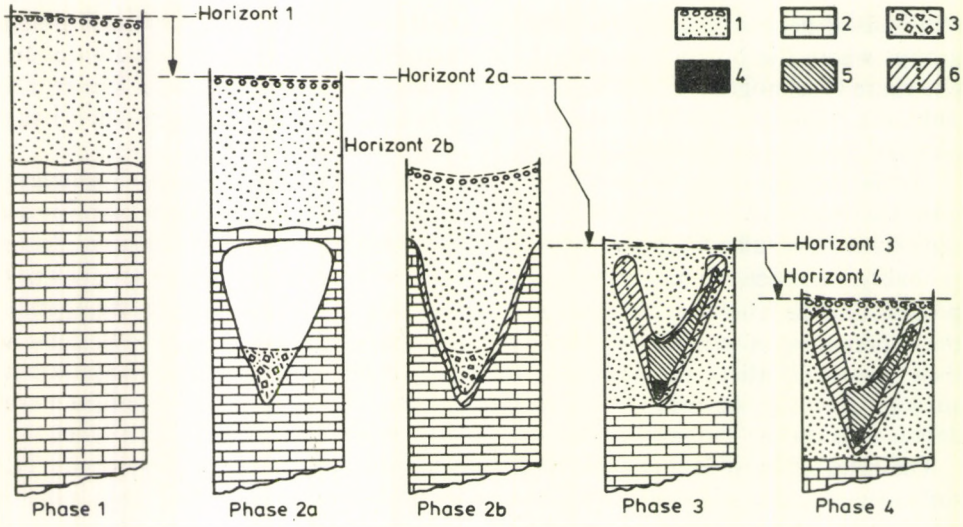


Figure 53. Evolution of a funnel-shaped bauxite body according to Knechtel (1963)
(1) weathering residue on a limestone (terra rossa); (2) limestone, clayey limestone; (3) coarse detritus; (4) bauxite; (5) clayey bauxite; (6) bauxitic clay

phases seems overly rigid. The assumption that the entire ore profile was formed *in situ* is contradicted by signs of transport and redeposition in the bauxite complex.

In sum, the role of subsequent karstification was greater in the majority of Kazakhstan-type deposits than in the Mediterranean-type deposits. The bedrock relief seen today is mostly a pseudo-paleorelief sculptured after bauxite deposition.

3.4.4. RELATION OF THE BAUXITE COMPLEX TO ITS COVER

Most of the deposits in the Valley and Ridge Province, in the Yenisei River region and in the Chadobetsk group are covered by just a few metres of Quaternary and recent sand, clay and regolith. In a geohistorical comparison, it seems probable that these deposits have been on the surface or close to it up to the present.

In the rest of the regions, the deposits carry some sort of cover which invariably overlies the bauxite unconformably; no incontrovertible conformity has been established anywhere. There is no angular unconformity between the deposits and their cover, however: the disconformity is usually indicated by a sharp contact with the cover and, in some deposit groups, by the unevenly eroded surface of the deposits. A case in point was described by Krivtsov (1968, 1969) from the Krasnooktyabrsk deposit group in the Western Turgay.

The immediate cover tends to be sandy–clay in its bulk; it includes carbonate rocks only in the Missouri region, and even there they are subordinate.

In six deposit groups, the bauxite is overlain by continental deposits: transgressive sequences could be identified above four groups only. Of the deposit groups with a continental cover, the Arkalyk group is the most important. In certain depressions

within that group, the bauxite complex is covered by 0·4 to 10 m of coarse-grained quartz sand or friable sandstone, overlain in turn by 10 to 100 m of multicoloured clay, clayey silt and sandy clay of freshwater, and locally brackish water facies. Wherever the multicoloured clay directly overlies the bauxite complex, its bottom one to five metres include bauxite pebbles and boulders of 10 to 40 cm diameter.

Also in the other deposit groups overlain by continental deposits (e.g. Tselinograd, Salair–Barzassk, Orsk), the immediate cover is a multicoloured clay as a rule, overlain in its turn by layers of clayey sand, silt and sandstone.

In the Western Turgay region, the accumulation of bauxite complex was followed by a slow transgression. The beginning of subsidence is indicated by some layers of carbonaceous clay and lignite in the top part of the bauxite complex (fig. 49). These are overlain within a sharp limit by marine strata, with abrasional gravel, sand and, locally, bauxite pebbles at their base (Krivtsov 1968, 1969), followed by a well-stratified marine clay. There is in the neighbouring Mugojar deposit group a transgressive sequence identical by and large with this one.

In the Alapaevsk deposit group on the eastern slopes of the Central Urals, the immediate cover is also a lignite or carbonaceous clay implying a swamp facies, overlain first by lacustrine–fluvial sandy clays, and then by neritic glauconitic sand.

In the Missouri State bauxite region, the bauxite complex used to be overlain directly by 1 to 3 m of neritic greenish-grey illitic clay, followed similarly by a few metres of marine nodular, clayey limestone. North of the region, the higher terms of the Upper Carboniferous are also preserved likewise in neritic facies. Keller *et al* (1954) stated these to have been originally present also in the bauxite region, to be eroded in the course of a subsequent uplifting. According to those authors, some of the bauxite complex also fell prey to this erosion. This hypothesis is also implied by their geological profile, figure 44.

3.5. ARIÈGE-TYPE DEPOSITS

These deposits are of a characteristic bipartite structure. The lower unit consists of sedimentary clays, more seldom of clay-marls or marls, which pass upward without a sharp limit into a bauxite through a bauxitic clay. The whole complex overlies the more or less karsted relief of carbonate rocks. The most typical representatives of this type are in the Ariège region on the north slope of the Pyreneen Mountains, this is why the writer has chosen this name for them. The Ariège deposits were studied in great detail by Combes who published his findings in a monograph (1969). He proved that the bauxite had formed by the *in situ* lateritisation of the clay or clay marl, but that the carbonate bedrock had also played a role. This type, then, is transitional between the karst and the lateritic bauxites.*

This group includes a mere four deposit groups, all in the Mediterranean zone. In addition to the eponymic Ariège group, and to the neighbouring deposits of Saint Paul de Fenouillet, the writer has placed here the Triassic bauxite deposits of Lika country, Yugoslavia, although some deposit features there appear to be transitional towards the

* In the upper part of the deposits bauxite pebbles of 0·5 to 3·0 m in size also occur. This indicates that redeposited bauxitic material also took part in the accumulation of the deposits.

Mediterranean deposit type. The Triassic deposits of Slovenia, Yugoslavia are of a decidedly Mediterranean character, but some of their features still exhibit a certain affinity with the Ariège type. Finally, despite its peculiar structure, the writer has placed in this type the small bauxite deposit of the Greek island of Skopelos.

The aggregate bauxite tonnage of the Ariège-type deposits is a mere 2% of the total world karst bauxite tonnage. Their distinction is justified on the strength of their peculiar depositological position rather than on account of their size.

3.5.1. DEPOSIT MORPHOLOGY

In Ariège, the lenticular deposit form is the most frequent. The lenses range from 100 to 500 m in length, from 100 to 200 m in width and from 4 to 15 m in thickness. Their bottom two-thirds is clayey-marl; the top third is bauxitic. The lenses are arranged close-spaced in a strip of some 40 km length. In the absence of systematic drilling exploration, the width of the strip is not known, but surface geology implies that it probably does not exceed 1 to 2 km. This deposit type has the remarkable feature that the above-described deposit profile may repeat itself, either with a thin limestone layer intercalated between the terms, or without any intercalation.

In places, next to the lenses close to the outcrop line some sinkhole-shaped deposits of 10 to 50 m diameter and 10 to 25 m depth. In these, the above-described regular profile of clay and bauxite is missing: there are instead, blocks of several cubic metres size of hard bauxite embedded in a soft laminated clay. This profile may clearly be the result of slumps and slides connected with subsequent karstification.

The Lika bauxite deposits are larger than those of the Ariège. The stratiform Rudopolje deposit has a ground plan area of more than one square kilometre. The deposits of Skoćaj and Vrace also exceed 0·5 km^2 in ground plan area. These large stratiform deposits are accompanied by minor lenticular ones.

The Lika bauxite region was monographically studied by Šinkovec (1970). Boreholes have revealed the deposits to be contained in flat saucer-shaped depressions in the bedrock limestone, as shown e.g. in the geological profile of the Vrace deposit (fig. 54). At the base, there is a layer of greenish grey claystone, 5 to 7 m thick, which gradually passes upward into 3 to 5 m of bauxitic clay and bauxite. The beginning of the next cycle of sedimentation is marked by a limestone conglomerate of 1 to 2 m thickness, overlain by 4 to 5 m of claystone. This passes upward into 1 to 2 m of bauxite. The third cycle starts with 2 to 8 m of limestone conglomerate, and closes with bauxitic clay and bauxite. The aggregate thickness of the entire complex is 15 to 30 m, with a gradual pinch-out towards the deposit edges. The Lika bauxites are the products of the *in situ* lateritisation of the basal claystone layers. The claystone in places contains reworked bauxite debris, ooids and pisoids; this is a Mediterranean feature already.

In the Skopelos Island bauxite deposits, the surface of a Liassic dolostone is overlain according to Papastamatiou (1964) by a few metres of basic tuff and tuffite, likewise Liassic. Locally lateritised, this layer includes small bauxite ore bodies of nest and lens shape. The largest lens, the one at Georgareika, is 1·5 m thick and 270 m long. According to Papastamatiou, bauxitisation is most advanced where the bauxite is in immediate contact with the bedrock limestone. He observed a continuous passage from tuffite into bauxite. The parent rock, then, is different from those of the Ariège and

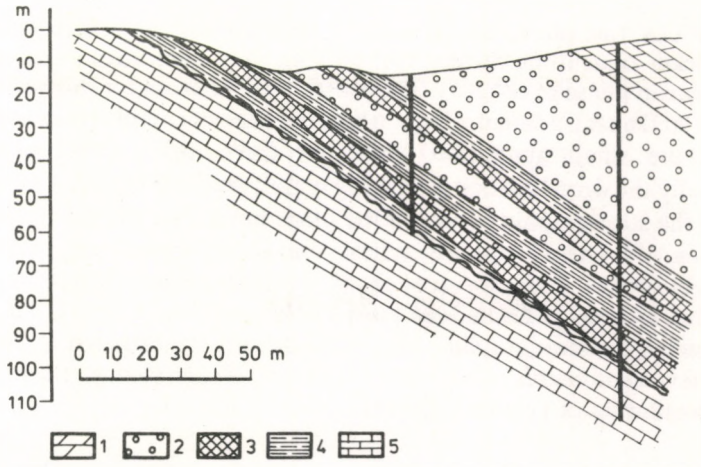


Figure 54. Geological profile of the Vrace bauxite deposit, Lika County, Yugoslavia. After Šinkovec (1970)
(1) dolostone, Norian; (2) limestone conglomerate; (3) bauxite; (4) clay and bauxitic clay; (5) limestone, Ladinian

Lika regions: the deposit nevertheless has its *in situ* lateritisation, combined with the influence of a carbonate bedrock, in common with the other Ariège-type deposits.

Deposit arrangement is strip-like in the Ariège, and rather more field-like in Lika and on Skopelos Island.

3.5.2. RELATION OF THE BAUXITE TO ITS BEDROCK

Formed under totally different conditions, the clays, marls or tuffs are not parts of the bauxite complex in the Ariège-type deposits. Genetically, these beds constitute the immediate bedrock, although there is a gradual passage between them and the overlying bauxitic clay and bauxite. The zone of transition is not thicker than one metre as a rule.

The carbonate complex which is overlain disconformably by the immediate bedrock is called the '*deep bedrock*' by the present writer. It is made up by several hundred metres of high-purity neritic nearshore limestone and dolostone in the Ariège and at the Saint Paul de Fenouillet region, and of several hundred metres of Anisian–Ladinian limestone and dolostone in the Lika region. In the latter region, there are thin interbeddings of tuff, tuffite, and clay in the Anisian layers and of sand or sandstone in the Ladinian limestone beds. At Rudopolje, the Ladinian limestone is underlain by beds of a tuffitic siliceous clay and marl. In places, the limestone was totally eroded, and the deposit complex is directly underlain by these tuffitic beds.

The *degree of karstification* is highly varied. That of the deep-bedrock relief is moderate to weak under the lenticular and stratiform deposits of all four deposit groups. Under the small sinkhole deposits of the Ariège region, on the other hand, a marked, dissected karst relief can be seen; it is a product of subsequent karstification. The chaotic mixing of the harder bauxite with the softer clay beds originally underlying it is due to slumps subsequent to the erosion of the karst relief. According to the writer's

own observations, this phenomenon is most marked near the outcrop line, under a cover not thicker than 10 m. Subsequent karstification continued in the Ariège also after the Pyrenean phase of orogeny. The most excellent proof for this was discovered by Combes (1969) near Cadarcet village, where the deposit complex is in an overturned position. In the small abandoned open pit there, the contact with what is stratigraphically the deep bedrock is a gently undulating, weakly karsted surface. The contact with the stratigraphic cover, on the other hand, exhibits strongly sculptured karst forms (fig. 55). These forms are vertical; that is they are not perpendicular to the bauxite deposit surface. These forms probably came into existense in the Quaternary, at a time when erosion had removed most of the covering deposits. The Urgonian limestone which is the stratigraphic cover of the bauxite is weathered, pulverulent in a depth of a few centimetres next to the contact. Pulverulency, then, is likewise secondary, postdating of necessity the episode of overturning.

3.5.3. RELATION OF THE BAUXITE TO ITS COVER

The deposits are overlain unconformably by the covering strata. This must be so since no *in situ* bauxitisation would have been possible otherwise. The hiatus, however, is remarkably short. The deposit top is a smooth, almost plane surface everywhere.

In the Ariège region, the bauxite is overlain in most places by 20 to 80 cm of yellow, brown or grey clay, locally with thin stringers of lignite embedded in it. It is in its turn overlain by a biogenic neritic-nearshore limestone. In the southeastern part of the bauxite region, near Péreille, the clay constituting the immediate cover thickens to 2 to 3 m, and the lignite embedded in it attains 10 to 30 cm. A swampy environment is signalled by disseminated pyrite and marcasite grains. The neritic limestone is introduced here by 10 to 40 cm of brackish water marl. The cover in the Saint Paul de Fenouillet group of deposits is similar.

In the Lika group of deposits the cover succession is transgressive as well, with the difference that there is a breccia and conglomerate, 5 to 40 m thick, of freshwater limestone at its base (fig. 54). Its lower part consists of comparatively fine limestone debris in a clayey matrix. The limestone debris grow coarser and the matrix becomes increasingly calcareous upward. The conglomerate is overlain by neritic Norian 'Hauptdolomite'.

On Skopelos Island, the bauxite is overlain by a few metres of shale and of dolostone conglomerate followed by neritic dolostone.

To summarise, the deposits were inundated after a shortish hiatus by a shallow sea in all four deposit groups.

3.6. SALENTO-TYPE DEPOSITS

During the geological periods subsequent to bauxite formation, the beds covering certain deposits came to be eroded; the bauxite was exposed and was itself eroded in places. In most cases in the course of transportation, the eroded bauxite got mixed with other sediments, underwent mineralogical alteration, and lost its bauxitic character. Known cases of eroded bauxite debris accumulating in depressions of a limestone

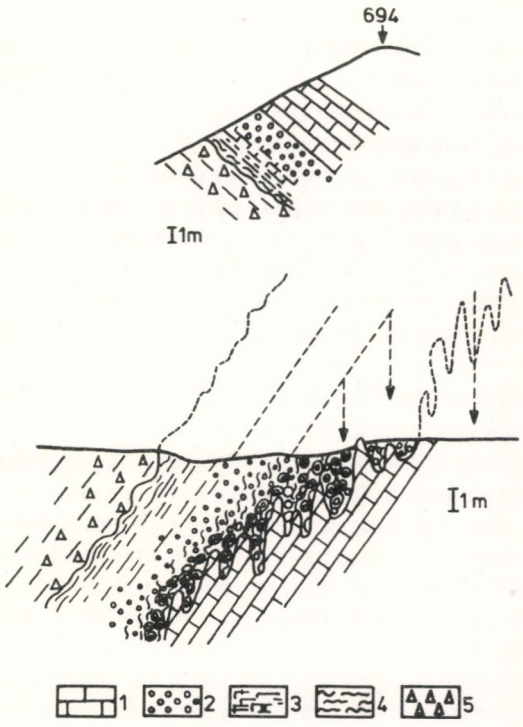


Figure 55. Secondary karstification around an overturned bauxite deposit near Cadarcet, Ariège, France, according to Combes (1969)

(1) limestone, Triassic; (2) bauxite; (3) bauxitic clay; (4) lamellar clay; (5) dolostone, Jurassic

surface to form secondary deposits are few and far between. The bauxite debris is invariably embedded in a more or less copious clayey matrix.

Only seven such deposit groups are known in all, whose aggregate tonnage is just 2% of world total karst bauxite tonnage. The bauxite tonnage stated for the Salento-type deposits in Appendix 1 refers to the total deposit mass (that is, to both the bauxite debris and the clayey matrix). The relative abundance of the two components may vary rather strongly from one deposit to the next.

Over and above the seven groups referred to, several small accumulations of bauxite have been reported in the literature; the tonnage of these, however, is so insignificant that they may be left out of our comparative-quantitative evaluation without any qualms.

Of the seven deposit groups mentioned, five are in the Mediterranean zone (Salento Peninsula, Italy; Óbarok–Újbarok and Nagyvázsony–Öcs, Hungary; Jajce–Baraći, Yugoslavia, and Zonguldak, Turkey). The two others are in the East Asian bauxite belt, in Vietnam. The writer has named the type after the deposits on the Salento Peninsula, which have the most pronounced depositological features of all European deposits. With regard to the tonnage, however, the two deposit groups in Vietnam are the most important.

Bauxite lenses of varying size on the surface of a Turonian limestone dot an area of more than 200 km^2 in the Terra d'Otranto area of the Salento Peninsula. Each lens is several hundred metres in diameter; their thickness is 5 to 8 m typically, 14 m as a maximum. The bedrock surface is weakly karsted in places and moderately karsted elsewhere; the deposit top is entirely smooth.

The deposits consist of a crumbling, earthy, red, bauxite-like clay, with hard bauxite pebbles of 0·5 to 3·0 cm diameter embedded in it. The pebbles consist of a remarkably high-grade, low-silica, ooidic–pisoidic bauxite. The lenses are exposed on the surface over a considerable area, but drilling has also traced them under Miocene, Pliocene, and Quaternary beds. The immediate cover is most often a clay, carbonaceous clay, or lignite, more seldom a limestone or sand. The cover succession is of lacustrine or brackish facies.

According to Cotecchia and Dell'Anna (1959), the bauxite pebbles derive from eroded bauxite deposits which were probably contemporaneous with the well-known Upper Cretaceous bauxite deposits of Apulia (those at Spinazzola and San Giovanni Rotondo). No primary deposit of this character has been found so far in the deposits' vicinity. The small size, hardness and rounding of the bauxite pebbles implies transportation over some distance (5 to 10 km). It is conceivable that the bauxite-like clay in which the pebbles are embedded corresponds to the softer, more clayey parts of the one-time bauxite deposits, elutriated and totally argillised in the course of transportation. The definitive settling of these issues is not yet possible, in the absence of sufficiently detailed mineralogical and petrographic studies.

The redeposited bauxite lenses near Zonguldak in Turkey are highly similar depositologically to the above-described ones. The bedrock is a Lower Carboniferous limestone; the cover is a Cenomanian sandstone. Also here, the matrix is a soft red clay of earthy consistency, which has lumps of hard high-grade bauxite embedded in it. No primary deposits have so far been discovered near the deposits, so that the source of the bauxite remains unknown.

In the Socialist Republic of Vietnam, in the provinces Lang Son, Ha Giang and Cao Bang, a reworked bauxite detritus was accumulated in depressions of Permo-Carboniferous and Upper Permian limestones, in the proximity of some primary, Palaeozoic deposits which are of Mediterranean type. Very hard, angular or hardly worn, the bauxite detritus ranges in size from a few millimetres to several cubic metres. It is embedded in a soft reddish clayey matrix, presumably an erosion product of Palaeozoic, Triassic and Jurassic strata; the matrix may include some elutriated clayey bauxite, too. The insoluble residue of the eroded limestones must also have enriched the clay fraction.

There are in places lumps of a very hard bauxitic iron ore, and more seldom, of volcanics and siliceous rocks accompanying the bauxite. The abundance ratio of bauxite detritus to matrix is highly variable (from 10 to 80% of the total mass). Recent geological study has shown the share of bauxite detritus to be 30 to 40% on average in the deposits at Dong Dang and Ma Meo of the Lang Son deposit group (Komlóssy 1976). The farther down the valleys the bauxite flowed, the smaller are the lumps in the matrix: their abundance—the potential recovery—declines in step. On the valley floors, the deposits pass gradually with no sharp limit into a red or yellowish clay.

Both a report by Chebotarev and Buy-Fu-Miy (1961) and the recent geological study (Komlóssy 1976) state the deposits to be 100 to 4000 m long, 50 to 300 m wide and 5 to 10 m thick as a rule. The bedrock surface is intensely karsted, and the deposit may be up to 50 m thick in the karst depressions.

According to personal observations of the writer there is no sharp limit in a number of places between the reworked deposits and the primary ones. The reason for this is that, here, the prime process was a simple removal by dissolution of the carbonate cover from above the deposits, rather than a mechanical erosion. Subsequently, dissolution transferred its action to the bedrock, with the result that the bauxite got gradually broken up and the fine red clay was washed into the fractures. Subsequently, the bauxite detritus, mixed with more and more clay, started to creep down the steeper slopes. The outlines of these creep forms can be recognised even today on the valley floors. The dough-like flows of reworked deposits rising above the valley bottoms and floors are quite conspicuous.

In addition to creep, sheet wash also acted on the slopes for brief spells, during heavy rains. In any case, transport distance was not much more than a few hundred metres; it certainly did not exceed one kilometre in the case of most reworked deposits. The process took place in the Neogene and may have continued in places during Quaternary. Most of the deposits are exposed: even on the valley floors there is only a thin cover of a few metres at most. Surface weathering has greatly affected the original mineralogical composition of the bauxite, resulting in the partial or total oxidation of its originally reduced geochemical facies.

The Baraći group of deposits near Jajce in Yugoslavia is transitional between the Salento and the Mediterranean types. It is underlain by a Senonian Rudist limestone and overlain by Miocene freshwater limestone, sandstone and conglomerate. In the entire bauxite region of Dalmatia–Bosnia, this is the youngest cover formation: this in itself heightens the plausibility of reworking. The material making up the deposits is not as heterogeneous as in the foregoing deposits. Reworking is implied merely by the conglomerate structure of the bauxite and by the ubiquity of detritic quartz grains. According to Trubelja and Vasiljević (1971), these deposits were redeposited from the nearby Senonian bauxites of Jajce–Bešpely. The present writer believes that some of the redeposition took place in the Eocene, but some of it went on into the Oligocene and early Miocene, when bauxitisation had already ceased. This in the writer's opinion is the reason for the transitional nature of these deposits.

The study of redeposition is especially fruitful in a Hungarian context, where the full gamut from reworked deposits to scattered bauxite pebbles and red clays of bauxitic origin can be observed. The most important reworked deposits are those in the vicinities of Nagyegyháza-puszta, Vázsonypuszta, Óbarok and Újbarok, to the southeast of Tatabánya town. These lenticular deposits ranging from 50 to 150 m in length and from 2 to 10 m in thickness fill out karst depressions in an Upper Triassic dolomite. They are overlain as a rule by Oligocene clays, sands and sandstones, but are covered in places by Quaternary clay only. Jaskó (1957a) stated the lenses to be made up of lumps of a harder, high grade bauxite embedded in a softer, more clayey bauxite. Jaskó could identify some worn Nummulites in the reworked bauxite, which had found their way in from the original Eocene cover, during the redeposition of the primary deposit. Both the erosion and redeposition presumably took place in the Lower

Oligocene. The writer is of the opinion that here, too, redeposition involved quite a short transport distance (a few hundred metres, one kilometre at most), because no significant amounts of alien matter got admixed to the bauxite, although its grade did deteriorate as compared with the neighbouring primary deposits.

There are small deposits due to similarly short-range redeposition all over the Transdanubian bauxite region, especially in the Sümeg, Nyirád, Szőc, Halimba, Iszkaszentgyörgy and Gánt areas, in the immediate vicinity of the primary deposits. To the south they are replaced by a zone predominantly of red clays which include bauxite pebbles in larger or smaller numbers. This zone extends from Zalahaláp through Monostorapáti and Tótvázsony to Vörösberény.

Redeposited bauxite pebbles are most abundant near Nagyvázsony. According to Jaskò (1957b), depressions in the top of an Upper Triassic dolomite are filled there with reworked bauxitic clay in a length of 200 to 400 m and in a thickness of 20 to 30 m. The deposit is covered by a few metres of Pliocene freshwater limestone. The soft red bauxite-like and bauxitic clay of the deposit contains hard rounded bauxite pebbles of walnut to apple size, disposed so as to form gravel beds alternating with pebble-less layers. The pebbles, whose size and abundance increase towards the edges of the deposit, are of high-grade, iron-rich bauxite; they do not as a rule constitute more than 10% of the material (30 to 40% as a maximum). Jaskó states that the clayey matrix itself is the elutriation product of an earlier, primary bauxite. The rounding, smaller size and lower percentage abundance of the pebbles indicate a longer transportation than in the above group, which gave alien clayey materials a better chance to get admixed to the original bauxite.

In the Vörösberény and Szentkirályszabadja areas, hard rounded bauxite pebbles are accumulated in rags of a Pliocene red clay found on the surface of a Middle Triassic dolomite. Bertalan (1953) estimated their abundance to be up to 60% in places. Even the Pliocene limestone overlying the clay in a thickness of a few metres contains some bauxite pebbles of 1 to 20 cm size. Samples collected and examined by the present writer have revealed that the pebbles consist of very high grade and mostly iron-rich bauxite. He believes the red clay to be most probably also of partly bauxitic origin, the red tint of the limestone being also due in places to elutriated bauxitic material.

The sites most remote from the known bauxite regions where hard bauxite pebbles embedded in red clay were found in the Gömör County karst region in Northeastern Hungary. Bárdossy (1968a) found them to consist also of iron-rich bauxite. They may be attributed to the redeposition of deposits similar to the extent Upper Cretaceous ones of nearby Slovakia. The writer holds the red clay to be of bauxite origin in part, although it has been entirely argillised, with admixtures of quartz sand and pebbles of quartz and limestone. Redeposition presumably took place in the Neogene, possibly continuing into the Quaternary.

In the writer's opinion, the reason why Salento-type deposits are so rare is that bauxite in the course of erosion falls apart and gets argillised quite rapidly. Only if the erosion process acted upon very hard and massive types of bauxite was there any chance for the accumulation of secondary deposits of considerable size. This was the situation in the Vietnamese bauxite region: this is why the most important deposits of the Salento type occur in that country.

3.7. TULSK-TYPE DEPOSITS

This type is represented by a single group of deposits totalling at most one million tons of bauxite, a mere 0·017% of world total karst bauxite tonnage. Bushinsky (1970) nevertheless distinguished them as an independent depositological type of karst bauxites, by virtue of their peculiar deposit features, unique geochemistry (including an absolute lack of titanium), and mineralogical constitution. The writer concurs with that opinion by adopting them as a separate bauxite type in his own system.

The deposits of the group are in the Moscow Basin, in the Tulsk and Ryazansk areas. The most important deposit is the one at Senurovsk. Trubina (1958) states that a friable bauxitic–clayey deposit of earthy appearance, of 10 to 20, though seldom up to 100 cm thickness overlies the surface of a Lower Carboniferous limestone. In the karst depressions of the bedrock, this deposit may swell up to a thickness of 8 m, taking forms of irregular outline (fig. 56). Horizontally, these ore bodies measure just a few metres as a rule; only exceptionally do they attain 100 to 150 m. An essential depositological feature is that they invariably occur at a small depth (3 to 30 m below the surface); they disappear wherever the cover is thicker than that. The immediate cover is an ochre-yellow Lower Carboniferous clay containing a few scattered gibbsitic nests and interbeddings; it is overlain by a grey pyritic clay with thin stringers of coal, followed by some coal measures of a thickness of 10 to 60 m.

In the opinion of Trubina (1958), this bauxite is a decomposition product of some pyrites; they formed after a substantial reduction of thickness cover due to the erosion of the Lower Carboniferous strata. The sulphuric acid generated by the pyrites'

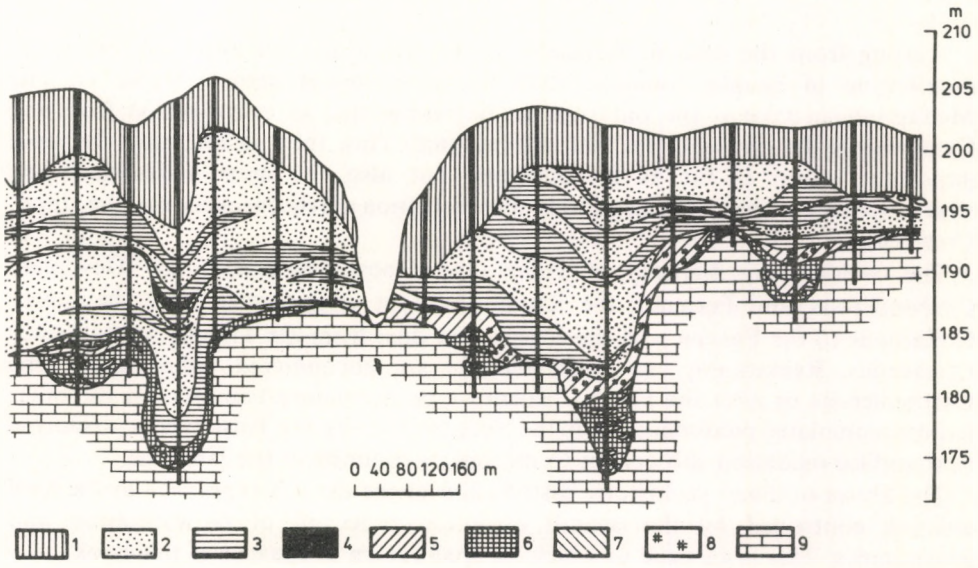


Figure 56. A Tulsk-type deposit near Senurovsk, Tulsk Department, USSR. After Trubina (1958)

(1) soil and loam, Quaternary; (2) sand, Lower Carboniferous; (3) grey clay, Lower Carboniferous; (4) coal, Lower Carboniferous; (5) Ochroid clay, Lower Carboniferous; (6) bauxite, bauxitic clay; (7) greyish-green clay, Lower Carboniferous; (8) bauxitic nests in clay; (9) bedrock limestone, Lower Carboniferous

decomposition gave rise to aluminium sulphate solutions, which, seeping downward, precipitated right above the carbonate bedrock. The writer deems this bauxite to be hardly older than Pliocene.

Another small deposit to be placed into this group is Portilla de Luna in Northwestern Spain, described as a Devonian bauxite by Font-Altaba and Closas (1960). In 1968, the writer had the occasion to make some detailed observations in this region in the company of Professor Fontboté. These observations and the chemical and mineralogical analysis of the samples taken, however, have revealed this deposit to be the same as at Tulsk, the product of a secondary bauxitisation, which took place in the Pliocene and possibly in the Quaternary rather than in the Devonian (Bárdossy and Fontboté 1977).

The site exhibits nests and veins of 10 to 20 cm size of white gibbsite and halloysite at and near the contact of an Eifelian clay shale with a Givetian limestone. These bodies are most abundant next to the contact in the structurally strongly deformed, almost upright layers. The nests are almost entirely devoid of titanium, just as the Tulsk deposits are. It would be unjustified to speak of an uninterrupted deposit, as the bodies are scattered in the shale and do not entirely replace the parent rock even where they are most abundant. This, then, is just a rudimentary form of this type of deposit, which was the reason why the writer has left them out of his comprehensive table of bauxite deposits (Appendix 1).

3.8. GEOHISTORIC DISTRIBUTION OF DEPOSIT TYPES

Starting from the data in Appendix 1, the writer has calculated a percentage distribution of bauxite tonnages over the geohistorical periods (table 7). The Mediterranean type is the only one to be represented in every period from the Cambrian to the present, the Silurian excepted. This, then, is not only the most important group as far as is concerned but also the most widespread one stratigraphically. The two peak periods of its formation are the Neogene and the Upper Cretaceous.

The owerhelming majority of Timan-type deposits formed in the Devonian, Carboniferous, and Permian; the Kazakhstan-type ones formed from the Lower Cretaceous to the Eocene, and the Ariège-type ones from the Triassic to the Lower Cretaceous. Remarkably, no such deposits are encountered either before the Carboniferous or after the Eocene. Salento type, secondary bauxites, on the other hand, accumulated predominantly in the Neogene. Finally, the Tulsk type, a product of near-surface oxidation subsequent to erosion, is confined to the Neogene.

The above-outlined geohistoric distribution of the deposit types was in the final analysis controlled by the overall evolution processes of sedimentation and geotectonics. This is an issue to which we shall return at the end of this book.

Table 7. Age distribution of deposit types with respect to tonnage.

Geological period	Mediterranean		Timan		Kazakhstan		Ariège		Salento		Tulsk	
	bauxite per cent	number of deposit groups	bauxite per cent	number of deposit groups	bauxite per cent	number of deposit groups	bauxite per cent	number of deposit groups	bauxite per cent	number of deposit groups	bauxite per cent	number of deposit groups
Quaternary	1	7	–	–	–	–	–	–	–	–	–	–
Miocene–Pliocene	32·8	11	–	–	–	–	–	–	85	3	100	1
Oligocene	0·1	3	–	–	–	–	–	–	11	3	–	–
Paleocene–Eocene	12	24	1	1	52	8	–	–	–	–	–	–
Upper Cretaceous	26	34	1	1	34	4	–	–	–	–	–	–
Lower Cretaceous	5	15	–	–	11	7	8	2	4	1	–	–
Jurassic	7	17	<1	1	–	–	1	1	–	–	–	–
Triassic	2	13	1	1	–	–	91	1	–	–	–	–
Permian	5	9	15	1	–	–	–	–	–	–	–	–
Carboniferous	1	2	71	8	3	1	–	–	–	–	–	–
Devonian	7	7	10	1	–	–	–	–	–	–	–	–
Silurian	–	–	–	–	–	–	–	–	–	–	–	–
Ordovician	0·1	1	–	–	–	–	–	–	–	–	–	–
Cambrian	1	2	–	–	–	–	–	–	–	–	–	–
Altogether	100	145	100	14	100	20	100	4	100	7	100	1

3.9. BAUXITE FORMATIONS

Bauxite deposits have a tendency to occur in typical bauxitic formations. These are sedimentary successions, formed under similar conditions of climate, geotectonics and sedimentary geology and which contain bauxite deposits in one or several horizons. A comparison of bauxitic formations by deposit types and bauxite belts permitted the writer to distinguish the following bauxitic formation types.

3.9.1. CARBONATIC BAUXITIC FORMATIONS

The Mediterranean-type bauxite deposits tend to be associated with formations several hundred metres thick, made up predominantly (sometimes almost exclusively) of carbonate rocks, mostly of neritic, more seldom of open-sea (deeper-water), even more seldom of brackish or freshwater facies. Next to the carbonate rocks, the most abundant sediments are marls and clay marls, but these are much less thick. In the immediate cover of the bauxite deposits, there are also thin beds of carbonaceous clay, coal or lignite. What these formations lack almost entirely are clastic deposits of silicate minerals. This is the justification for the term 'carbonatic formation'.

Bauxite tends to occur in several horizons, each of which represents a local emersion or an unconformity. Starting from the stratigraphic data and from the currently accepted international absolute time scale (van Eysinga 1978), the writer has calculated the duration in millions of years of the hiatus between the immediate bedrock and the immediate cover of the deposits. The stratigraphic ages of the latter are stated in Appendix 1. The actual hiatus was, of course, shorter, as some of the bedrock was

eroded after the emersion. The estimation of this eroded portion being uncertain, it is preferable to work with the length of the entire (apparent) hiatus. The 145 Mediterranean-type deposit groups have yielded the following values.

Length of hiatus in million years	Number of deposit groups
1– 5	22
5– 10	21
10– 20	26
20– 30	16
30– 40	7
40– 60	7
60– 80	9
80–100	5
100–130	4
130–150	4
150–170	13
170–200	4
200–220	1
220–240	1
500–550	1
uncertain	4

More than half of all deposit groups, then, are found in hiatuses less than 30 million years in age. These deposit groups are all intercalated in the purest carbonatic formations, in those were non-carbonatic lithofacies are the most subordinate. They are most numerous in the bauxite regions of the Dinarids, Hellenids, Urals, the Turkestanian Ranges and the Caribbean Islands. The number of deposit groups in hiatuses 100 to 200 million years long is also quite large. In the bauxitic formations including these, clayey–marly lithofacies are more abundant than in the preceding group. For an example, let us cite the 11 Mediterranean-type deposit groups of the Transdanubian Hills, Hungary.

The Sohodol-Cîmpeni, Rumania, deposit intercalated in a hiatus of singular length (500 to 550 million years) cannot, in the depositological respect, be considered pure Mediterranean type; it exhibits transitional features towards the Kazakhstan type, most of whose deposit groups are found in hiatuses as long as this.

All the Ariège-type deposits are found in carbonatic formations, in hiatuses of extreme brevity:

Lika (Croatia)	1 to 5	million years
Ariège	5 to 40	million years
Saint Paul de Fenouillet	5 to 10	million years
Skopelos Island	1 to 5	million years

All four Ariège-type deposit groups are in the Mediterranean belt, scattered among deposit groups of Mediterranean type, in the same sort of carbonatic formations. This is why, from now on, we shall lump these two types together in our analysis of carbonatic formations.

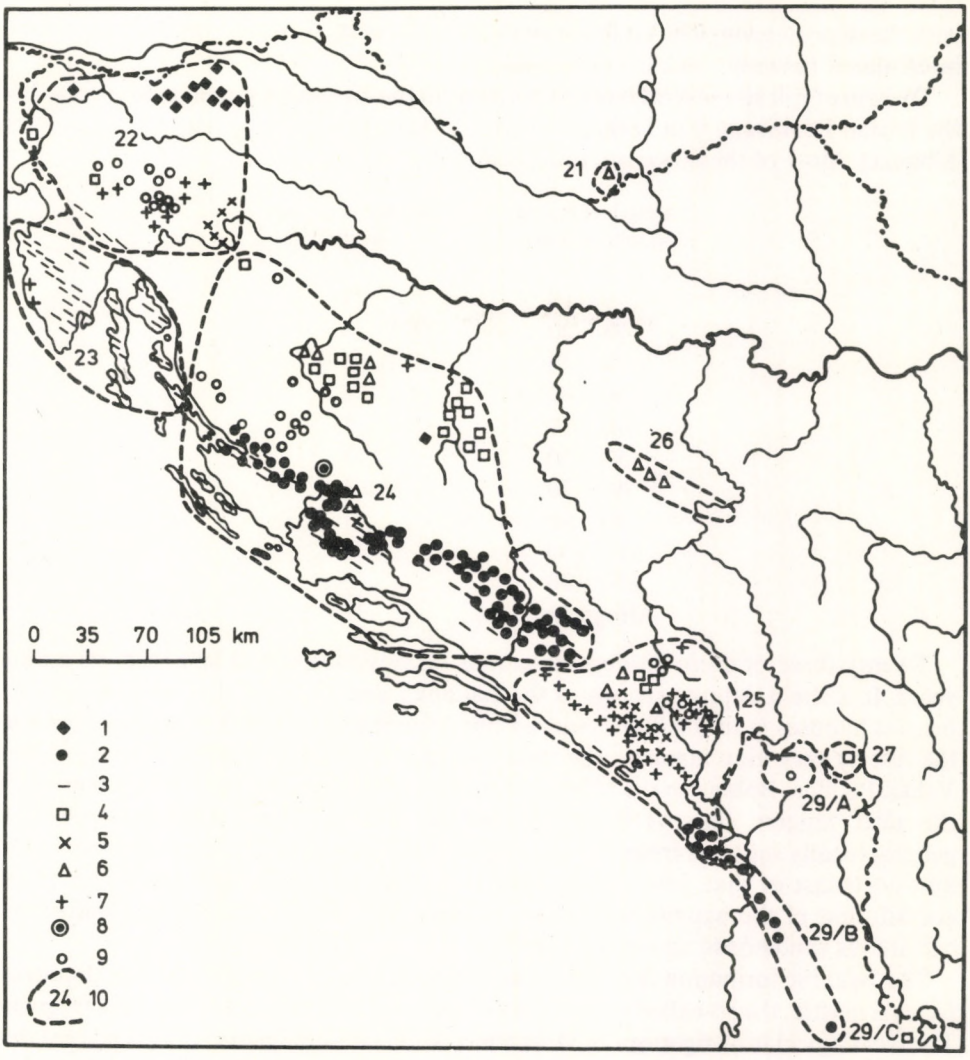


Figure 57. Geographic distribution and age of the bauxite deposit groups of Yugoslavia and Albania. After Grubić (1975), complemented by the writer

(1) Oligocene deposits; (2) Middle Eocene deposits; (3) Lower Eocene and Paleocene deposits; (4) Senonian and Turonian deposits; (5) Cenomanian deposits; (6) Lower Cretaceous deposits; (7) Jurassic deposits; (8) Upper Triassic deposits; (9) Middle Triassic deposits; (10) serial number and outline of bauxite group

Let us now examine in somewhat greater detail the *pure carbonatic bauxitic formations*, taking for an example the bauxite deposit groups of the Dinaric range, including those of Slovenia, Croatia, Bosnia, the Herzegovina, Crna Gora (Montenegro) and Albania. Their geographic positions and ages are shown in figure 57; their stratigraphic columns, in colour figure 1 (separate, at the back of the book). In most deposit groups, the bauxite formation is made up of limestones and dolostones, in places with associated limestone conglomerates and breccias. It is only in connection

with the deposit groups of Triassic age that one finds layers of volcanic tuff and tuffite intercalated between the carbonate rocks.

There are typically several bauxite horizons in the formation (seven in Slovenia, 2 on the Istrian Peninsula, 9 in Dalmatia and the Herzegovina, 5 in Crna Gora, and 2 in Albania). Most of these occupy brief hiatuses:

Length of hiatus in million years	Number of deposit groups
1– 5	5
5– 10	4
10– 20	8
20– 30	6
30– 40	2
40– 50	1
50– 60	1
60– 70	1
70– 80	1
110–130	1
170–190	3
Altogether	33

Twenty-three of these 33 deposit groups occupy hiatuses of less than 30 million years. It is the 5 deposit groups of the Ladinian and Carnian that are found in the briefest hiatuses (1 to 5 million years). Four deposit groups are distinguished from all the rest by very long hiatuses: the three Northern Slovenian groups (Savinja Valley, Velika Planina, Bohinj) in the 170 to 190 million years class and the Vlasenica group in the next longest class (110 to 130 million years). These deposit groups are geographically separate from the main zone of bauxite deposits; they lie to the north and northeast of that zone. They differ from the pure carbonatic ones also in the constitution of the bauxite formation, in that there are also marls, clay marls, and calcareous sandstones above the bauxites.

The bauxitic formation is rather homogeneous also as to faciology. The dominant facies is neritic; the so-called carbonate plateau or platform facies is the most frequent. The faciological investigations of D'Argenio, Radoičić and Sgrosso (1971) proved that in the Triassic and Jurassic successions the facies called 'rim complex' and 'back-reef complex' to be the most frequent, true reef facies being somewhat rarer. Deeper-water and bathyal facies are also rare or, alternatively, bauxites are missing from the regions where they occur.

The strata immediately underlying the bauxite horizons are invariably neritic; no regressive succession can be observed anywhere. The emersion, then, must have been quite rapid, its traces wiped out by subsequent erosion. The deposition of bauxite was succeeded in most deposit groups by a very rapid transgression. In most places, even the immediate cover of the bauxite is of marine, littoral facies. Only in the deposit groups of Paleocene and Eocene age was there a somewhat slower ingression. For example, the Paleocene bauxite horizon of Kozina and Istria is overlain by Paleocene 'Cosina' limestone of freshwater or lagoon facies, 40 to 60 m thick. Only above the Cosina formation, in the Lower and Middle Eocene, are there any neritic

limestones. The Cosina layers are found also in the cover of the bauxite deposits of Obrovac, Novigrad and Drniš, but they are less thick there. In the central part of the Istrian Peninsula, around Karojba and Pazin, the immediate cover is a carbonaceous clay of swamp facies, 1 to 3 m thick, overlain by 5 to 10 m of brackish water gastropod limestone followed by a neritic limestone. This transgressive succession is Lower and Middle Eocene (Bušer and Lukács 1972).

The carbonatic bauxitic formation of the region ends in the Middle and Upper Eocene, partly with a general emersion, and partly with the deposition of sandy-clay 'flysch-like' deposits. The flysch cannot by any means be ascribed to the bauxitic formation any more. In the Imotski and Mostar areas, the flysch succession begins just a few metres above the Middle Eocene bauxite horizon. The writer regards this fundamental change of lithology as the upper limit of the bauxitic formation.

Also in the Hellenid Ranges, bauxite occurs in a pure carbonatic formation. The time span corresponding to the formation and the number of bauxite horizons are different in the different principal structural units. In the so-called *Parnassos Zone*, the bauxitic formation with its three bauxite horizons encompasses the interval from the Upper Triassic to the Middle Paleocene; it ends with the incipient deposition of a Paleocene flysch succession. Each of the three bauxite horizons occupies a very short hiatus of 1 to 10 million years, terminated by an extremely rapid transgression. Made up almost exclusively of limestones and dolostones in a total thickness of 1500 to 1700 m, this succession is one of the finest examples of the pure carbonatic bauxitic formation.

In the adjacent *Subpelagonian Zone*, the carbonatic bauxitic formation extends from the Triassic to the middle of the Malm, with a single bauxite horizon in the Lower Malm. The lithologic succession is made up of Triassic dolostone, dolomitic Liassic limestone, and pure Dogger and Malm limestone. The hiatus including the bauxite horizon spans some 5 to 10 million years. The emersion was followed by a rapid transgression; even the immediate cover of the bauxite is a neritic limestone. From the Upper Malm on, the nature of sedimentation changes profoundly; clay shales, radiolarites and ophiolites were formed in the Upper Malm and Lower Cretaceous. These are overlain by an Upper Cretaceous flyschoid complex and by typical flysch, Paleocene and Eocene age. It is owing to this change in sedimentation that the two uppermost bauxite horizons of the neighbouring Parnassos zone are lacking here.

The best instance of another basic type of the carbonatic bauxitic formation is in the Transdanubian Hills of Hungary. It spans the stratigraphic interval from the Middle Triassic to the Upper Eocene. The twelve deposit groups of this region occupy a hiatus much longer than the previous ones: 70 to 90 million years at Alsópere and Tés, 130 to 140 million years in the Halimba Basin, and 140 to 170 years in the remaining ten deposit groups. The stratigraphic columns of the deposit groups also showing the facies of the sediments are presented in colour figure 2 (separate, at the back of the book).

There are, however, minor bauxite deposits and indications also in much shorter hiatuses; the ore bodies of the small open pits of Sümeg–Szőlőhegy and Surgoth-tag, e.g. are underlain by Campanian Hippurites limestone and overlain by Middle Eocene strata (Vadász 1946; Bárdossy 1961c). Prospect drilling in recent years has discovered bauxite lenses in the same stratigraphic position also in the Csabpuszta and Kolontár areas near Nyirád (Károly *et al* 1970). The hiatus spans 20 to 25 million years there.

In the Ajka and Padragkút areas, prospect drilling discovered bauxites inserted between Albian Requienia limestone and Senonian terrestrial deposits (Szantner and Szabó 1970): there are some minor traces of bauxite even within the Requienia limestone. The aggregate tonnage of these small ore bodies and indications is so small that the writer has preferred not to include them in Appendix 1; their stratigraphy, however, is presented in colour figure 2. The aggregate tonnage of these deposits does not attain even one percent of the total bauxite tonnage of the Transdanubian Hills; they do not therefore invalidate the general statement concerning the length of hiatus.

The bauxitic formation of the Transdanubian Hills can be characterised as follows after the stratigraphic studies of Fülöp (1964, 1966), Konda (1970), and Végh–Neubrandt (1969).

The formation starts with a transgression in the Lower Triassic. At the beginning, the principal lithofacies are sandstones, dolomitic sandstones and dolomitic marls; from the middle of the Campilian on, limestones and dolostones become dominant. Only the Carnian and Rhaetic stages include layers of marl. The Carnian–Norian 'Hauptdolomite' and the Rhaetic 'Dachsteinkalk' are typical neritic 'carbonate plateau' facies, formed in a sea hardly deeper than reef-facies depth (for the Hauptdolomite) or just slightly deeper than that (for the Dachsteinkalk) (oral communication by Végh–Neubrandt 1974). Although in most deposit groups it is these Upper Triassic rocks that immediately underlie the bauxite, palaeogeographic considerations make it highly probable that marine sedimentation continued into the Jurassic in the areas where the deposit groups occur. Fülöp (1971) stated the bottom of the Liassic to be of predominantly littoral facies; it passes during a gradual subsidence into the neritic and subneritic facies, and finally into the bathyal facies in the Bathonian and Callovian stages. These were followed by a slow incipient regression, which by the Tithonian had again produced some neritic facies.

In the Southwestern Bakony Hills, rags of Liassic strata spared by erosion were identified under the bauxite (Károly *et al* 1970). Palaeographic considerations permit to infer the Jurassic to have extended also into these regions, although it was largely eroded afterwards. At Iszkaszentgyörgy and in the Vértes Hills, there is no trace of Jurassic rocks under the bauxite, but the proximity of Jurassic strata and their facies render their prior existence in these areas rather likely.

Fülöp (1964) states the Berriasian stage in the Southwestern Bakony to be made up of sublittoral limestone. Pure carbonate sedimentation continued in the Valanginian stage with a crinoidal limestone, to be replaced by a calcareous marl with chert nodes in the Hauterivian and by radiolarian marls in the Barremian. In the Northern Bakony, pure carbonate sedimentation continued up to the end of the Hauterivian, whereas the Barremian produced glauconitic sandstones and cephalopod limestones, all neritic. The northeastern shore of the Lower Cretaceous sea was at the height of Zirc according to Fülöp; towards the southwest, deeper and deeper basin facies followed. Fülöp's map shows that the zone of the bauxite deposits, which had emerged late in the Jurassic or early in the Cretaceous, was not reached by this sea. Fülöp deems the bauxite to be a facies heteropic to the Lower Cretaceous marine strata deposited in the neighbourhood.

Early in the Aptian, a very brief emersion followed, after which most of the Transdanubian Hills were again inundated by the sea, up to the town of Tata (fig. 58).

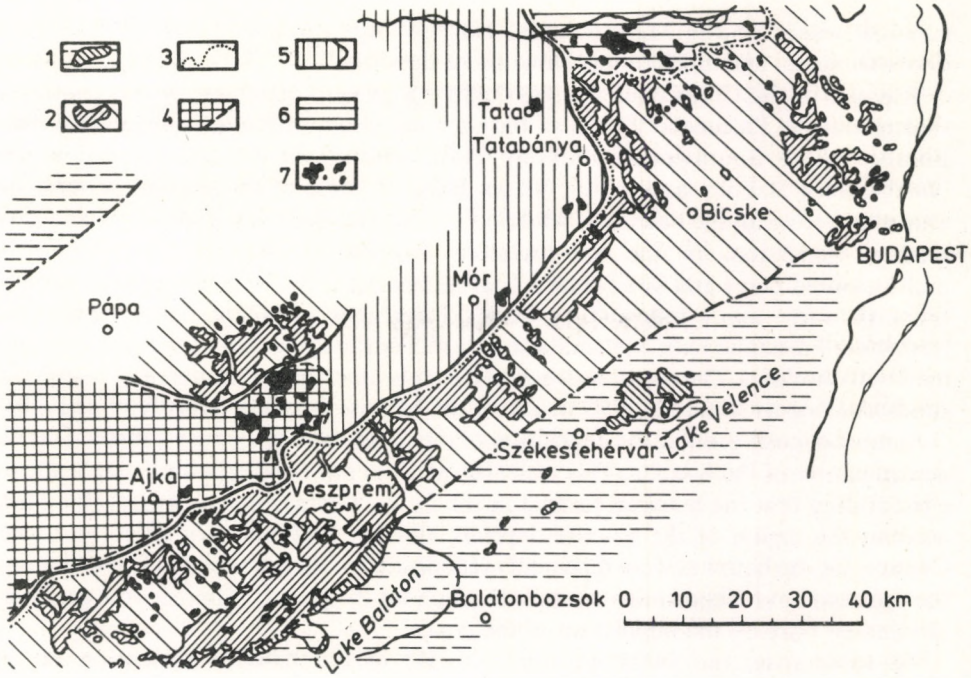


Figure 58. Distribution of Valanginian–Aptian strata in the Transdanubian Hills, Hungary, according to Fülöp (1964)

(1) Paleozoic, on surface and buried; (2) Triassic, on surface and buried; (3) boundary of Berriasian strata; (4) extent and boundary of Valanginian–Hauterivian–Barremian–Aptian strata; (5) extent and boundary of Aptian crinoid limestone; (6) Valanginian–Barremian strata on surface; (7) Valanginian–Aptian strata on surface

After another brief emersion, clays and marls with Munieria, of lagoon and brackish water facies, came to be deposited about the end of the Aptian, with intercalated euhaline deposits (Fülöp 1961). Remarkably enough, not even traces of bauxite have so far been discovered in these two brief hiatuses. In the Alsópere and Tés areas, on the other hand, where the Munieria clays and marls transgress beyond the border of the older Cretaceous strata, there are bauxite deposits below them and overlying Rhaetic and Lower Liassic Dachstein-type limestone.

The Albian is characterised by neritic limestones with rudists and orbitolinids and by sandy limestones. The Cenomanian is represented by glauconitic marls followed by marls with Globotruncana and Turrilites (Fülöp 1961). In the Turonian, the entire Transdanubian Hills area was laid dry. In the Northern Bakony and in the Vértes Hills, this terrestrial episode presumably continued into the Senonian. In the Southwestern Bakony Hills, on the other hand, another transgression began in the Santonian, with a freshwater dolostone and limestone conglomerate and multicoloured clay marl at its base. This is the immediate cover of the bauxite deposit in the Halimba Basin. It is overlain by a brackish water succession with coal beds, followed by a Gryphea marl. According to Fülöp (1961), this latter was deposited in a muddy shallow sea, rather close inshore. The Hippurites limestones of the Campanian already indicate a clear

euhaline water. In the Maestrichtian, marls of sublittoral and neritic facies with Globotruncana and Inoceramus came to be deposited.

The end of the Cretaceous was marked by a general emersion, which continued throughout the Paleocene. The next marine inundation took place during the Lower Eocene in the Southwestern Bakony Hills, and in the Middle Eocene in the Northeastern Bakony and in the Vértes Hills. In most of the deposit groups, the immediate cover of the bauxite is Middle Eocene. The Eocene inundation was a slow gradual ingression. In the Southwestern Bakony, the bauxite is overlain by multicoloured clays and swamp-facies carbonaceous clays, overlain in their turn by brackish water marls followed by neritic limestones. At Gánt, some banks of a freshwater limestone with Melania also occur. They are overlain by a brackish water marl with molluscs. The neritic succession of sandy limestone, clay marl and calcareous sandstone begins some 80 to 90 m above the bauxite deposits (Károly *et al* 1970).

In most deposit groups, the deposition of the cover took place very shortly after the accumulation of the bauxite. This observation does not conflict with Fülöp's above-cited finding that the bauxites formed in the Lower Cretaceous, because, taking into account the region of the bauxite deposits also left high and dry after the Lower Cretaceous, the bauxites were repeatedly redeposited in all probability. The writer is of the opinion that redepositions went on throughout the period of immersion, and were terminated only by the deposition of the cover.

To summarise, the following may be said of the bauxitic formation of the Transdanubian Hills.

(i) The pure limestone and dolostone lithofacies are slightly less dominant than in the pure carbonatic type bauxitic formations described previously. Marls, clays, sandy limestones, calcareous sandstones and carbonaceous clays are more abundant. The abundance of clayey and sandy material increases especially from the Barremian period on.

(ii) Although the ages of accumulation of the bauxites in their present sites are different, their overwhelming majority nevertheless falls within a single very long hiatus, in which they form a single bauxite horizon. It is only in the Csabpuszta and Kolontár areas in the Southwestern Bakony Hills that there are two bauxite horizons, one above another. The 3 to 9 bauxite horizons of the pure carbonic-type bauxitic formation are unknown here.

(iii) The percentage of clayey bauxites is higher than in the pure carbonic deposits.

(iv) The transgression above the bauxite is slower than in the pure carbonatic formations: the first euhaline deposits are found at a height typically of 10 to 50 m, at a maximum of 90 m above the deposit top.

A carbonaceous, clayey swamp facies plays a greater role in the immediate cover.

Within the carbonatic bauxitic formations, then, a *pure carbonate type* is to be distinguished from a *contaminated type* (i.e. contaminated with clay and sand). These two types differ not only as to lithofacies but also in their entire geological evolution.

The bauxitic formation which contains the Nagyharsány deposit group in the Villány Hills, Southern Hungary, resembles the pure carbonatic formation of the Dinaric ranges. According to Fülöp (1966, 1971), a neritic limestone and dolostone of the Anisian and Ladinian is overlain there unconformably by a Middle Liassic littoral

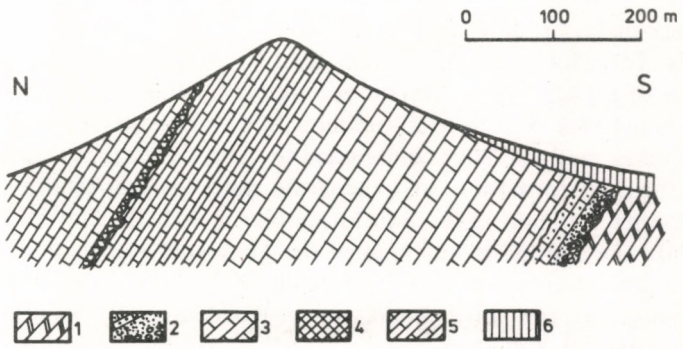


Figure 59. Geological profile of Harsány Hill, Hungary, according to Fülöp (1971)
(1) dolostone and limestone, Anisian and Ladinian; (2) limestone breccia overlain by limestone, Pliensbachian; (3) limestone, Callovian–Tithonian; (4) bauxite, bauxitic clay; (5) limestone, Barremian; (6) loess and loam, Quaternary

limestone breccia, followed by a few metres of neritic limestone. Above another hiatus, there are 400 to 450 m of neritic Callovian, Oxfordian, Kimmeridgian and Lower Tithonian limestones. It was during the short emersion of 17 to 22 million years following these latter strata that the bauxite came to be accumulated. A swift transgression followed in the Lower Barremian, with 2 to 5 m of clayey nodular limestone at its base. It is overlain by a thick-banked pure limestone of massive texture. A pure carbonatic, neritic sedimentation continued up to the middle of the Albian. The succession is illustrated by a geological profile of Harsány Hill (fig. 59). The movements of the Austrian orogenic phase interrupted the limestone facies and replaced it by marls and aleurolites at the middle of the Albian: in a short while, the entire Villány Hills area was laid dry. On Harsány Hill, the Cretaceous was eroded down to the Lower Aptian limestones. The remaining Cretaceous succession is 170 to 200 m thick.

This, then, is a pure carbonatic bauxitic formation of marine facies, with three brief emersions. Only the uppermost one is associated with any bauxite, although in the Dinaric Ranges bauxite did also form in the other two. The reason for this difference is not well understood. According to Fülöp (1966), the marine transgression invaded the area from the south and southeast: it is there that strata similar in age and facies to those at Villány should be looked for. There is unfortunately a cover of thick young basin sediments there, making direct geological observation impossible.

A pure carbonatic bauxitic formation is found at Pădurea Craiului in the Bihar Mountains, Rumania. The Neocomian bauxite deposits there are almost contemporaneous with those at Nagyharsány, and the hiatus in which they are found is also very brief (10 to 20 million years). Fülöp (1966) states that the geological constitution of the two regions is strikingly similar.

The deposit groups of Unterlaussa and Dreistätten in Austria and of Mojtin and Markušovce–Drienovec in Slovakia are embedded in bauxitic formations very similar to those of the Transdanubian Hills. The hiatus in which the bauxite deposits proper are inserted is also of similar length. These deposit groups have a single known bauxite horizon. An exceedingly long hiatus, a single bauxite horizon and a clayey-sandy contamination of the overlying beds in the three Northern Slovenian deposit groups

and in the deposit group of Vlasenica, as described above, relate the above to the deposits of the Transdanubian Hills rather than to any other region.

It is possible to explain the geographic zoning of the bauxitic formations in simple palaeogeographic terms. In the writer's opinion, however, the geotectonic evolution of the respective areas may also have contributed to the circumstance that the two types of carbonatic bauxitic formation are in separate zones geographically.

Deposits inserted into pure carbonatic formations include the Devonian ones of the Northern Urals, the Carboniferous ones of the Turkestan Range, and the Neogene ones of the Caribbean area. The bauxite deposits of the Languedoc and Provence in Southern France, on the other hand, are embedded in carbonatic formations contaminated with clay and sand.

3.9.2. BAUXITE FORMATIONS INCLUDING COAL SEAMS, CLAYS AND CARBONATES (THE 'COAL–CLAY–CARBONATE FORMATION')

It is the Timan-type deposits that are found in formations of this type, with the exception of the deposit groups at Naurzumsk in the USSR and on the southeastern rim of the Coastal Plain in the US, but including the deposits of Missouri State. In the 12 deposit groups involved, the length of the stratigraphic gap occupied by the bauxitic complex is as follows.

Length of hiatus, in million years	Number of deposit groups
10– 20	1
20– 30	2
40– 50	1
110–170	3
130–180	1
170–180	1
120–220	1
200–270	2

The deposits that occupy hiatuses shorter than 30 million years are those of the Timan Mountains and of the Taimyr Peninsula. The deposit groups of China and of Missouri State, on the other hand, are found in long hiatuses.

The bauxitic formation in the Timan bauxite region starts with Middle Devonian claystones, sandstones and aleurolites. These contain intercalations of limestone whose number increases upward. In the Upper Devonian, limestones and dolostones alternate. The immediate bedrock of the bauxite is a marine limestone with fairly frequent clayey intercalations. The greatest thickness of the Devonian succession is 800 m (Demina 1971). Its deposition was followed by a brief immersion and by the deposition of the Lower Carboniferous bauxitic complex, in angular unconformity on the eroded bedrock surface.

The bauxite is overlain by 5 to 30 m of Visean coal measures, with carbonaceous clay and lenses of coal at the bottom and aleurolites and sandstones on top. The age

difference between the beds underlying and immediately overlying the bauxite is 20 to 30 million years. The coal measures pass with continuity into a succession, 2 to 50 m thick, of multicoloured clays and aleurites in alternation. These in turn pass into 100 to 200 m of Upper Visean and Namurian neritic limestone and dolomitic limestone, as with numerous clayey intercalations.

In the area of the deposits, the strongly eroded surface of these carbonate rocks is overlain by Quaternary sands and clays. In the adjacent great synclines, the marine carbonate deposits of the Middle and Upper Carboniferous and of the Lower Permian are also preserved. These too contain many clayey interbeddings. In the Upper Permian, the carbonate rocks are replaced by increasingly abundant clay, aleurite and sandstone, marking the end of the bauxitic formation. Gulyanitsky (1969) states that the total thickness of the Carboniferous is 600 to 800 m; that of the Permian, 300 to 1500. The Upper Carboniferous and Permian were probably also present in the bauxite area, but were eroded from their more uplifted positions.

In summary, the bauxitic formation has the following salient features:

(a) the carbonate rocks making up the bedrock and the high cover are impure, rich in clayey interbeddings;

(b) the bauxitic complex proper also includes aleuritic and sandy sediments in addition to bauxite;

(c) the immediate cover includes coal seams more often than not. The cover succession indicates a slow ingression, with the first marine facies turning up at a height of 50 to 120 m above the bauxite deposits.

Bauxite is also embedded in a bauxitic formation of similar age and constitution at the Taimyr Peninsula (Budnikov 1971; Mikhailov *et al* 1973).

In the deposit groups of Southern China, the bauxitic formation emerges from a Lower Cambrian shale, which is overlain by Middle and Upper Cambrian dolomitic limestones. The Ordovician is represented by marine limestones, with clayey intercalations in its upper part. The Silurian mainly consists of marine shales; the Denovian starts with sandstones, but its upper part consists of dolomitic limestones (Wang Chung-hsing 1965). The deposition of these beds was followed by an episode of emersion and erosion. The bauxite overlies limestones mostly of Cambrian, less often of Ordovician age.

In Kweichow Province, there are two separate bauxite horizons (fig. 60): the lower horizon is overlain by Middle Carboniferous carbonaceous shales with stringers of coal, followed by pure marine limestones in a process of slow ingression. The region then emerged once more, and the upper bauxite horizon, deposited in the hiatus,

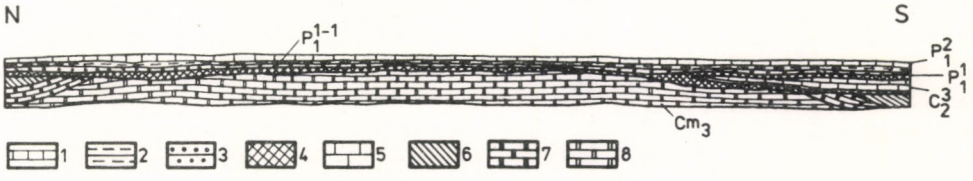


Figure 60. Synoptic profile of the Kweichow bauxite region, China. After Wang Chung-hsing (1965)
(1) limestone, Permian; (2) claystone, bone; (3) sandstone; (4) bauxite, bauxitic clay; (5) limestone, Carboniferous; (6) shale, Silurian; (7) limestone with claystone intercalations, Ordovician; (8) dolomitic limestone, Upper Cambrian

subsequently came to be overlain by Lower Permian coal measures, including beds of shale, sandstone and coal. Another slow ingression produced some more marine limestone. The Upper Permian gave rise to another set of coal measures, overlain by another succession of marine limestones. Finally, the Triassic succession, the closing term of the bauxitic formation in this area, is an alternation of marine limestones, dolostones and aleurolites.

In Central and Northeast China—in Honan, Shantung and Liaoning provinces—the bauxitic formation is made up somewhat differently. It starts with Ordovician limestones and dolostones. Some karst bauxite deposits then formed during a subsequent emersion. They are overlain by coal measures of Middle and Upper Carboniferous age, including an alternation of coal beds, shales, aleurolites, sandstones and lacustrine limestones, as well as seven horizons of Tikhvin-type bauxite. The uppermost of these is at a height of 200 to 500 m above the karst bauxite horizon. Finally, a slow ingression also produced marine limestones. The bauxitic formation ends in the Upper Permian.

In this area, then, the basic features of the Timan region are seen to recur: (a) coal measures in the immediate cover, (b) a subsequent slow ingression; (c) clays, aleurites and sandstones intercalated in the carbonate rocks of both the near and the high cover. On the other hand, the presence of Tikhvin-type bauxite deposits in the cover is a unique local feature.

The bauxitic formation of the Missouri State bauxite region stands closest to the type just described, as shown clearly by the geological profile in figure 45.

3.9.3. CLAYEY-SANDY BAUXITIC FORMATIONS

Formations of this type are connected with the Kazakhstan-type deposits, with exception of the Missouri State region. The bauxitic formations containing the Timan-type deposit groups of Naurzumsk in the USSR and those of the southeastern rim of the Coastal Plain, US, are also of this type. In the 21 deposit groups involved, the length of the stratigraphic gap occupied by the bauxite complex is as follows.

Length of hiatus, in million years	Number of deposit groups
10– 20	1
140–150	1
200–220	2
200–300	1
220–280	4
260–300	2
290–330	1
300–550	1
400–450	1
420–470	2
500–550	1
550–600	1
600–700	3

With the single exception of the Southeastern Coastal Plain, all deposit groups occupy exceedingly long hiatuses: it is in this type of bauxitic formation that the hiatuses are longest.

The bauxitic complex is in the majority of the deposit groups closer in geological age to the cover than to the bedrock; the time gap towards the latter may be several hundred million years long. The bauxite, then, formed in an entirely different phase of the structural and sedimentary evolution of the given area than the bedrock succession. This is why, in the writer's opinion, the bedrock succession should not be included at all with the bauxitic formation which, accordingly, should be taken to start with the bauxite deposit proper. The only exception is the Southeastern Coastal Plain, whose bauxitic formation is to be characterised separately below.

An idea of how the formation is constituted can be gained from the example of the deposit groups of the Kazakhstan, which may be considered the most typical. The bauxitic complex, regarded as the bottom term of the bauxitic formation here, is described in Section 3.4. It is overlain after a brief hiatus by a cover that is of continental facies on the margins of the basin. In the Arkalyk group of deposits, for example, it includes a river-laid sandstone overlain by limnic–lacustrine multicoloured clay, clayey aleurite and sandy clay. At its top, there is an aleurite with thin stringers of gravel. It marks the end of the bauxitic formation; the area emerged for good in the Miocene. The cover succession is 150 to 170 m thick; the thickness of the entire bauxitic formation is 170 to 220 m.

In the deeper-lying central and western parts of the adjacent Turgay Plateau, on the other hand, the bauxitic complex is overlain by a marine Albian–Cenomanian cover (fig. 61, separate, in the back of the book). The transgression was also a gradual one here, introduced by a swamp facies of carbonaceous clay and lignite, and followed up by brackish water clay and aleurite. It was only thereafter that a neritic glauconitic sand and sandstone came to be deposited. The higher cover is an alternation of marine clays, clay marls, diatomaceous clays, aleurites and sandstones. The bauxitic formation of this region ends with an early Miocene emersion. The total thickness of the cover succession is 200 to 250 m in the central part of the basin, and 130 to 180 m in the deposit groups of the Western Turgay.

In the Alapayevsk group of deposits on the eastern slopes of the Central Urals, the bauxitic formation consists of a transgressive succession of similar constitution; the thickness ranges from 20 to 150 m.

In the deposit groups at Chadobetsk and along the Yenisei River, the deposits are covered by just a few metres of Quaternary clay and scree (figs 44, 46 and 47). The bauxitic formation is confined here to the deposit proper. In addition to bauxite, it includes clayey, silty and sandy types of sediment. The bedrock is Proterozoic and Cambrian, whereas spore and pollen analysis has given an Albian–Cenomanian age for the deposit groups at Chadobetsk and along the Angara River, and a Paleocene one for the remaining three deposit groups along the Yenisei. It follows from the geological history of the region that the areas in question have remained high and dry ever since the deposition of the bauxitic complex.

The situation is similar in the Valley and Ridge Province of Southeastern US, where a Cambrian and Ordovician bedrock and a Paleocene and Lower Eocene bauxitic

complex are covered by just a few metres of Quaternary clay, sand and gravel (figs 42 and 43).

In the deposit groups on the southeastern rim of the Coastal Plain, US, the bauxitic formation is made up differently. It begins with an Upper Cretaceous neritic clay overlain by 30 to 50 m of marine micaceous sand and friable sandstone which include lenses of kaolin-type clay. They are overlain in turn by the Paleocene clayey limestone that is the immediate bedrock of the bauxite. Twenty five to 40 m thick on the southwestern margin of the area, it gradually thins out and then disappears towards the northeast. The bauxitic complex, of a typical thickness of 15 to 30 m and 40 m as a maximum, overlies the Paleocene. It is in turn overlain after a brief hiatus by a thin layer of swamp-facies lignite or by river-laid cross-bedded sandstone, followed by Lower Eocene neritic glauconitic sand and by lamellar clay. The formation is concluded by a Middle Eocene sand bed: the region emerged for good after its deposition. The succession is well illustrated by the geological profiles in figures 39 and 40. The bauxitic formation is 150 to 200 m thick. The principal difference to the previously described formations is that, here, the bauxitic complex is found near the middle rather than at the base of the formation. Accordingly, the hiatus is also substantially shorter.

The principal features, then, of the clayey-sandy bauxitic formations are as follows.

(a) The lithology is overwhelmingly clayey, silty and sandy: carbonate rocks are totally lacking, with the exception of a single deposit group,

(b) With a single exception, the hiatus between the bedrock and the bauxitic complex is of very great length. The bauxitic complex is therefore to be regarded as the first term of the bauxitic formation,

(c) In a majority of deposit groups, the entire bauxitic formation is continental; of the 21 groups, only in six is the cover succession of a transgressive nature,

(d) The thickness of the bauxitic formation nowhere exceeds 300 m, and is, indeed, less than 100 m in several deposit groups.

3.9.4. SUMMARY

The basic features of sedimentation are different in the three types of bauxitic formation. The carbonatic formation is dominated by neritic sediments of carbonate type. The terrigenous contribution is greater in the coal–clay–carbonate formation, and dominant in the clayey-sandy formation. This change is accompanied by a decrease in the share of the marine facies and by an increase in the share of the continental ones. Total thickness of formation ranges from 1000 to 3000 m in a majority of the carbonatic formations; it is somewhat less in the coal–clay–carbonate formations, and a mere 100 to 300 metres in the clayey-sandy formations.

Within the carbonatic formation, there is an extreme sub-type, the pure carbonate formation, whereas the sub-type contaminated with clay and sand is already transitional towards the coal–clay–carbonate type bauxitic formation.

The position of the bauxite complex within the formation and its lithology are different in the different formations. There are several bauxite horizons in the carbonatic type, especially in the pure carbonate sub-type. The coal–clay–carbonate

formation includes karst bauxite deposits in a single horizon within the formation. The single bauxite horizon of the clayey-sandy formation is, with a single exception, at the base of the formation.

It is the bauxite deposits of the pure carbonate formation that are purest (i.e. contain the greatest tonnages of high-grade bauxite). The greater the share of terrigenous components in the bauxitic formation, the higher is their percentage also in the bauxite complex, and the poorer is the average bauxite grade. This is reflected by the distribution of total karst bauxite tonnage over the formations: 75% in carbonatic formations, 15% in coal–clay–carbonate formations, and 8% in clayey-sandy formations.

Although the writer's subdivision of karst bauxite deposits into six depositological types was based exclusively on their constitution and depositological features, the formations which incorporate the deposits can also be grouped according to depositological type: the Mediterranean and Ariège types tend to occur in carbonatic formations, the Timan-type ones in coal–clay–carbonate formations, and the Kazakhstan type in clayey–sandy formations. The exceptions to this rule comprise just three minor deposit groups. Owing to their reworked nature, the Salento-type deposits cannot be assigned plausibly enough to any one of the bauxitic formations. The same holds for the Tulsk type of deposits which, though of special position and origin, is insignificant as to tonnage.

Depositological type is, in the final reckoning, a function of the bauxitic formation. The constitution of a deposit (bauxite complex) then, is not accidental, but an organic component in the geological evolution of the bauxitic formation of which it forms part.

Remarkably, those successions which include deposits of the Tikhvin-type exhibit no features that would permit the definition of a distinct bauxitic formation. In other words, no separate bauxitic formation typical of the Tikhvin-type deposits can be distinguished. Some Tikhvin-type deposits are found in coal–clay–carbonate formations, others in clayey–sandy formations. It is not, then, accidental that Tikhvin-type deposits tend to lie in the proximity of Timan and Kazakhstan-type deposits, seeing that they tend to occur in the same formations. Indeed, as revealed by the Chinese deposit groups, they may occur together in one and the same succession.

4. LITHOLOGY

4.1. CLASSIFICATION AND FREQUENCY OF BAUXITE TYPES

4.1.1. PREVIOUS CLASSIFICATIONS

The great variety of bauxite types has for long been an incentive towards a systematisation with a view to providing a better overview. Prior to the 1940s, no adequate method was available for the accurate determination of the bauxites' mineralogical compositon: hence, the classifications of that period used to be based on *chemical composition.*

Harrassowitz (1926) found that he could best characterise bauxite types in terms of the molecular ratio of silica to alumina, i.e. of an index which he called the *Ki* coefficient:

$$Ki = \frac{\text{Weight\% } SiO_2}{\text{Weight\% } Al_2O_3} \times 1{\cdot}7 .$$

The factor 1·7 serves for the transformation of weight into molecular percentages. Harrassowitz introduced the term *siallites* to denote all rocks composed of pure aluminium silicates (of kaolinite), the *Ki* parameter of which is a round 2·0. For the other extreme of the range, which he presumed to consist of pure aluminium hydrates, he introduced the term *allites.* (The *Ki* coefficient of a pure allite is zero.) He defined *allitic siallites* and *siallitic allites,* as intermediate members between the two extremes. The total quantity of Al and Fe hydrates must be less than 10% in a siallite and more than 90% in an allite. Allitic siallites and siallitic allites cover the ranges from 10 to 50% and from 50 to 90%, respectively. Harrassowitz reserved the term 'bauxite' for allites of monohydratic constitution; he called the trihydratic ones '*laterites*'. This distinction did not gain any broad acceptance. Only his terms allite and siallite survived—in Soviet literature above all—, with gradually changing contents.

Malyavkin (1937) used the term allite to denote rocks in which free aluminium hydrates predominate. Siallites as construed by him were clayey rocks related to the allites, possibly containing minor amounts of aluminium hydrates. Malyavkin inverted Harrassowitz's original ratio to Al_2O_3/SiO_2, and used direct weight percentages. He called his ratio the *module* or *silica module,* and drew the limit between allites and siallites at a module of 1·0. The use of the module is widespread throughout Soviet literature and still much in use today.

Vishnyakov (1940) retained the allite–siallite limit set at the module 1·0, but called 'bauxites' all rocks having a module higher than 2·6. In his terminology, then, the term allite was confined to rocks higher in silica, and more clayey, than bauxite. It was in the same sense that Goretsky (1960) used the terms allite and bauxite. He defined bauxite

as a rock containing more than 28% Al_2O_3 and having a module higher than 2·6, and drew the limit between allites and siallites at a module of 0·87. Krivtsov (1968, 1969), on the other hand, called 'siallites' the rocks in the module range from 0·85 to 1·0, 'allites' those in the 1·0 to 2·1 range, and 'bauxites' those with modules higher than 2·1. He used the generic term 'bauxitic rock' for all allites and siallites. According to Krivtsov, it is the bauxitic rocks that constitute the transition between bauxites and clayey rocks.

Bushinsky (1971) proposed to use these terms in the original Harrassowitzian sense, but with the following module limits: siallite, less than 1·0; allitic siallite, between 1·0 and 2·0; siallitic allite, between 2·0 and 2·3; and allite, above 2·3. He wished to confine the term bauxite to industrial-grade ores of aluminium.

In the relevant Western literature, J. De Lapparent's (1930) chemically based classification came into widespread use in the 1930s. It was a twofold one, based on silica contents on the one hand and on iron contents on the other. He distinguished them according to the silica ratio

'bauxites mégalumineuses':	$Al_2O_3/SiO_2 = > 20$
'bauxites alumineuses':	$Al_2O_3/SiO_2 = 10–20$
'bauxites siliceuses':	$Al_2O_3/SiO_2 = 4–10$
'bauxites mégasiliceuses':	$Al_2O_3/SiO_2 = < 4$,

and, according to the iron content,

'bauxites microferriques':	$Fe_2O_3 < 10\%$
'bauxites ferriques':	$Fe_2O_3 = 10–25$
'bauxites mégaferriques':	$Fe_2O_3 = > 25$.

De Lapparent's above classification, based upon a study of French bauxite types, was fairly adequate at the time.

Pedro (1966) proposed a geochemical classification to be based on a distinction of 'free' aluminium (aluminium hydrates) and 'bound' aluminium (aluminium silicates). On this basis, he developed the following system:

bauxite:	free Al_2O_3: 70 to 100%;	$Ki = 0·0–0·6$
clayey bauxite:	free Al_2O_3: 50 to 70%;	$Ki = 0·6–1·0$
bauxitic clay:	free Al_2O_3: 20 to 50%;	$Ki = 1·0–1·6$
clay:	free Al_2O_3: <20%;	$Ki = 1·6–2·0$

The novel instruments of mineralogical analysis introduced in the 1940s and 1950—the thermal and X-ray methods especially—permitted the development of classifications based on *quantitative mineralogy*, which express the composition of a bauxite in the most adequate manner. The first classification of this type (a triangular diagram, fig. 62) was due to Konta (1958). The three apices of the triangle were marked in the three most important mineral groups of bauxite: (a) aluminium hydroxides, (b) haematite and iron hydroxides, and (c) clay minerals. True, Konta's geometric principle of classification permits an easy overwiew, but does not deal with the transitional rock types occupying the central part of the diagram.

Valeton's (1972) classification is also of a geometric nature (fig. 63). It distinguishes a greater number of rock types, but does not dealt with the central field of transitional composition, either.

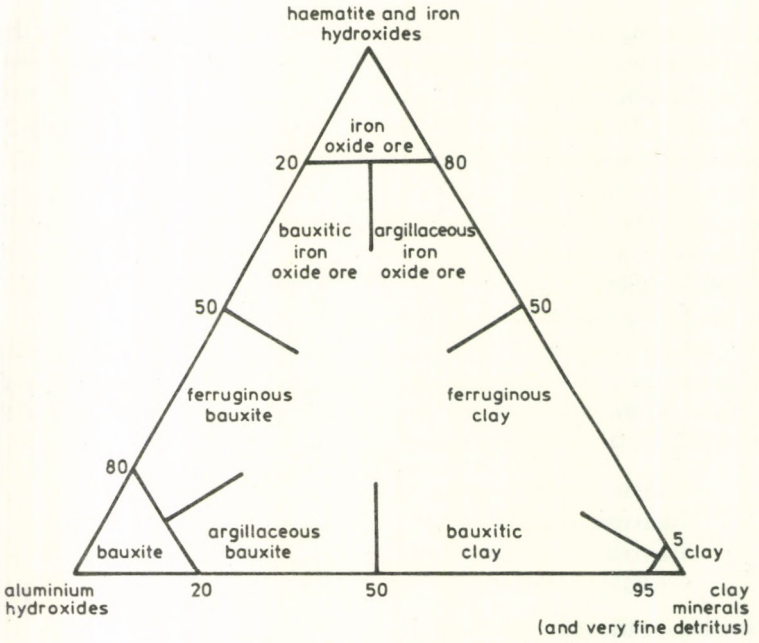


Figure 62. Petrographic classification of bauxites according to Konta (1958)

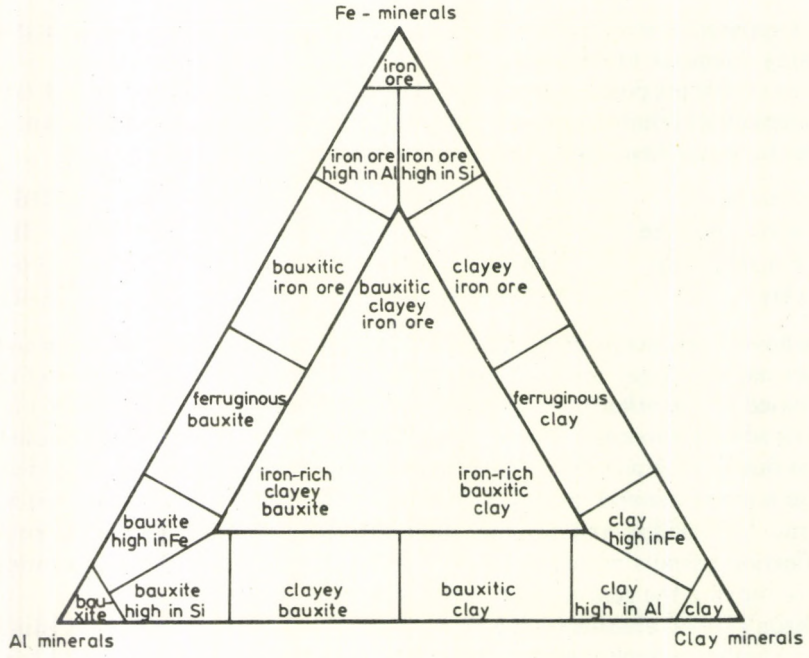


Figure 63. Petrographic classification of bauxites according to Valeton (1972)

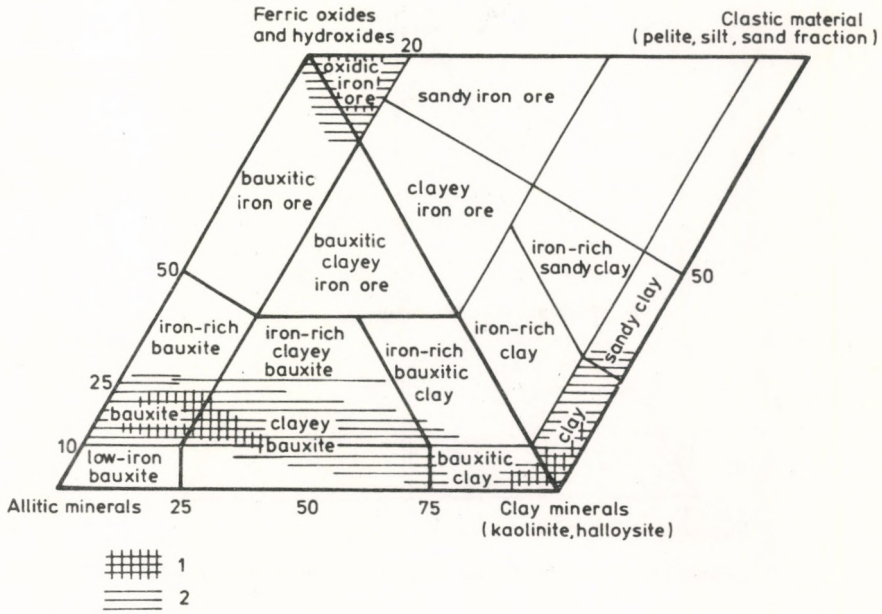


Figure 64. Petrographic classification of the bauxites and their relation to the argillaceous rocks and sedimentary iron ores, according to Bárdossy (1964)

(1) most frequent rock facies in the nature; (2) frequent rock facies

Both classifications have the prime drawback of paying no attention to the abundances of bauxite types as observed in nature. They are logical but theoretical systems which have no direct bearing on Nature. Sensitive to these issues, the present writer proposed a classification in 1964 based on the estimated relative abundances of individual bauxite types. He retained the three main mineral groups of Konta's and Valeton's triangular diagrams but, in subdividing the central field, he made no attempt to outline regular, geometric forms (fig. 64). A second triangle attached to the first one served to illustrate relationships with clayey and clastic rock types.

4.1.2. A NEW PROPOSAL OF CLASSIFICATION

In writing this book, the writer strove to take into account the natural abundances of bauxite types to an even greater extent than before. He separated each deposit group into bauxite types and estimated their abundances. An overview of the data thus obtained made it clear that the above diagram could be further simplified. The definitive triangle diagram distinguished only seven rock types as against the nine of the previous classification (fig. 65), as a result of a contraction of the low-abundance clayey iron-rich types.

The writer has placed also anatase and rutile into the apex of the aluminium hydroxide and oxide minerals. These titanium minerals are closely associated with the aluminium minerals genetically, and their total abundance seldom exceeds 3%. This

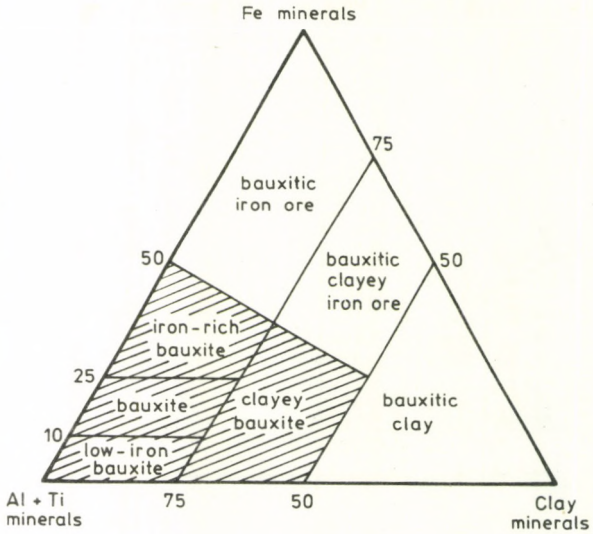


Figure 65. Proposed novel petrographic classification of bauxites

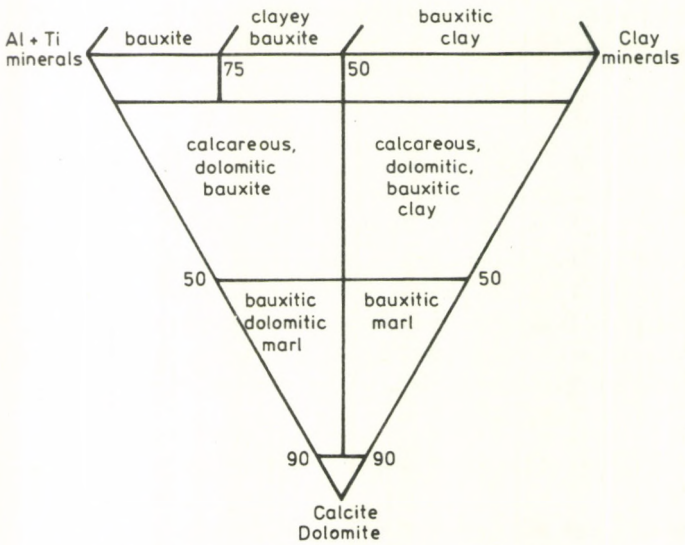


Figure 66. Proposed novel classification of carbonatic bauxite types

was a change for the better inasmuch as it permitted the plotting of the mineralogical compositions of most bauxites directly, with no prior recomputation, into the triangle diagram. Pyrite, marcasite and siderite, which may play a significant role in some reduced bauxite types, were also incorporated with the iron minerals in addition to the iron oxides and hydroxides. There remained the difficulty of how to deal with the iron-rich septechlorites and leptochlorites, which are clay minerals but also important

carriers of iron. As they usually contain iron and silica in about equal amounts, the writer has felt the best solution to be to add one-half of their percentage weight to the iron minerals, and the other half to the clay minerals.

It is the shaded part of the triangle that contains the bauxites in the broader sense. Within it, four bauxite types are distinguished. The three bauxite types of the unshaded area are subsumed under the generic term '*bauxitic rocks*': these represent the transitions towards the clays and the sedimentary iron ores.

It is not, of course, also possible to represent all other minerals of bauxite in this single triangle diagram. Auxiliary diagrams must be adduced to deal with bauxite types of special composition. Bauxites containing finely dispersed calcite and/or dolomite are shown in figure 66. The upper edge of their triangle diagram fits the bottom edge of the basic triangle. Accordingly, 'bauxite' corresponds here to '*calcareous* or *dolomitic bauxite*'; 'bauxitic clay' corresponds to '*calcareous* or *dolomitic bauxitic clay*'. If the content of calcite or dolomite exceeds 50%, the terms *bauxitic limestone*, *bauxitic dolostone* or *bauxitic marl* are to be used. If the calcite/dolomite content exceeds 90%, the adjective 'bauxitic' becomes superfluous.

In some deposit groups, there are limestone or dolostone breccias having a bauxite or a clayey bauxite matrix. If their bauxite content exceeds 10%, the writer uses the term *bauxitic limestone* (or *dolostone*) *breccia* for them.

In places, bauxite passes into coal, carbonaceous clay or lignite. If the organics content is in the range from 10 to 50%, one may speak of *carbonaceous (lignitic) bauxite* or a *carbonaceous (lignitic) bauxitic clay*. If the organics content exceeds 50%, the names *bauxitic lignite* or *bauxitic coal* are to be used.

Bauxites containing more than 10% phosphate minerals are called *phosphoritic bauxites*. These may in places pass into phosphorite (rock phosphate) deposits. The transitional type, *bauxitic phosphorite*, contains 50 to 90% phosphate minerals.

A few bauxites contain more than 10% manganese minerals. They are called *manganiferous bauxites*. No bauxitic manganese ores have been identified so far.

Several bauxite deposits contain kaolinitic clays or iron-rich kaolinitic clays which to the naked eye are very similar to bauxite, sometimes to the point of being undistinguishable. Their colour and texture are the same as those of the surrounding bauxite. If the bauxite is unstratified, then so is this clay. It contains no bauxite minerals (gibbsite, boehmite or diaspore), but its 'accessory' minerals are the same as those of bauxite: haematite or goethite, anatase, rutile and kaolinite. It also differs from the usual run of clays in that it contains insignificant amounts of clastic material (quartz, feldspars, sericite). Périnet (1961), the first to study these clays, called them *pseudo-bauxites*. As regards mineralogical composition, they cannot be regarded as an independent rock type: they are simply clays or iron-rich clays. In view of their lithologic features and close genetic attachment to the bauxite deposits, however, the present writer nevertheless deems it necessary to distinguish them as a lithofacies, by the name '*bauxitoid clay*' (bauxite-like clay).

No other bauxitic rock types transitional into other sedimentary rocks are known. True, bauxite lumps of smaller or larger size can be found in the sandy and coarse clastic sediments overlying certain bauxite deposits. These, however, are secondary, mechanical admixtures, and cannot therefore be regarded as transitions towards (or away from) bauxite.

Although theories deriving bauxites from pyroclastics are not lacking, no bauxitic pyroclastics have so far been identified. No bauxitic silica rock is known, either, and a bauxitic evaporite seems to be a contradiction in terms.

Finally, the writer wishes to specify those terms connected with bauxite classification which he does not recommend.

Schneiderhöhn (1944) classified bauxites in the strict sense according to their principal aluminium minerals, calling diasporic bauxites *diasporities*, boehmitic bauxites *boehmitites* and gibbsitic bauxites *gibbsitites*. This subdivision did not gain acceptance, as it would only have burdened with superfluous new terms a terminology that was complicated and equivocal enough to start with. These terms could at best denote the pure bauxite types, but the principal bauxite minerals occur rather frequently in mixtures also, a fact that frustrates the comprehensive application of this terminology.

Önay (1949) proposed calling the weakly metamorphic bauxites of predominantly diasporic composition *diasporites*, and all non-metamorphic bauxite types bauxites, regardless of their aluminium minerals. The acceptance of this proposal would but complicate matters superfluously as, according to it, one and the same diasporic bauxite would have to be called a bauxite if non-metamorphic and a diasporite if weakly metamorphic. Now the decision as to exactly what bauxites should be regarded as weakly metamorphic can be taken only after a profound textural and mineralogical examination.

American literature often terms diasporic bauxites *diaspore clays*. According to Allen (1935), in the Missouri bauxite region, this term denotes bauxites containing more than 68% Al_2O_3. After visiting the deposits and performing comparative laboratory analyses on samples collected there, however, the writer regards these as simply bauxites for which no separate term is justified. The term 'clay' is misleading anyhow, as these high-purity bauxites contain very few clay minerals.

Some authors use the term 'flint clays' for low-iron bauxites and bauxitic clays of ceramic and refractory grade. According to Grim's (1967) original definition, *flint clay* is an abbreviation of 'flint-like claystone', denoting a massive, non-plastic claystone of conchoidal fracture which consists overwhelmingly of kaolinite, contains no quartz or chalcedony, and resembles flint macroscopically. This definition does not connotate any bauxite minerals or bauxitic features. Some of the bauxite types of the Missouri bauxite region are indeed called flint clays. Allen (1935) and Keller *et al* (1954), however, are of the opinion that this is largely a commercial and industrial term.

The white low-iron bauxite and bauxitic clay of the Maktesh Ramon group of deposits, Israel, is also called a flint clay if its average Fe_2O_3 content is less than 1·5% (Slatkine and Heller 1961). In Hungary, the bauxitic clay and bauxite-like clay of Pilisvörösvár near to Budapest was named a flint clay by Nemecz and Varju (1967). In those authors' view, a flint clay is a hard rock of conchoidal fracture and a very fine grain (minus-2 μm); it is white or red, consisting predominantly of kaolinite, low in aluminium minerals and haematite.

The writer has examined samples of typical flint clay from Pilisvörösvár, the Missouri deposits and Maktesh Ramon by macroscopic and microscopic methods from the mineralogical and petrographic aspect. He has found that they do not differ from other karst bauxites: they can be fitted without any difficulty into the system of

karst bauxite types as outlined above. He does not deprecate the use of the term flint clay in practice, in applied geology, in connection with ceramic or refractory grade material, or with clay deposits, but deems its use in bauxite geology inappropriate for the reasons pointed out above.

Finally, the writer would emphasise that the bauxite classification proposed by him is of a fundamentally scientific nature. In industrial practice, it would be preferable to use the term *aluminium ore* rather than the term bauxite. The limits of regarding a bauxite as ore-grade material vary from one country to the next and from one period of time to the next, as a function of evolution in refining technology and of the economic conditions of the given country.

4.1.3. ABUNDANCE OF BAUXITE TYPES

Taking as a basis the data available in literature, as well as those furnished by his field trips and laboratory studies, the writer has estimated the relative abundances of bauxite types in the individual deposit groups. The results are presented in Appendix 2. The writer wishes to emphasise in this context that these are just estimates, which at best express rough relative abundances of the individual rock types.

Subsequently, the writer has calculated the tonnage-weighted abundances of these rock types over total world karst bauxite tonnage, with the following results.

Bauxite in the strict sense	39%
Low-iron bauxite	4
Iron-rich bauxite	9
Clayey bauxite	20
Bauxites in the broad sense together	72% total
Bauxitic iron ore	2
Clayey bauxitic iron ore	3
Bauxitic clay	14
Bauxite-like clay	6
Other special types of bauxite taken together	3
Altogether	100%.

An overwhelming majority of all karst bauxites, then, can be plotted within the triangular diagram shown as figure 65. Most bauxites do not lie in the vicinity of the aluminium-mineral apex of the triangle, as represented by Konta and Valeton, but at some distance towards the iron-minerals apex. In nature, a bauxite type with 10 to 25% iron minerals is the most widespread: hence, this is the true typical bauxite.

Of the transitions between bauxite and the other rocks, those towards the clays are the most frequent. Remarkably enough, the transition is accompanied by a monotonic decrease of abundance: clayey bauxite, fairly copious at 20%, is followed by bauxitic clay with 14% and bauxite-like clay with a mere 6%. A similar decrease is observed towards the sedimentary iron ores: the abundance of iron-rich bauxite is 9%, that of bauxitic iron ore only 2%. As regards tonnage, nearly three-quarters of all bauxite deposits are made up of bauxite in the broader sense of the term; the bauxitic rocks of transitional nature only make up the remaining quarter of total tonnage. This, of

course, is but a world-wide average, from which the composition of this or that deposit group may differ rather widely.

A survey of the data in Appendix 2 reveals the deposit groups within each deposit type to be similar as to lithologic composition.

The share of *bauxite s.s.* is greatest in the Mediterranean-type deposits and, within that type, in those deposit groups embedded in pure carbonate formations. Its percentage is 60% in the great deposit groups of the Caribbean, 30 to 70% in the Hellenids, 30 to 75% in the Dinarids, and a round 90% on Rennel Island. (The latter is the highest known percentage within a deposit group anywhere.) The deposit groups in the less pure carbonate formations contain less bauxite s.s. (For example, 25 to 55% in Southern France; 10 to 50% in the Transdanubian Hills.)

In all other deposit types, the percentage of bauxite s.s. is much less: 10 to 25% in the Ariège type, 3 to 20% in the Timan type, 0 to 20% in the Kazakhstan type. Two deposit groups of this last type (western rim of the Coastal Plain and Missouri) contain no bauxite s.s. at all; nor do the Tulsk-type deposits. Its abundance is 5 to 20% in the Salento type, with the exception of the Jajce–Baraći deposit group, where it attains 50%. (This deposit group was stated in the depositological chapter to be transitional between the Salento and the Mediterranean types. Its strikingly high content of bauxite s.s. confirms this assessment.)

The abundance of *low-iron bauxite* is much less than that of bauxite s.s.: a mere 4% on a worl-wide karst-bauxite average. Its percentage is greatest in the Tulsk type, where it makes up some 20% of the deposits, unaccompanied by either bauxite s.s. or iron-rich bauxite. Next in line is the Timan type. All 14 deposit groups of this latter contain 5 to 20% iron-poor bauxite. It is noteworthy that the Southern Timan group of deposits contains much more low-iron bauxite (20%) than bauxite s.s. (5%). The Mediterranean type of deposits is characterised in this respect by a comparatively great variety. Of the 145 deposit groups, 57 are not known to contain any low-iron bauxite: most of the others contain 2 to 15%. There are, on the other hand, substantial percentages in some other deposit groups: the group of Nadrt, Slovenia, Yugoslavia with 40% serves as an example.

Of the Hungarian deposit groups, Nagyharsány contains 20% low-iron bauxite; Gánt and Nézsa contain 10%; the rest contain 2 to 5%, except for the deposits in the Halimba and Nagyegyháza basins, where no low-iron bauxite has been identified so far.

Low-iron bauxite is subordinate in the Ariège type: it is altogether absent from two of its deposit groups; in two others, it varies between 5 and 10%. The Kazakhstan-type deposits are similar to the Ariège-type ones in that 11 of the 20 deposit groups classed there lack low-iron bauxites altogether and the rest also contain less than 5%. The only exceptions are the two deposit groups of the Valley and Ridge Province and of Missouri State, where iron-poor bauxite attains 10 to 20%: these, however, contain relatively little bauxite s.s. (in fact, the Missouri group contains none). No low-iron bauxite is known from any one of the seven deposit groups of the Salento type.

Iron-rich bauxite is more abundant than low-iron bauxite (9% on a world-wide average). The averages of the individual deposit groups do not scatter widely about this mean; they are mostly between 5 and 20%. Of the 191 karst bauxite deposit groups, no iron-rich bauxite has been found so far at all in 38 and only in 7 deposit groups does its share exceed 20%.

Deviations between the deposit types are less than in the foregoing cases. Salento-type deposits contain 5 to 25% iron-rich bauxite. Kazakhstan-type deposits also contain rather much (10 to 20%), with the exception of the Southeastern US and of Missouri State. A majority of Mediterranean deposits contains 5 to 15%, but some are fairly rich in this type of bauxite (Grebnićka Planina, Yugoslavia, 45%; Yazd, Iran, 40%; in Greece, Euboea Island has 40%, Eleusis Mandra and Megara have 30%; Niué Island has 30%; Lang Son and Cao Bang in Vietnam, 25%).

None of the Timan-type deposits contains more than 10% iron-rich bauxite. In the deposits of the Ariège type, its average content also ranges from 5 to 10%. Finally, the Tulsk-type deposits contain no iron-rich bauxite at all.

Most of the Hungarian deposit groups contain between 5 and 10% iron-rich bauxite; the minimum (2%) is at Nagyharsány, the maximum (20%) at Nagyegyháza.

The role of *bauxitic iron ore* in karst bauxites is rather subordinate, 2% on average: that of *clayey bauxitic iron ore* is slightly greater at 3%. The two types of rock tend to occur together. They are most abundant in the deposit group of Payas-Islahye, Turkey, where the share of bauxitic iron ore is estimated to be 20% and that of clayey bauxitic iron ore at 40% (that is, the two, taken together, make up more than half of the deposit). In all other deposit groups, the role of these two rock types is much more insignificant: even taken together, they seldom exceed 30%.

There are no substantial differences in their abundance from one deposit type to the next. They are most abundant in the Kazakhstan type of deposits, being present in most deposit groups of that type. Their joint share is 15 to 30% in the deposit groups of Chadobetsk and along the Yenisei River, and 10 to 15% in the rest. They are ubiquitous in the Chinese deposit groups of the Timan type (5 to 10% bauxitic iron ore and 5 to 15% clayey bauxitic iron ore). Both rock types are absent from a majority of Mediterranean-type deposit groups. After the Payas-Islahye deposit group just cited, their share is greatest at Grebnićka Planina in Jugoslavia (5 + 5%), at Kerman (10 + 10%) and Bukan (0 + 10%) in Iran. The scanty data at the writer's disposal also permit an inference of their comparative abundance in the deposit groups of Afghanistan, in the Pamirs and at Sisophon–Battambang in Campuchea. No bauxitic iron ore or clayey bauxitic iron ore is known in any of the Hungarian deposit groups.

The average abundance of *clayey bauxite* in karst bauxites is 20%. This is the geographically most widespread bauxite type, unknown only in the deposits of the Loyalty, Fiji and Tonga island groups of the Pacific. Its abundance does not significantly differ from one deposit type to the next. The range of its variation is 10 to 30% in most deposit groups: only in 19 groups does it attain between 40 and 50%. In Hungary, it is most abundant at Alsópere (round 40%). The clayey bauxite and bauxitic clay of the Bijele Poljane deposit group of Crna Gora (Montenegro) is iron-poor: this is one of the most important occurrences of low-iron bauxitic rock types. It figures in Yugoslav literature as low-iron or '*white bauxite*': but lithologically, most of it is clayey bauxite.

The average abundance of *bauxitic clay* is 14%. This is also a geographically widespread rock type; just as clayey bauxite, it is known in every deposit group except those of the three Pacific island groups. It exhibits, however, some variability from one deposit type of the next. Its share is least in those deposit groups of the Mediterranean type connected with the pure carbonate formations, e.g. 2 to 10% in the Dinarids and

Central Greece, 10 to 20% in the Urals. In the other deposits of Mediterranean type, those connected with the impure carbonate formations, it usually attains from 10 to 30%, more seldom 40%. In Hungary, it is the deposit groups of Alsópere and Nézsa that, at 30%, are richest in bauxitic clay: the rest only contain 10 to 20%.

Some small Mediterranean-type deposits of the Caribbean are made up predominantly of bauxitic clay (Cayman Islands, 90%; Bahamas, 70%; Puerto Rico, 50%). Remarkably, the great deposit groups in the proximity of these, on Jamaica and Hispaniola, only contain 5 to 15% bauxitic clay.

The share of bauxitic clay is 10 to 20% in the Timan-type deposits, 20 to 30% in the Kazakhstan type, 30 to 40% in the Ariège, 10 to 40% in the Salento type, and 40% in the Tulsk type.

The average abundance of *bauxite-like clay* in the karst bauxites is 6%. It is absent from most of the Mediterranean-type deposit groups connected with the pure carbonate formations. In the rest, including Hungary, its share is from 3 to 10%. The Timan- and Ariège-type deposit groups contain 10 to 20%; the Kazakhstan-type ones, 10 to 30%. At 10 to 40%, it is most abundant in the Salento-type deposits, as an obvious consequence of their redeposited nature.

Calcareous and *dolomitic bauxites* are rare, amounting to 0·6% of the world total karst bauxite tonnage. Calcareous bauxite is more widespread, known as it is from 18 deposit groups. Its abundance is mostly in the 2 to 10% range. Only the Ordovician deposit group near Tselinograd, Kazakhstan, USSR, contains about 50%, and the La Boissière group in Southern France about 25%. Lower percentages (about 5%) of dolomitic bauxite accompany calcareous bauxite in the deposit groups of the Northern Urals and those around Shiraz in Iran.

Dolomitic bauxite is known from three deposit groups so far. Its average abundance is an estimated 5 to 10% at Halimba and at Nagyegyháza–Mány in Hungary. Finally, of the bauxite samples collected on Skopelos Island, Greece, the writer's own laboratory analyses have shown about 10% to be dolomitic. The depositological and genetic position of this last deposit group, however, is in need of further clarification.

Bauxitic-limestone, *-dolomite* and *-marl* are more rare than even the foregoing rock types, making up 0·1% at most of total karst bauxite tonnage. They have been identified so far in a mere 7 deposit groups, all of which also contain calcareous and/or dolomitic bauxite. In none of the deposit groups does the abundance of these three rock types exceed 10%.

Bauxitic dolomite has so far been indentified only in the Halimba Basin and on Skopelos Island. Bauxitic marl has been described from the Northern Urals only. Bauxitic limestone is known from the deposit groups of the Northern Urals, of Berdsk-Obukhovsk, of Unterlaussa (Austria), La Boissière and Tselinograd. These rock types are found most often in the top part of the deposit; they constitute a passage towards its carbonate cover where that latter overlies the bauxite conformably.

Limestone breccias with a bauxitic cement make up some 0·4% of world karst bauxite tonnage. They have been identified in a total of 11 deposit groups, all of which belong to the Mediterranean deposit type. In the deposit groups of the Southern, Central and Northern Urals, at Berdsk–Obukhovsk, in the Turkestan Range, at Bukantau and Tamdytau Ranges, they constitute an 'under-ore breccia', making up 5% at most of the total bauxitic complex. This rock type was discussed in Section 3.2.3.4. It is essentially a

pseudobreccia produced by bedrock karstification subsequent to bauxite deposition. A photo of such a pseudobreccia was presented as figure 32.

The remaining five deposit groups, on the other hand, contain true bauxitic limestone breccias. The abundance of these is greatest in the deposit groups near Dragalj, Petrovići and Velimlje of Crna Gora (Montenegro), estimated at about 20% by the writer on the occasion of his field trips there. Red bauxite, in its top horizons especially, contains close-packed angular or little-worn limestone lumps of 5 to 10 cm size, even larger locally, in stringers more or less parallel to the bedding. In the deposit groups of the Abruzzi and of Campania in Italy, and also at Unterlaussa in Austria, limestone breccias in a bauxitic matrix are known from the tapering, marginal parts of the deposits. According to Özlü (1977), this rock type forms three horizons in the Cenomanian bauxite deposits of the Western Taurids, Turkey.

Of the world total karst bauxite tonnage, 1·8% is made up of *carbonaceous (lignitic) bauxite* and *carbonaceous (lignitic) bauxitic clay*. These rock types have been identified in a total of 23 deposit groups. Their share is greatest in the Timan deposit type; even there, it is less than 10%, however. They are known from 7 of the 14 deposit groups of that type. In the 6 Chinese deposit groups, a carbonaceous bauxitic shale is found in the top parts of the deposits and along the intercalated stringers of coal. There is some bauxitic bone also in the Southern Timan deposit group, underlying the coal and carbonaceous clay of the immediate cover.

In ten of the Mediterranean-type bauxite deposits, these rock types are found as a rule under coal and carbonaceous clay in the immediate cover. They are most abundant (about 10%) at the Sohodol-Cîmpeni deposit group of Rumania, where the top part of the deposits is made up of black bauxitic bone (Papiu *et al* 1970). At Brignoles, France, bauxite in places passes gradually into the overlying coal (Nicolas 1970). The Devonian bauxites of the Northern Urals contain spots of a black *bituminous bauxite* in their top parts, with an organics content up to 12% (Dementev and Rodchenko 1966).

In Hungary, 1 to 2% carbonaceous bauxitic clay is attached to the brown coals or carbonaceous clays of the immediate cover. The only small intercalation of lignitic bauxite within a bauxite body was identified in the Bagolyhegy deposit of the Gánt group (Kiss and Vörös 1965).

Rock types of this nature are known from six of the Kazakhstan-type deposit groups. They tend to overlie the bauxite bodies within the bauxitic complex. According to Kirpal' (1972a), in the Arkalyk deposits, there is a black low-density bauxite lignite rock of real lignitic appearance. A lignitic bauxitic clay in a different, rather special position was described by Tyurin (1971) from beneath the main bauxite body of the Taldyk-Ashisaysk deposit at Mugodjar, USSR. It was 1 to 12 m thick.

No carbonaceous bauxite types have so far been identified in the deposit groups of the Ariège, Salento and Tulsk types.

Phosphoritic bauxite is more rare than even the foregoing types, amounting to 0·1% at most of total world karst bauxite tonnage. It has been identified so far exclusively in the deposit groups of the Loyalty, Fiji and Tonga island groups of the Pacific: there, however, it is fairly abundant at 20 to 50%. According to an oral communication by Gsell (1971), *phosphoritic bauxite* passes into a bauxitic phosphoritite on Lau Island of the Fiji Group.

Manganiferous bauxite is known so far from a single deposit group (Bédarieux in France). It is the top parts of the deposits that are enriched in manganese minerals: their abundance, however, seldom exceeds 10%; the maximum observed so far is 15%. Manganiferous bauxite can be recognised even by eye by its fallow greyish-brown colour. Field observations by the writer imply this bauxite type to make up at most 1 to 2% of the total deposit tonnage of Bédarieux.

4.2. PARTICLE SIZE OF BAUXITE

The individual mineral particles of most karst bauxites are finer than 1 μm. Particles in the 1 to 100 μm size range are much more rare. The device most suitable for the study of particle sizes and morphologhy in bauxites made up of particles finer than 1 μm is the electron microscope. Larger particles can be studied also by means of the electron microprobe and in thin sections by petrographic microscopes. The literature makes hardly any mention of true particle size distributions in bauxites. Hungarian bauxites have been studied in this respect by Árkosi who developed a method for the quantitative evaluation of particle size distributions in electron micrographs. First, an aqueous dispersion of the bauxite sample is prepared. In order to avoid aggregation, the dispersion is applied to the sample holder in comparatively great dilution. In the micrograph, the major axis (length) and the perpendicular axis (width) of each particle can be measured. The line-of-sight dimension (minor axis or thickness) can only be estimated. The meaningful evaluation of one sample requires 600 to 1000 measurements. Árkosi was kind enough to prepare electron micrographs of some typical samples of our bauxite collection, for which the writer takes this occasion to express his sincere gratitude.

Figure 67 shows plots of grain size distributions as established by Árkosi on two typical bauxite samples, one from Iszkaszentgyörgy and one from Nyirád, Hungary. The percentage distributions of particle length and width have been plotted in the same graph. Particle sizes are nearly identical in the two deposit groups, with 0·2 to 0·3 μm as the most frequent length and 0·1 to 0·2 μm as the most frequent width. The estimated most frequent thickness is 0·1 to 0·3 μm. The histograms show the particle size distribution to be highly uniform: the abundance of particles larger than 0·6 μm is insignificant; there are no particles larger than 0·9 μm. Árkosi found similar distributions in Halimba and Gánt bauxite. This uniformity of particle size distribution is rather remarkable because the mineralogical composition of these four deposit groups is not in any way identical.

Particle elongation can be quantified by calculating the length-to-width ratio of each individual particle. The ratio thus obtained is the greater, the more elongate the particle. Figure 68 shows the histograms of elongation of the two bauxite samples of figure 67. Elongation is less than two units in 67% of the particles in the Iszkaszentgyörgy sample and in 66·6% of the particles in the Nyirád sample. The majority of the particles, then, is hardly elongate. Particles of elongation greater than 3·5 make up less than one percent of both bauxites. The same has been found to be the case with samples of Gánt and Halimba bauxite.

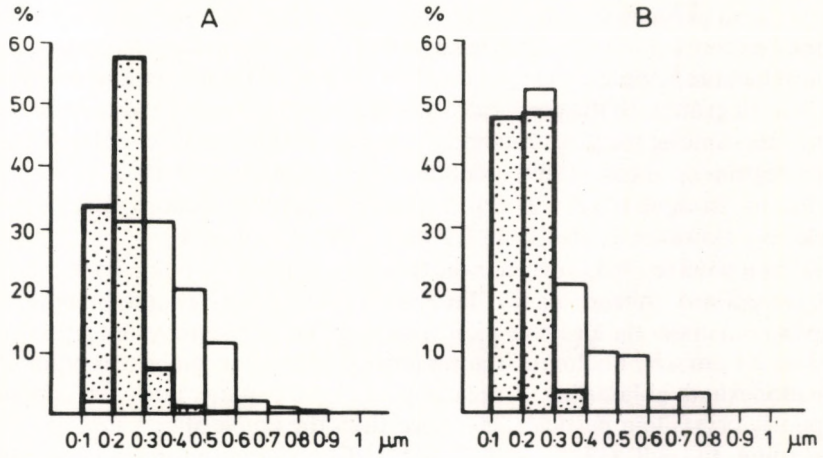


Figure 67. Particle size histograms of Hungarian bauxites as determined by transmission electron microscopy. The dappled parts indicate particle width (intermediate axis) distribution; the empty columns indicate particle length (major axis) distribution

(A) a type sample of the bauxite mine, Iszkaszentgyörgy, József II deposit; (B) idem, Nyirád, Lens Dűlt-Nyires VII

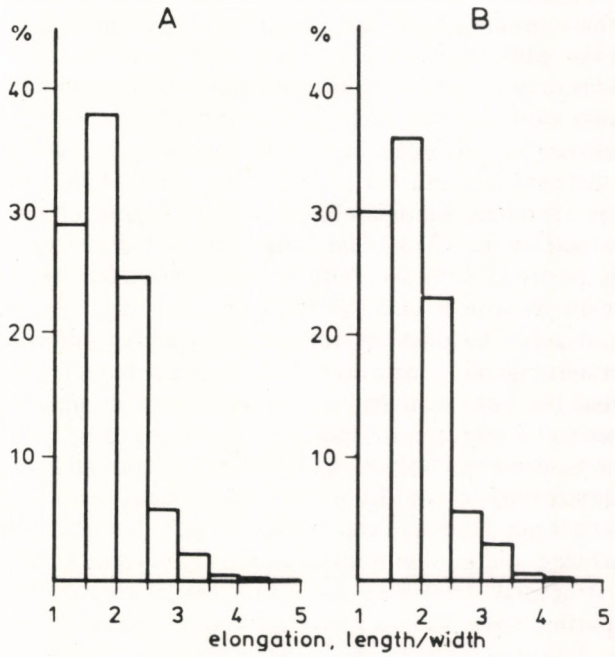


Figure 68. Frequency histograms of particle elongation in the two bauxites specified in Figure 67

The mineral particles making up the Hungarian bauxites have irregular outlines as a rule; they seldom exhibit regular crystal faces. Figure 69 presents electron micrographs of typical bauxite samples, one each from the Nyirád, Halimba, Iszkaszentgyörgy, and Gánt deposit groups. In the preparates, the individual particles tend to form clusters of varying size. Among the predominantly irregular-shaped particles of the bauxitic clay of Cserszegtomaj, Árkosi (1969) identified a few kaolinite crystals of about 1 μm size, bounded by straight edges defining a pseudohexagonal outline. The predominant particle size, however, is also from 0·1 to 0·2 μm in that rock.

As far as it is currently known, it is the Quaternary bauxites of the Pacific islands that are finest-grained. According to Tercinier (1971), Lifou Island bauxite (Loyalty Group) is seen under the electron microscope to contain boehmite and gibbsite crystals of 0·03 by 0·1 μm size. Boehmite is automorphic, columnar; gibbsite is either irregular or pseudohexagonal-lamellar. A greater part of the bauxite, however, is made up of very poorly crystalline particles finer even than the above-stated size.

According to Hill (1955) and Strahl (1971), Jamaican bauxite is composed predominantly of particles finer than 1 μm. The electron micrographs presented in their papers show near-isometric particles of exceedingly irregular, floccule-like outline. The particles have a tendency to form clusters of 40 to 100 μm size, in which they are presumably assisted by their highly irregular shapes. Certain clayey bauxite types contain some halloysite which tends to assume elongate, tubular forms.

Electron microscope study by Bárdossy *et al* (1978) revealed the particle size of bauxite from the Manchester Plateau, Jamaica, to be predominantly in the 0·15 to 0·25 μm range. The finest grains are of 0·05, the coarsest ones of 0·5 μm size.

Árkosi found the gibbsitic bauxites of the Arkalyk and Western Turgay deposit groups in Kazakhstan to be composed predominantly of particles of 0·2 to 0·6 μn size, isometric or hardly elongate, and irregular in outline.

Malm bauxite from Central Greece have an average particle size of 0·3 μm, with most particles falling between 0·2 and 0·6 μm (Bárdossy and Mack 1967). The average particle size of the Turonian–Senonian bauxite of that region, on the other hand, is 1 μm, with most particles in the 0·3 to 2·0 μm range. The particles of the Malm bauxite are nearly isometric, partly of irregular outline, partly bounded by crystal faces. The Turonian–Senonian bauxite is also isometric, but the majority of its particles is bounded by crystal faces. This phenomenon is in the writer's opinion due primarily to a difference in mineralogical constitution: the Malm bauxite is predominantly boehmitic, whereas the Turonian–Senonian bauxite is predominantly diasporic.

All the evidence so far indicates diasporic bauxite types to be coarser-grained with more regular particle morphologies than boehmitic and gibbsitic bauxites. The dominant particle size range of the Devonian diasporic bauxites of the Northern Urals, e.g. is from 0·4 to 5·0 μm. Diaspore crystals are largest, columnar; the iron minerals form smaller particles, and kaolinite forms thin flaky crystals smaller than 1 μm.

An increased structural stress may result in a partial recrystallisation of bauxite entailing larger particle sizes. The very strongly kneaded bauxite found in the deposit group of Tap Na, Vietnam, for example, largely consists of crystals of 5 to 50 μm size. Bauxite found on the Chalkidike Peninsula in Greece underwent in addition to strong structural stresses a weak metamorphism. The most frequent particle size of its minerals is 1 to 20 μm, but it contains some diaspore crystals of 100 to 200 μm size, too.

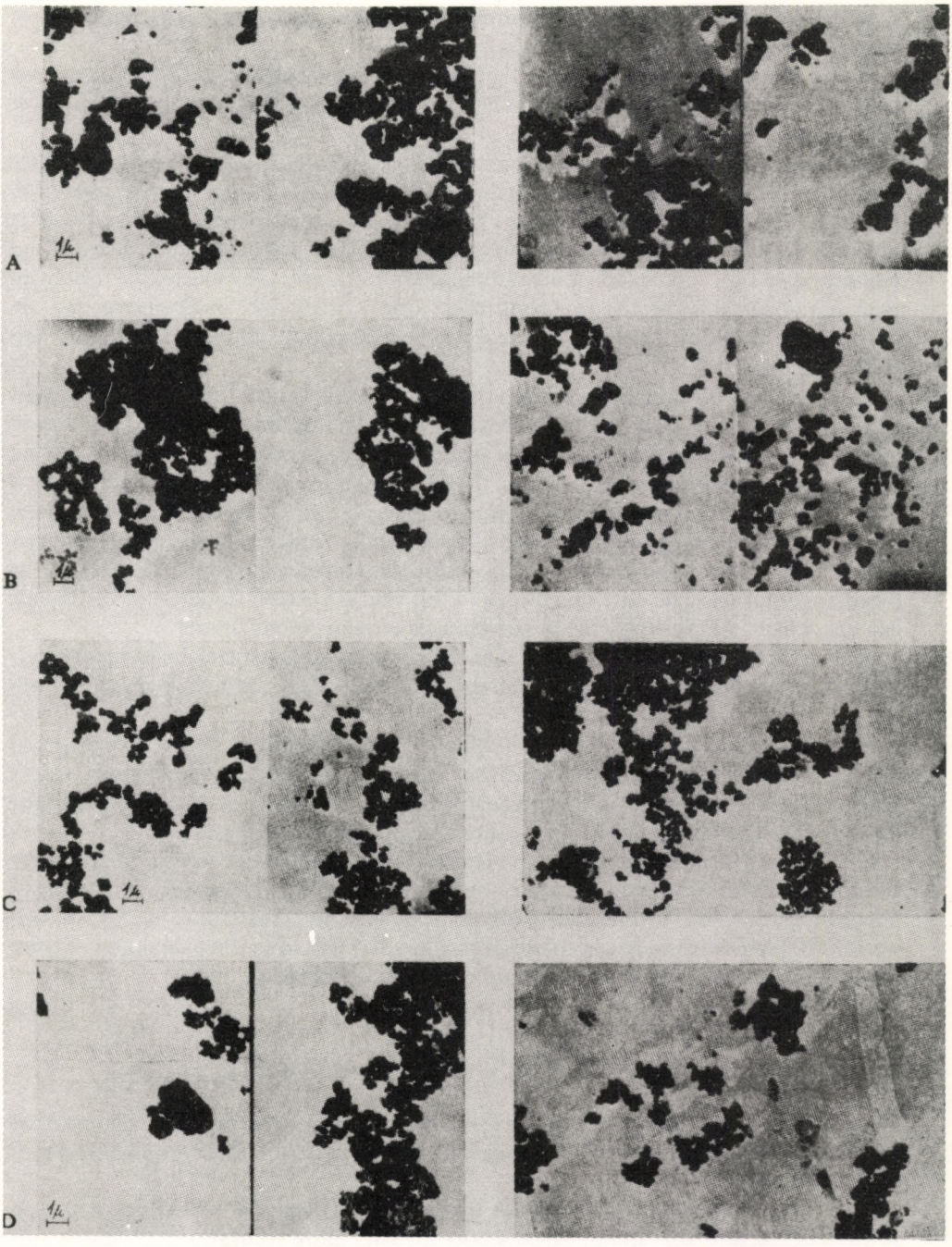


Figure 69. Electron micrographs of typical Hungarian bauxites, prepared by Arkosi

(A) Nyirád, Lens Dűlt-Nyires VII; (B) Halimba, Lens Malomvölgy XI/A; (C) Iszkaszentgyörgy, József II ore body; (D) Gánt, Újfeltárás deposit

Finally, strong metamorphism transforms bauxite into emery, the average grain size of which is in the millimetre range.

In contrast to the above, secondary minerals filling out pores, cavities or fine fissures tend to reach several millimetres.* They are to be discussed in more detail in the chapter on mineralogy.

Bárdossy *et al* (1978) also examined particle sizes in numerous lateritic bauxites. They found gibbsite crystals to be very large in young high-level type laterites (50 to 100 μm as a rule, but with frequent grains of millimetre size in pores and cavities). The average particle size of similarly young and similarly exposed (uncovered) karst bauxites is less by two orders of magnitude (0·05 to 0·2 μm). The reason for this much finer grain is felt to be the retarding effect that a carbonate environment has on crystallization. This idea was confirmed by the laboratory findings of Bárdossy and White (1979).

The particles of most karst bauxites, then, are of colloidal size, and therefore exhibit the peculiar physico-chemical features to colloids. Taking Buzágh's (1951) system of colloids as a basis, karst bauxites on the strength of their particle structure are so-called transitional systems; within that class, they belong to the so-called aggregates. As to their consistency, they are *xerogels*, as their disperse phase is solid and their medium of dispersion is gaseous. The term xerogel does not imply bauxites to be crystallographically amorphous: on the contrary, X-ray diffraction studies have revealed them to be largely crsystalline. The only exceptions to this rule are certain bauxites of the Pacific islands, which contain substantial percentages of amorphous matter.

The gel structure of karst bauxite, then, merely implies a statistically homogeneous aggregate of sub-microscopic particles, the interstices of the particles being filled with air.

Particle surfaces carry extremely thin films of adhesive water. These films may add up to a 10 to 20% moisture content in the deposit (in the natural, 'earth-moist' state). Bauxite samples taken out of the mine dry out rather rapidly at room temperature, losing the bulk of their original moisture content ('air-dry bauxite').

Bauxite as a xerogel has very great specific surface and surface tension. This is why some of its adsorbed water content (0·5 to 2·0 wt. %) can be expelled only by heating to between 100 and 400 °C. The finer-grained the bauxite, the greater the quantity of such water. The phenomenon has been established on the basis of a joint evaluation of more than one hundred DTA graphs and diffractograms (Bárdossy 1970a).

4.3. THE TEXTURE OF BAUXITE

Texture is meant to denote here the form and size of the mineral and rock (bauxite) particles making up a bauxite, the ways of their aggregation, their relative positions and arrangement.

The macroscopic features of bauxite texture can be observed by the naked eye on surfaces of fracture. Cut and polished sections permit more accurate observations. For comparison, these have been recorded by the writer on black-and-white and colour

* To fall in the size range from several hundred micrometres.

photographs. For the study of microscopic and submicroscopic textural features, it is the electron microprobe that is best suited, in addition to optical microscopy on thin sections. The application of the electron microprobe to bauxite was developed less than a decade ago (Bárdossy and Pantó 1970, 1971). These studies were complemented by the study and photography of thin sections for comparatively low-iron bauxites and of polished sections for iron-rich ones.

4.3.1. THE TEXTURAL ELEMENTS OF BAUXITE

In international literature, there is no consensus as to the terms to be used to characterise bauxite textures: and, what is worse, one and the same term often covers widely divergent contents. This is why the present writer felt obliged to develop a uniform system of textures applicable to all karst bauxites.

A texture is made up of *textural elements*, and so the first task is to propose a system of how such elements are to be distinguished.* The simple ones consist of a single monomineralic grain; the composite ones are complexes of several such grains. Textural elements, however, can be classified much more satisfactorily by a joint application of morphological and genetic criteria. The classification developed by the writer, to be outlined below, is based on such considerations.

4.3.1.1. MATRIX

An aggregate of uniform distribution of the finest particles, of more or less uniform size, contained in the bauxite under consideration. All the other—comparatively larger —textural elements of the bauxite are embedded in it. The matrix may belong to one of the following types, depending on its average particle size.

(i) *Pelitomorphic matrix*. Average particle size, less than 1 μm. No individual mineral particles can be distinguished on polished surfaces (colour figure 3). Even in an electron microprobe image providing a magnification of 600 diameters, clusters of particles can be distinguished at best (fig. 70). The individual mineral grains making up the cluster are resolved only under a magnification of several thousand times, as provided by the electron microscope (fig. 71).

(ii) *Microgranular matrix*. Average particle size, 1 to 5 μm. Individual mineral grains can be distinguished, although rather vaguely, on microprobe images and in thin sections.

(iii) *Panidiomorphic matrix*. Average particle size, 5 to 100 μm. This is the texture usually found in strongly recrystallised old bauxites and in any weakly metamorphic bauxite. Individual mineral grains can be readily distinguished in a thin section. The grains show polygonal, mosaic or dovetail contacts.

(iv) *Macrocrystalline matrix*. Average particle size, greater than 100 μm. This type of matrix is observed exclusively in metamorphic bauxites (emeries).

* On a mineralogical basis, simple and composite textural.



Figure 70. Electron microprobe image of the sample shown in colour figure 3. Electron image, 600 ×



Figure 71. Transmission electron micrograph of the sample shown in colour figure 3. 25,000 ×

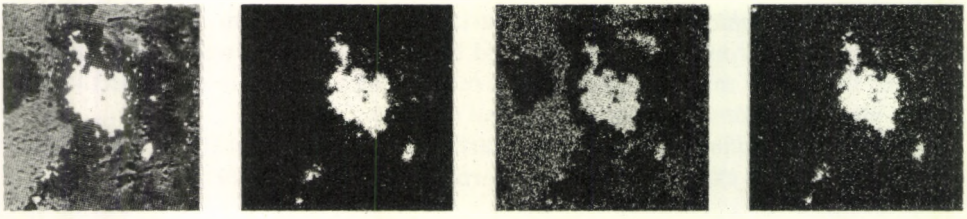


Figure 72. Electron microprobe image of an authigenic chalcopyrite grain. Devonian bauxite from the Northern Urals, USSR, 600 ×

(1) electron image; (2) X-ray image of sulphur distribution; (3) X-ray image of iron distribution; (4) X-ray image of copper distribution

4.3.1.2. AUTHIGENIC MINERAL GRAINS

Individual mineral grains formed within the bauxite in its present lie, they are readily distinguished from the embedding matrix by their larger size. They are also found in addition to the matrix in other textural elements such as pisoids. They are larger in pisoids than the mineral grains of their immediate surroundings.

Authigenic mineral grains exhibit no traces of transport or wear. Some are automorphic, delimited by crystal edges; others are irregularly branching (fig. 72). The latter are clearly the fillings of pre-existing micropores. Spherical siderite grains (spherolites) are also grouped with the authigenic mineral grains.

4.3.1.3. PSEUDOMORPHS

Individual mineral grains or clusters of such, occupying the place of a pre-existing mineral grain, preserved its shape. Pseudomorphs are as rare in karst bauxites as they are frequent in lateritic bauxites. They are usually difficult to identify, as their outlines tend to be obscured by secondary chemical processes (fig. 73).

4.3.1.4. CLASTIC MINERAL GRAINS

Usually grains of hard mineral species resistant to wear and surface weathering, their clastic origin is revealed by worn shapes (fig. 74). They are larger than the mineral grains of the embedding matrix, their usual size range being from 5 to 200 μm.

4.3.1.5. CLASTIC BAUXITE PARTICLES

Bauxite fragments delivered to their present lie by transport and redeposition. The writer has defined their upper size limit as 2 mm, and uses the term *microfragments* for those of less than 60 μm size. Angular or worn, they tend to be richer in iron and poorer in silica than the embedding matrix (fig. 75). The probable reason for this is that the

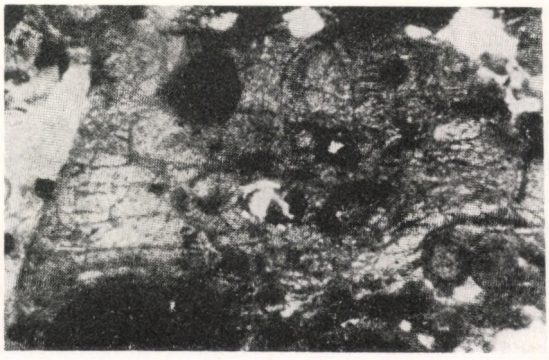


Figure 73. Boehmite: a pseudomorph after pyroxene. Cenomanian–Turonian bauxite, Cusano Mutri, Campania, Italy, 100×, parallel nicols

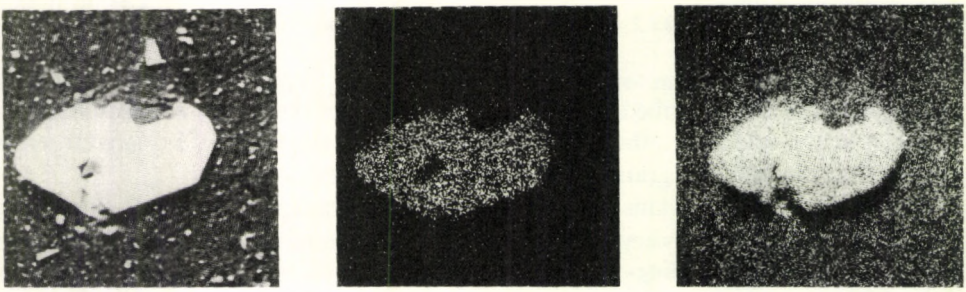


Figure 74. A zircon grain of detritic origin. Upper Cretaceous bauxite, Mazaugues deposit, Brignoles, France, 600×, electron microprobe

(1) electron image; (2) X-ray image of zirconium distribution; (3) X-ray image of silicon distribution

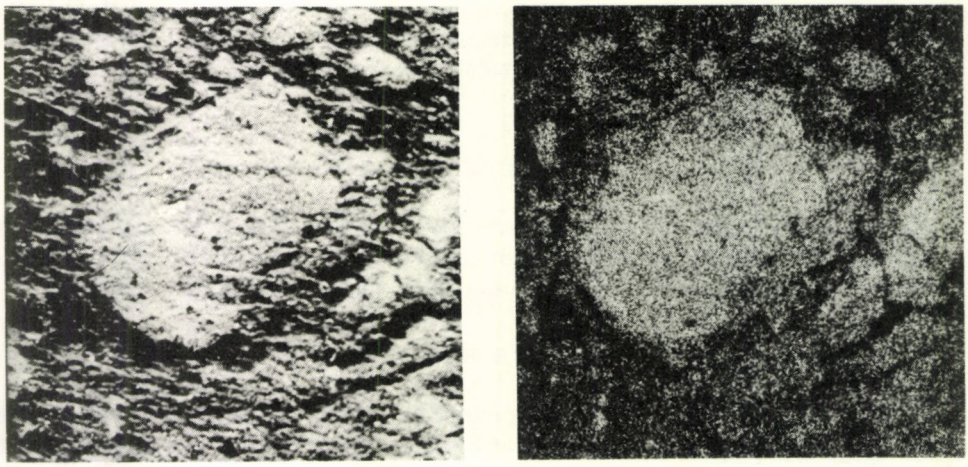


Figure 75. Clastic bauxite particles in a pelitomorphic matrix. Dűlt-Nyires VII lens, Nyirád, Hungary. Electron microprobe images, 300×

(A) electron image; (B) X-ray image of iron distribution

more clayey types of bauxite tend to elutriate more readily in the course of reworking. Bauxites richer in iron are harder as a rule and, hence, more resistant to wear. Some deposits contain bauxite fragments very high in iron, whose typically collomorphic texture implies that they presumably originated in a pre-existing lateritic ferricrete crust (a 'cuirasse').

Some authors consider bauxite fragments to be concretions formed *in situ*. Nia (1968), for example, denies the clastic origin of the bauxite particles described by Bárdossy and Mack (1967) from the Turonian bauxite horizon of Central Greece, even in cases where the grain surface is worn or exhibits a broken ooid, and adduces the processes of concretion formation as an explanation. In certain cases, the concretion may have broken up, and its parts may have drifted apart. No convincing evidence is produced, however, for the cause either of breakage or of drift.

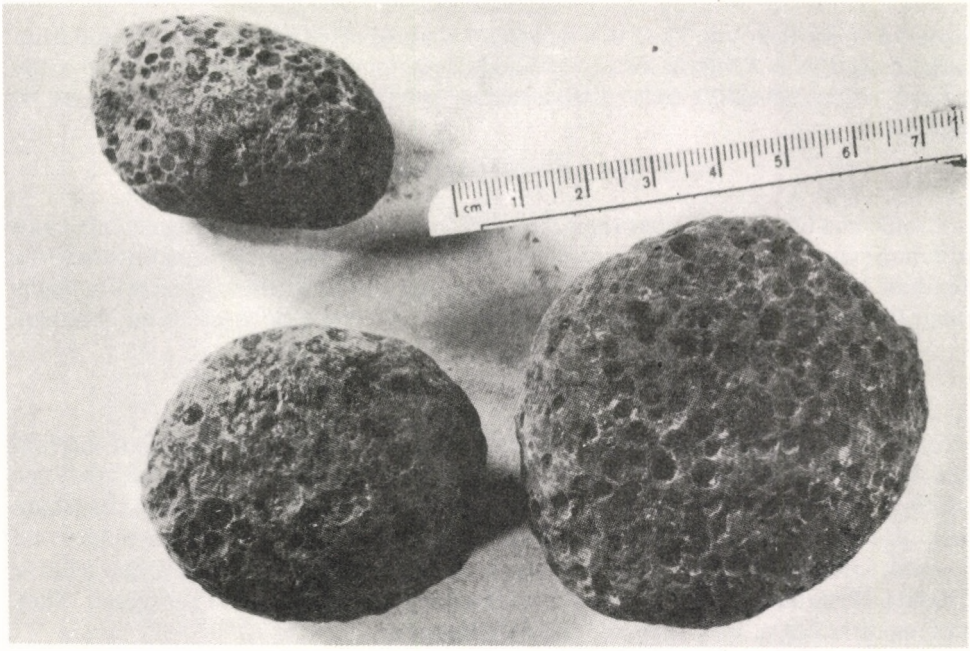


Figure 76. Composite bauxite pebbles. Eplény I open pit, Hungary

In the present writer's opinion, clastic particles can be distinguished from concretions by their shapes (by the fact that their sharp outlines cut across pre-existing textural elements, and by their bauxite-like mineral composition concretion in the writer's experience are monomineralic or composed of two or three mineral species at most).

4.3.1.6. BAUXITE PEBBLES AND DETRITUS (CLASTS)

These differ from the clastic bauxite particles by being of large size. Called detritus (clasts) if angular or sharp-edged and pebbles if rounded, they can be further subdivided in terms of the particle size limits usual in sedimentary petrography:

0·2 to 0·5 cm: fine bauxite pebbles or detritus (clasts)
0·5 to 2·0 cm: medium bauxite pebbles or detritus (clasts)
2·0 to 10·0 cm: coarse bauxite pebbles or detritus (clasts).

Bauxite pebbles made up partly or entirely of clastic bauxite grains or smaller bauxite pebbles are said to be composite. They indicate the repeated primary reworking of bauxite (fig. 76). Bauxite pebbles and detritus are usually richer in iron and harder than their surroundings. This is also attributed to selective wear.

Pastukhova (1975) described the *in situ* bauxitisation of pebbles and fragments of alumosilicatic rocks of clastic origin from several bauxite deposits in the Kazakhstan. The process of bauxitisation has preserved the original forms of the particles of sand and pebble size. Pastukhova proposed the term *'relict clastic grains'* with a view to distinguishing them as separate textural elements. Such particles may indeed have

formed in some deposits. Their distinction from transported bauxite pebbles and from bauxite detritus is liable to be very difficult, sometimes impossible. This is why the present writer considers their distinction as independent textural elements to be unjustified.

4.3.1.7. BAUXITE BOULDERS

Clastic bauxite lumps larger than 10 cm in diameter, they are slightly rounded as a rule, hard and iron-rich. Their surface layer is often impregnated by iron minerals to a depth of 2 to 5 cm; the enrichment of iron is strongest at the surface. Typical diameters range from 10 to 50 cm, but some boulders of one metre size have also been observed.

4.3.1.8. SPHEROIDS ('ROUNDGRAINS')

Spherical or ellipsoidal particles of either homogeneous or disordered internal structure, they were called 'pellets' by Bushinsky (1971). Seen from outside, they are readily mistaken for the ooids and pisoids to be described below, with which they share their spherical or ellipsoidal outer form. The insides of roundgrains exhibit no concentric shells, however. Most roundgrains are of clastic, reworked origin (that is completely rounded particles of bauxite). More seldom, they may be authigenic. They tend to be richer in iron than the surrounding matrix.

Some roundgrains are entirely homogeneous inside (fig 77), others are inhomogeneous in spots. They are often crisscrossed by an irregularly branching network of fissures filled with aluminium or iron minerals. These fissures end at the roundgrains surface, and do not pass into the embedding matrix (fig. 78). The fissures presumably formed in the course of reworking, as a result of temporary dessication or gel compaction. Some roundgrains broken during reworking had their broken edges worn down in the course of transport (fig. 79).

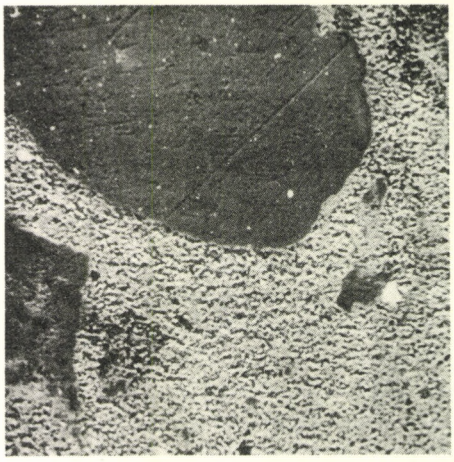


Figure 77. A roundgrain of pelitomorphic texture embedded in a microgranular matrix. Devonian bauxite, Northern Urals, USSR. Electron microprobe, 300 ×, electron image

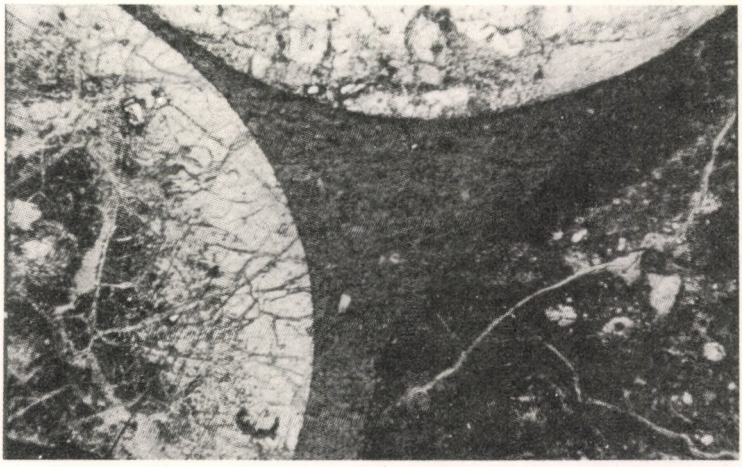


Figure 78. A roundgrain traversed by veins. Halimba–Malomvölgy, Hungary. Polished section, 70×



Figure 79. A broken and subsequently worn roundgrain Mazaugues deposit, Brignoles, France. Electron microprobe, 300×, electron image

4.3.1.9. PARTICLES OF CONCENTRIC (ORBICULATE) STRUCTURE

Particles made up of darker and lighter concentric shells are clearly visible under the microscope or in the electron microprobe image. The shells surround a central nucleus. The number of shells is 2 to 10 as a rule but may exceed 20 sometimes. Adjacent shells may merge gradually into one another or be sharply distinct. A shell may be from 1 to 100 μm thick. Within individual shells, a vague concentric microstructure can often be observed; shells of homogeneous texture all over are more rare. Some shells may embed authigenic or clastic mineral grains.

The texture of the nucleus is different from that of the outer, shelly (orbiculate) part. It consists of a larger clastic or authigenic mineral particle, a clastic bauxite grain, or

some other, smaller transported textural element. More seldom, it is made up of several smaller particles.

Concentric particles are most often spherical or ellipsoidal. Using a time-honoured terminology, they can be classifified as follows.

micro-ooids — diameter less than 100 μm
ooids — diameter 100 to 1000 μm
pisoids — diameter 1 to 5 mm
macropisoids — diameter greater than 5 mm.

Nia (1968) called all concentric particles of diagenetic origin pisoliths, irrespective of their size. Although the four groups distinguished above have many features in common, the present writer nevertheless feels it necessary to distinguish them. Firstly, they denote textural differences well visible to the naked eye, and secondly, they have distinct sedimentological and diagenetic connotations.

Using a term coined by Carrozzi, the writer calls *spastoids* all concentric particles of irregular form, regardless of size.

(i) *Micro-ooids*. Comparatively rare, they tend to occur in ooidic bauxites. Owing to their small size, they tend to have just a few shells, most often rather vaguely delimited ones. Well-defined nuclei are rare (fig. 80).

(ii) *Ooids*. The most frequent concentric particles of bauxites; their shells tend to be sharply distinct. Their nuclei are often largish authigenic mineral grains (e.g. of haematite, figure 81). It is clear in the figure how even the innermost, iron-rich shells gradually smooth out the rambling outline of the authigenic haematite grain. The third shell from the centre already has a regular circular outline in cross section. The right-hand picture (the X-ray distribution image of $Fe_{K\alpha}$) reveals the shells to be alternately higher and lower in iron.



Figure 80. Indistinct concentric shells in micro-ooids. Senonian bauxite, Bosanska Krupa, Yugoslavia. Electron microprobe, 600×, electron image

The nucleus of an ooid may be just as often a clastic mineral grain, most often a zircon grain, more seldom a fine particle of clastic bauxite (fig. 82). In that case, too, the outward-increasing sphericity of the particles is readily observed.

Structures similar to those shown in figure 83 are comparatively rare. The ooid in that figure has no well-defined nucleus; the innermost shell surrounds an aggregate, of similar texture as the matrix, of crypto- and microcrystalline bauxite minerals. The shells are eccentrically arranged. This is a comparatively rare phenomenon; most of the

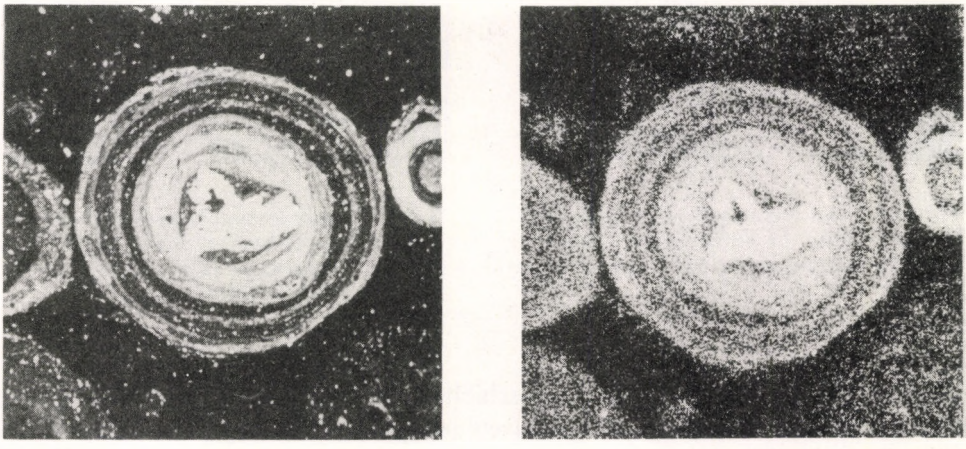


Figure 81. An ooid with an authigenic haematite grain for a nucleus. Senonian bauxite, Bosanska Krupa (Yugoslavia). Electron microprobe images, 300 ×
(A) electron image; (B) X-ray image of iron distribution

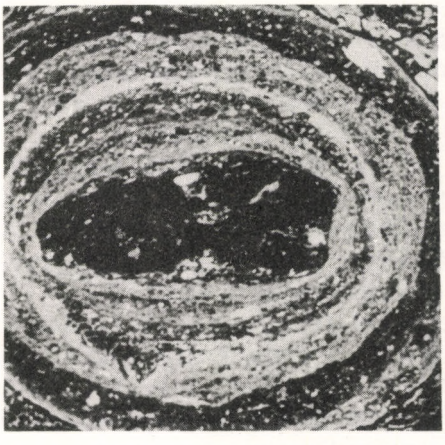


Figure 82. An ooid with a clastic bauxite particle for a nucleus. Triassic bauxite, Skočaj deposit, Lika County, Yugoslavia. Electron microprobe, 300 ×, electron image

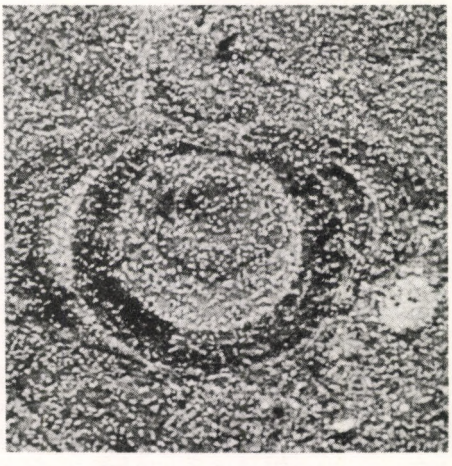


Figure 83. An ooid the texture of whose nucleus resembles that of the matrix. Hosszúharasztos deposit, Gánt, Hungary. Electron microprobe, 300 ×, electron image

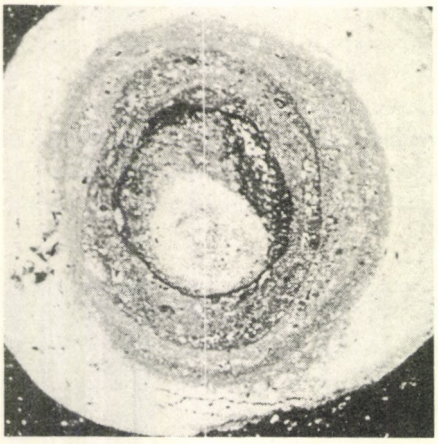


Figure 84. An eccentrically concentric ooid. Dragoni deposit, Monti Maggiore, Campania, Italy. Electron microprobe, 300 ×, electron image

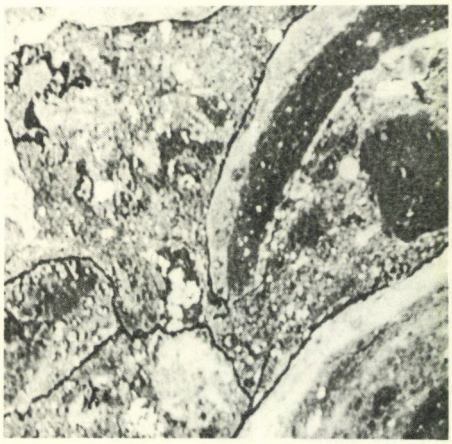


Figure 85. Detail of the outer shells of a redeposited ooid. Triassic bauxite, Skočaj deposit, Lika country, Yugoslavia. Electron microprobe, 300 ×, electron image

ooids ever studied by the writer have exhibited a regular concentric arrangement of shells. The writer has found comparatively many ooids of eccentric structure in the Italian bauxites (fig. 84).

The nucleus-to-shells area ratio is rather varied. In many-shelled ooids, the nucleus may take up just one-fifth to one-fourth of the entire ooid. In ooids with comparatively large nuclei or in ooids of few shells, the nucleus may occupy up to two-thirds of the entire cross-sectional area of the ooid.

The outer shells of some larger ooids embed minute micro-ooids. This can be explained by assuming the rates of growth of adjacent ooids to have been different. The faster-growing ooid engulfed its neighbours, and continued to grow after having done so. The first to describe this phenomenon was Jurković (1965) in connection with the Senonian bauxite of the Suvaja-Šolaja group of deposits in Yugoslavia.

The present writer regards the majority of undamaged ooids as authigenic, diagenetic textural elements. Broken ooids or ooids of worn edges have, however, evidently undergone transport; they were formed during the brief interruptions of primary bauxite redeposition. Reworking had broken up some of the ooids so that only fragments of their harder, more iron-rich outer shells were left over (fig. 85).

Such broken ooids are also fairly frequent in the Turonian bauxite horizon of Central Greece. According to Nia (1968), these are not due to transport but to fragmentation along radial fractures produced by gel compaction; the fragments then drifted apart in a still semi-plastic matrix. This mechanism of formation is hard to imagine as no broken fragments that could be fitted together can be found anywhere in the vicinity (the present writer has made repeated attempts to discover such fits over fairly large rock surfaces). Moreover, the broken edges of the ooid fragments often exhibit a slight wear, as an incontrovertible proof of transport.

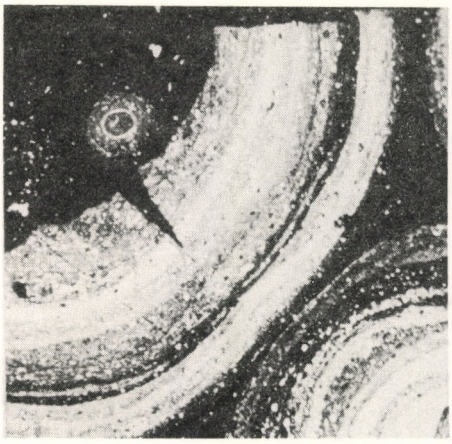


Figure 86. An ooid fragment, redeposited and cracked after sedimentation. Senonian bauxite, Bosanska Krupa, Yugoslavia. Electron microprobe, 300 ×, electron image

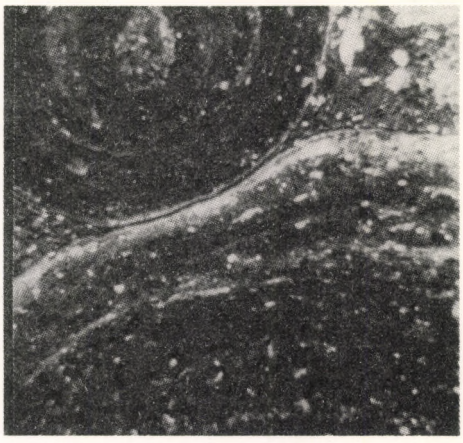


Figure 87. Plastic deformation in ooids. One ooid denting another while being itself deformed. Triassic bauxite, Skočaj deposit, Lika country, Yugoslavia. Electron microprobe, 300 ×, electron image

In the bauxite of the Bosanska Krupa deposit group in Yugoslavia, the author has found a broken ooid cracked by overburden pressure after deposition (fig. 86). In addition to the wedge-shaped crack, the ooid also underwent some plastic deformation where it pressed against an adjacent ooid. The ooid fragment, then, was not completely hardened at the time of deposition: it was still capable of plastic deformation.

In certain bauxites, ooids are packed close together, with the more hardened ones denting the softer ones (fig. 87). This mechanical heaping can be readily distinguished from cases of authigenic ooid growth where ooids lying side by side come to touch and intergrow, as it were. In such cases, the shells in contact gradually pinch out near the contact surface.

In bauxite deposits exposed to structural stress or to considerable overburden pressure, the ooids are flattened parallel to the deposit top, acquiring a lenticular or elliptical shape in cross section (fig. 88).

In some bauxites, shell formation started anew after the deposition of the broken ooids (fig. 89). The successive shells gradually smooth out the broken outline, exhibiting forms of outward-increasing sphericity. This phenomenon provides some circumstancial evidence as to the formation of all concentric particles. It seems as though, during the diagenesis of the matrix the material tended to assume a textural arrangement of least specific surface, a sphere or ellipsoid. This is the reason for the increasing sphericity of the outer shells.

Ooids including a smaller ooid, broken or entire, are termed *composite ooids* by the writer.

Rather seldom, particles of oxidised magnetite, pyrite, marcasite or siderite are surrounded by a low-iron zone 10 to 30 μm wide, surrounded in its turn by a zone of about the same width made up of finely dispersed haematite or goethite. These textural

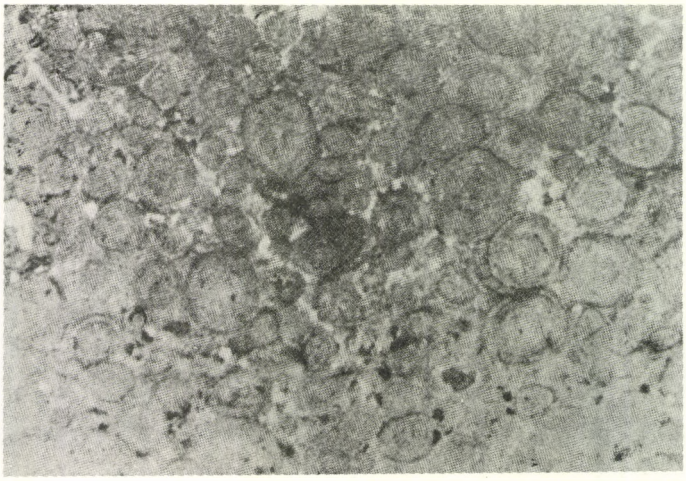


Figure 88. Ooids deformed by overburden pressure. Carboniferous bauxite, Gun district, Honan Province, China. Thin section, after Schüller (1957)

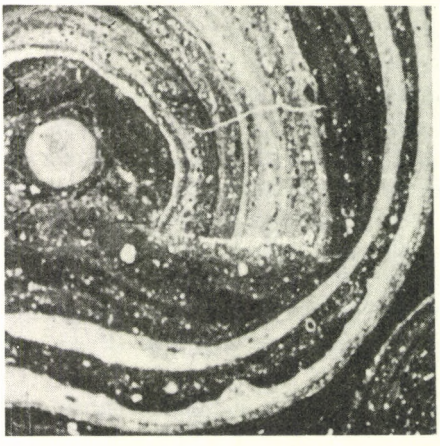


Figure 89. Part of a broken ooid, growing on after redeposition. Senonian bauxite, Bosanska Krupa, Yugoslavia. Electron microprobe, 300×, electron image

elements of epi- or supergene origin, recalling simple ooids in appearance, are termed *pseudo-ooids* by the writer. They are due to the outward diffusion and precipitation of ferric hydroxide during the process of oxidation.

(iii) *Pisoids*. They differ as a rule only in their larger size from the ooids; their textural features are much the same otherwise. A larger size, however, does not invariably imply a larger number of shells. As a rule, either the nucleus of the pisoid is greater, or the shells are thicker. The outermost shell tends to be richest in iron. Pisoids tend to be darker-coloured, higher in iron than the embedding matrix. They are frequently also harder, in which case they do not break when the bauxite is broken, but emerge like scattered ball-bearing balls above the fracture surface.

Flattened lens or ellipsoid shapes are more frequent among pisoids than among ooids. The writer could observe in a number of instances ooids of a well-preserved spherical shape embedded in the vicinity of flattened pisoids. It is to be presumed that the pisoids, being larger, are flattened under their own weight as well as by overburden pressure. Combes (1969) also is of the opinion that the flattened pisoids found in the bauxites of the Ariège were deformed by their own weight prior to full hardening.

In bauxites that underwent strong secondary processes, the original concentric structure of the pisoids may have been destroyed. The original material may have been leached away completely, leaving the interior porous or spongy, with a replacement in the form of secondary minerals (mostly goethite, less seldom calcite) taking its place.

The pisoids of reduced bauxites often contain very fine authigenic grains of pyrite or marcasite. Wherever a bauxite of this sort is subsequently oxidised somewhere near the ground surface, the pyrite/marcasite gets oxidised and is removed in a sulphate solution. The remaining material constitutes a spongy filling in the interior of the one-time pisoid; the places of the former pyrite/marcasite grains are marked by minute angular cavities.

(iv) *Macropisoids* ('beans'). Rare in karst bauxites, they are rather frequent in lateritic bauxites. Their usual size range is from 5 to 15 mm; larger ones are very rare. The writer has so far found none greater than three centimetres in karst bauxites: nor have any been described in the literature. Just as in the pisoids, it is not the number of shells that increases with size but the diameter of the nucleus above all and, more seldom, the thickness of the shells. The outermost shell tends to be thickest and richest in iron. They are darker-coloured than the embedding matrix; and are often crisscrossed by hair cracks (colour figure 4). Macropisoids often exhibit an epigene deferrification, especially near the deposit tops. Their internal structures are usually destroyed: they may be filled out with very loose, porous secondary precipitations of goethite. Sometimes, only a spherical hole marks their prior existence.

(v) *Spastoids.* These textural elements are much more rare than the foregoing ones. The first to describe them were Papiu and Mînzatu in 1969, from the bauxite at Pădurea Craiului in Rumania. Their usual size range is from 0·5 to 3 mm. They are made up of a smaller number of shells that are less clear-cut than in the ooids and pisoids.

The amoeba-like rambling surface of the spastoid in figure 90 is coated with a gibbsite film a few micrometres thick.

4.3.1.10. FILMS AND COATINGS ON PRE-EXISTING TEXTURAL ELEMENTS

In the bauxite types exposed to dia- or epigenetic processes of some intensity, various textural elements are often observed to be surrounded by a film or coating ranging in thickness from a few μm to 2 mm. The film or coating, most often yellowish or brownish in colour, tends to consist predominantly of goethite. It is usually homogeneous, with just a vague zoning in certain cases. Coatings tend to gradually round out the angular forms of the textural elements that they cover. Their outer surfaces form a sharp limit towards the matrix. Figure 91 shows an iron-rich bauxite particle of clastic origin in a secondary coating.

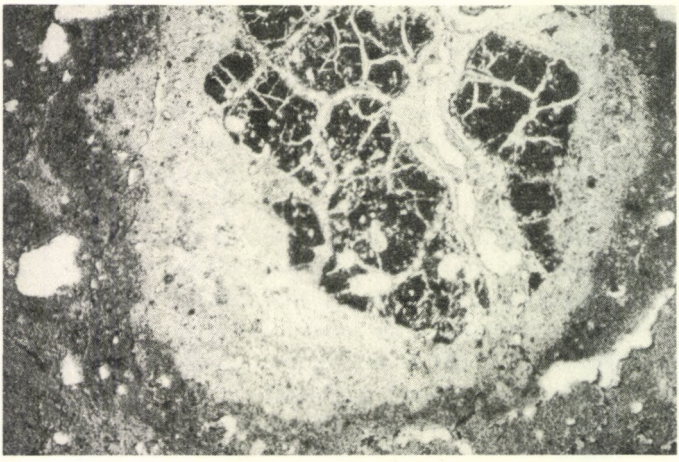


Figure 90. A spastoid in Paleocene–Lower Eocene bauxite. Georgia, US. Thin section, parallel nicols, 60 ×



Figure 91. A secondary coating surrounding an iron-rich bauxite grain of clastic origin. Dragoni deposit, Monte Maggiore, Campania, Italy. Electron microprobe, 300 ×, electron image

These films and coatings, then, are similar in many respects to the spherical shells of the ooids and pisoids, but are later then those in a majority of cases. The writer has observed in one of the open pits of the Villeveyrac Basin in France a fine example proving their secondary origin. The brick-red bauxite there contains numerous pisoids and bauxite fragments of 1 to 10 mm size, as well as some bauxite pebbles. All these are somewhat darker, harder and more compact than the matrix (cf the parts marked 3 of figure 92). Near the top of the deposit, the matrix was discoloured to a pink colour by a secondary leaching of iron: the more compact pisoids and bauxite pebbles, however, retained their original dark-red colour (parts marked 1 of figure 92). In the bottom part of the deferrified bauxite, in a zone of 20 to 40 cm thickness, the pisoids and bauxite



Figure 92. Secondary goethite coatings on bauxite pebbles and pisoids. Villeveyrac deposit, Languedoc, France

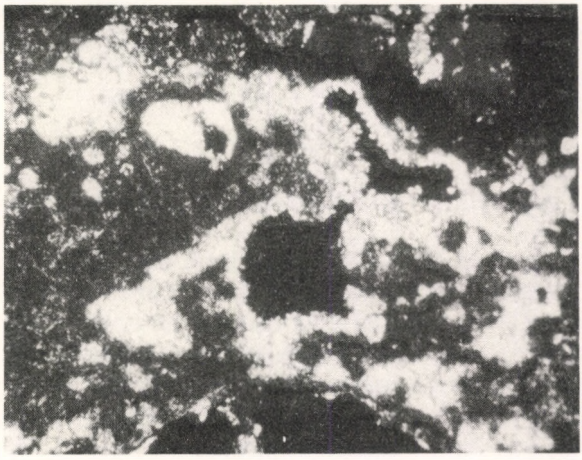


Figure 93. Secondary segregations of gibbsite filling pores in a bauxite. Paleocene–Lower Eocene bauxite, Chattanooga, Tennessee, US. Thin section, crossed nicols, 60 ×

pebbles are coated by 0·5 to 2·5 mm of ochroid or yellowish brown goethite (parts marked 2 of figure 92). No such coatings could be detected either in the red bauxite below or in the deferrified bauxite above. In the writer's opinion, this distribution came about by the precipitation on the more compact, more impermeable textural elements, in the form of coatings, of some of the iron dissolved in the course of deferrification.

4.3.1.11. PORE FILLINGS

The micropores and pores of bauxite are often filled with minerals of secondary origin. These are sharply distinct from the surrounding bauxite. Their forms and outlines are irregular, copying the shape of the former pore. The minerals in the pore fillings are much coarser grained as a rule than the surrounding bauxite (fig. 93). The larger pores may be incompletely filled, with secondary crystals sitting on the pore walls as in a geode.

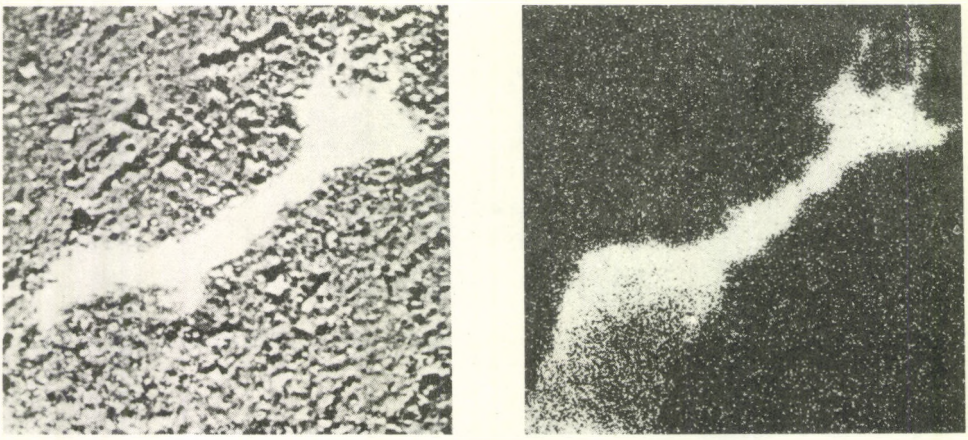


Figure 94. An iron-rich segregation of irregular form in the bauxite matrix. Mazaugues deposit, Brignoles, France. Electron microprobe, 600 ×
(A) electron image; (B) X-ray image of iron distribution

4.3.1.12. SEGREGATIONS OF IRREGULAR FORMS (NODES)

Irregular in outline, these are not separated from their surroundings neatly in all directions (fig. 94). Their internal structure is uniform or irregularly inhomogeneous. An upper size limit of 2 cm has been fixed for them by the writer. The term 'node' carries the connotation of a roundish shape. These elements tend to be of late-diagenetic or epigene origin. Due to processes partly of impregnation and partly of displacement, they do not fill out pre-existing pores. Most are rich in iron. In the literature, they have been repeatedly—and wrongly—classed as pisoids.

4.3.1.13. NESTS, CONCRETIONS

These are of epi- or supergene origin, larger than 2 cm, irregular as to shape (although some are reniform or spheroidal). Some, especially those rich in iron, are not neatly distinct from their surroundings. Goethite constitutes concretions of a peculiar radial–fibrous internal texture in some deposits at Sümeg and Nyirád, Hungary

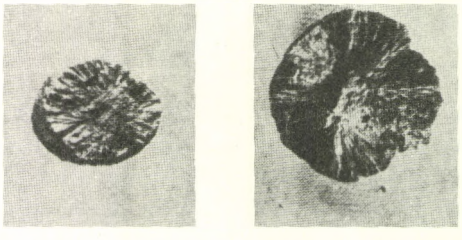


Figure 95. Goethite concretions of radial-fibrous structure. Lens VI, Nyirád, Darvastó, Hungary, natural size



Figure 96. Vestiges of roots filled with haematite and goethite. Eplény II open pit, Hungary

(fig. 95). These concretions are very hard and compact, with shiny surfaces. The broken or worn surfaces of some of them imply a process of reworking.

The manifold mineralogy of concretions and nodes is to be discussed in the mineralogical chapter.

Concretions formed along former plant roots constitute a distinct subgroup within this group. They tend to be tubular, of to 20 mm diameter, tapering downwards, round or oval in cross section. Ranging from 4 to 20 cm in length they usually are vertical or near-vertical in the bauxite. Their outer parts are harder; their insides are porous or spongy (fig. 96). In the writer's opinion, they formed after, or at the time of, the decomposition of the original roots, under reducing conditions.

4.3.1.14. FISSURE FILLINGS

It is appropriate to distinguish two subgroups of fissures here. One is confined to within certain textural elements (e.g. pisoids, roundgrains, bauxite pebbles), and does not pass into the surrounding matrix. Its formation is connected with the origin of the textural element in question, usually to processes of gel compaction and de-watering (syneresis). Some roundgrains are traversed by quite dense, grid-like or radiating networks of hair veins (fig. 78).

The other group forms networks in the matrix. The finest such veins are of just a few μm diameter and of rather limited length (fig. 97). The writer feels that these too were formed by gel compaction.

Most fissure fillings are more or less plane, straight in section, cutting through the textural elements in their way: they are sharply distinct from the surrounding bauxite. They vary in thickness from a tenth of a millimetre to several centimetres. The minerals

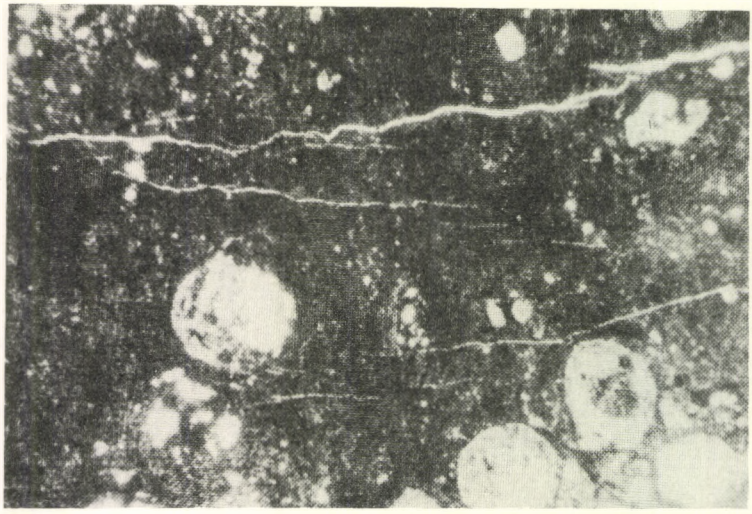


Figure 97. A fine network of diaspore-rich veins traversing the bauxite matrix. Paleocene bauxite, Kozina, Slovenia, Yugoslavia. Thin section 60×

making up the fillings are of unusual size (in the 0·1 mm to several millimetres range). Most fissures are due to structural strains, with displacements of millimetre to centimetre size evident along some of them.

Some especially large fissures are not filled out completely with minerals: only on the two walls of bauxite are there large automorphic, inward-projecting crystals. These too shall be described in the chapter on mineralogy.

There are a few rare cases where the bauxite is not sharply distinct from the veins, where the veins fan out and merge into the matrix, as cf the titanium-rich veins found in the Pădurea Craiului bauxite, Rumania (fig. 98). The photo shows clearly how the roundgrains embedded in the matrix are bypassed by the veins.

4.3.2. TYPES OF TEXTURE

Textural types can be distinguished on the basis of the frequency distribution of textural elements. Their terminology and classification currently depends entirely on the personal judgment of the observer and cannot therefore be generalised. The writer feels that objective textural characterisations of general validity must inevitably be based on the relative abundances, expressed in percentage areas, of the textural elements involved. No systematic measurements of the sort have so far been carried out on bauxites.

In the course of his studies, the writer has tried to develop a method suitable for an accurate assessment of bauxite textures. Thin sections and polished sections, however, are suited only for the assessment of textures made up of textural elements finer than one millimetre. The measurement can be performed using the usual microscope attachments (integrating tables). The textural elements of bauxites, however, cover as a rule such a broad size range that it is necessary to evaluate a polished rock surface at

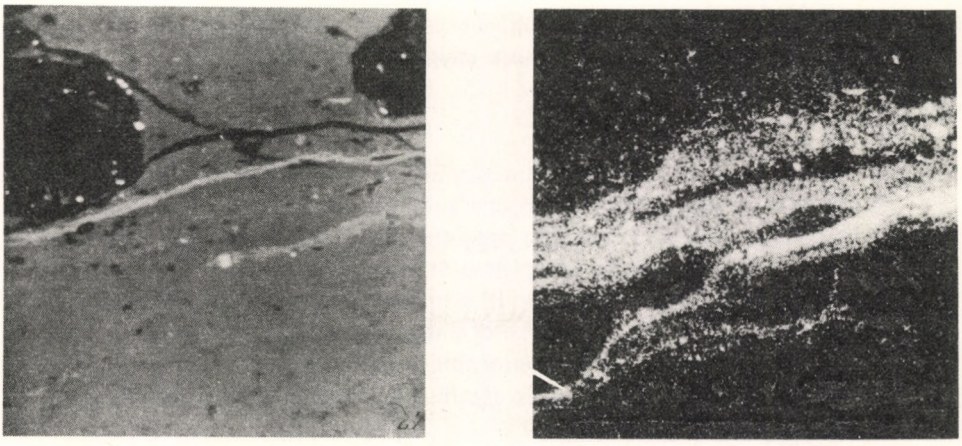


Figure 98. A network of titaniferous veins in the bauxite matrix. Pădurea Craiului, Bihar Mountains, Rumania. Electron microprobe images, 268 ×
(A) electron image; (B) X-ray image of titanium distribution

least 5 cm square, for which the point counting method using a plastic foil is the best suited. The larger the textural elements, the larger is the surface that must be evaluated for the result to fall within the desired range of statistical confidence.

In some earlier work—e.g. Gedeon (1931)—the relative abundances of the matrix and the pisoids, roundgrains and bauxite pebbles embedded in it were stated in percentage weights. This is feasible only if the bauxite can be elutriated so that the components can be separated and then weighed. Elutriation must not result in the disintegration of the textural elements. Bauxites satisfying these criteria, however, are few and far between, and so this method does not lend itself to a general application.

It would in principle be desirable to subject the types of bauxite made up of textural elements of clastic origin to particle size analysis just as one deals with the clastic sediments. Because of the above stated difficulties, the only way of obtaining systematic particle size analyses, as far as the writer can perceive, is the measurement of thin sections or cut rock surfaces. This, however, is so labour-consuming that its broader acceptance is unlikely in the not too remote future.

The bauxite matrix is the only textural element which in itself constitutes a texture. Textures consisting of matrix only are called *homogeneous*. In all other cases, the matrix embeds one or more types of textural elements, giving rise to a *heterogeneous texture*. If the abundance of some textural element exceeds 50%, then the texture as a whole is to be named after it (e.g. pisoidic bauxite). If none of the textural elements attains the abundance of 50%, then all those whose abundance exceeds 25% are to be enumerated in decreasing order of abundance (e.g. pisoidic–ooidic–microclastic bauxite).

The determination and consequent identification of these *composite textural types* is a time-consuming task. Textural elements may combine in such a variety of ways that any attempt to encompass them all would lead to an impenetrable maze of very little usefulness. *A table stating the percentage abundances of textural elements is therefore*

more informative than even the most detailed textural classification. This is why only the most important simple textural types are enumerated below:

(A) *Textural types of authigenic origin*

(1) *Pelitomorphic texture.* It consists predominantly of a pelitomorphic matrix (fig. 70). It may be of syngenetic, epigene and supergene origin.
(2) *Microgranular texture.* It consists predominantly of a microgranular matrix. It may be of syn-, diagenetic or epigene origin.
(3) *Panidiomorphic–granular texture.* It consists predominantly of idiomorphic mineral grains (fig. 99). It may be of diagenetic and epigene origin.
(4) *Pseudoporphyric texture.* A pelitomorphic or microgranular texture embedding large authigenic mineral grains in significant abundance (fig. 100); it may be of diagenetic or epigene origin.
(5) *Ooidic texture.* Consists predominantly of ooids (fig. 101); of diagenetic origin.
(6) *Pisoidic texture.* Consists predominantly of pisoids; of diagenetic origin.
(7) *Macropisoidic texture.* Consists predominantly of macropisoids (colour figure 4); of diagenetic origin.
(8) *Fluidal–collomorphic texture.* A matrix, possibly containing other fine textural elements, arranged in fluidal or rhytmic forms, most often in the spaces between bauxite fragments and pebbles (fig. 102). In such cases it is due to the matrix giving way under the pebbles' pressure, that is, to a mechanical effect. In other cases, it is diagenetic and epigene, a result of a colloidal rearrangement of material. A rare textural type of limited distribution in karst bauxites.
(9) *Pseudobrecciate texture.* The matrix is traversed by a dense network of veins due to gel ageing. The lumps of matrix separated by the veins look at a first glance like the debris making up a bauxite breccia. The good fit between the outlines of adjacent lumps, however, implies an authigenic origin. It may be of diagenetic and epigene origin.
(10) *Nodular texture.* Made up predominantly of nodes of roundish outline or of concretions of irregular form; diagenetic or epigene.

(B) *Textural types of allothigenic–clastic origin*

(11) *Microclastic texture.* Made up predominantly of microfragments of clastic origin, of less than 60 μm size (colour figure 5).
(12) *Arenitic texture.* Made up predominantly of clastic bauxite particles in the 60 to 2000 μm size range (colour figure 6). The term psammitic texture used by some authors is a synonym.
(13) *Roundgrained texture.* Made up predominantly of roundgrains (colour figure 7).
(14) *Conglomerate texture.* Made up predominantly of bauxite pebbles larger than 2 mm (colour figure 8).
(15) *Breccia texture.* Made up predominantly of bauxite fragments in the 2 to 100 mm size range (fig. 103).
(16) *Agglomerate texture.* Made up predominantly of bauxite boulders larger than 10 cm (fig. 104).

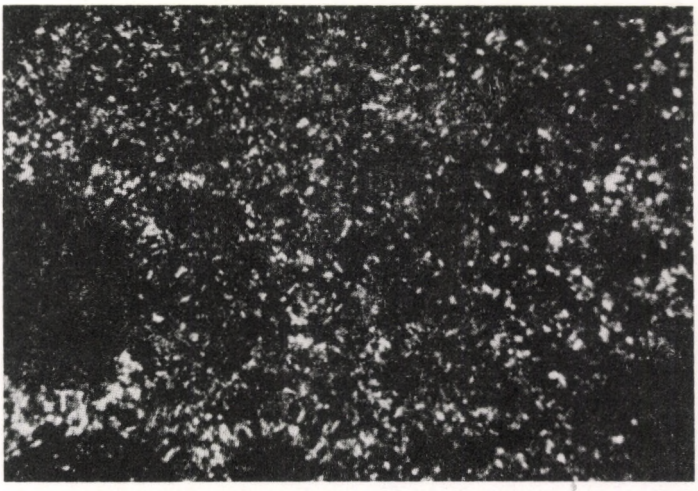


Figure 99. Bauxite of panidiomorphic texture. Kozina, Slovenia, Yugoslavia. Thin section, crossed nicols, 60 ×

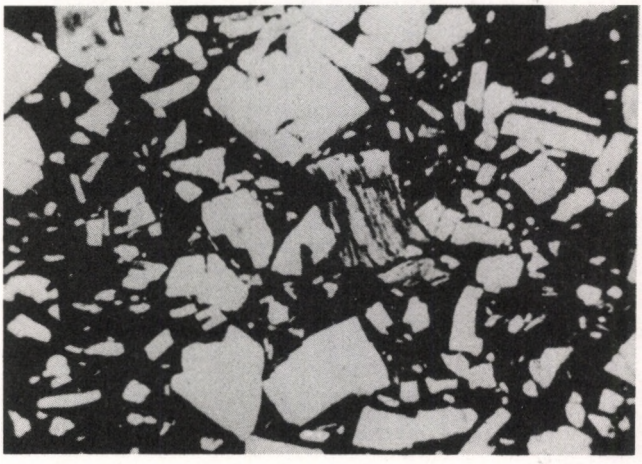


Figure 100. Bauxitic clay of pseudoporphyric texture. Badek, Lika County, Yugoslavia. Thin section, crossed nicols, 20 ×. After Šinkovec (1970)

(C) *Textural types of metamorphic bauxites*

(17) *Granoblastic texture.* Made up of macrocrystalline isometric grains of roughly equal size.
(18) *Porphyroblastic texture.* A panidiomorphic matrix embedding porphyroblasts of larger size.
(19) *Lepidoblastic texture.* Made up predominantly of flaky or platy macrocrystals.

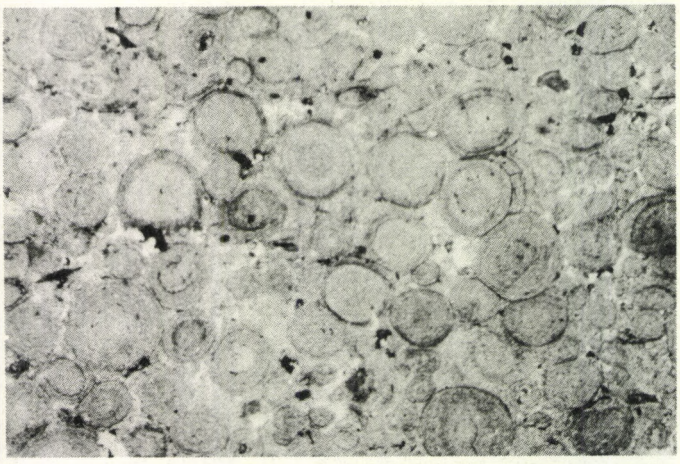


Figure 101. Bauxite of ooidic texture. Nadrt, Slovenia, Yugoslavia. Thin section, parallel nicols, 30×

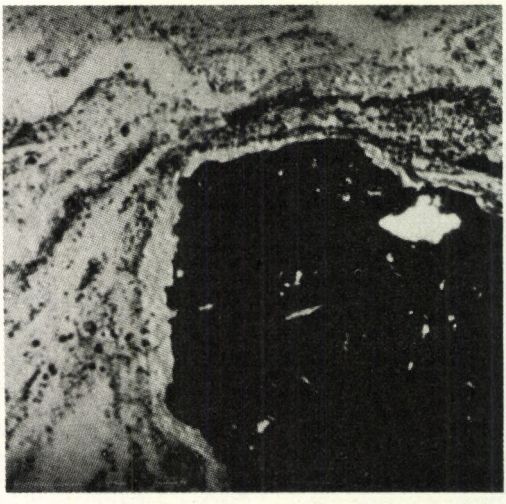


Figure 102. A bauxite of fluidal–collomorphic texture surrounding a clastic bauxite grain. Behbehan, Iran. Electron microprobe, 268×, electron image

Authigenic texture types are made up of predominantly authigenic textural elements. Textures in the authigenic group may be syn- and diagenetic, epigene or supergene in origin. Pre-existing textural elements may be attacked by secondary processes, by pyrite oxidation and iron leaching in particular. In such bauxites, the original texture will be partly or wholly effaced. These typical epi- and supergene textures are predominantly pelitomorphic or collomorphic.

Textures of allothigenic–clastic origin are made up predominantly of textural elements of clastic origin. Like all clastic rocks, they are best categorised by particle size distribution.

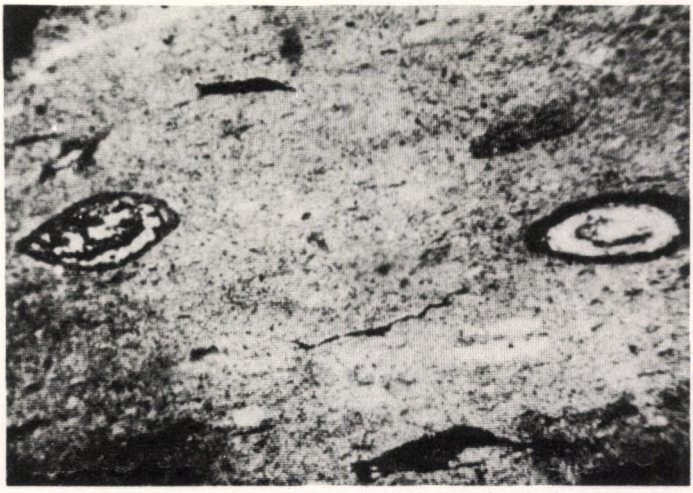


Figure 103. Bauxite of breccia texture. Halimba, Malomvölgy X lens, Hungary. A block of redeposited bauxite in the open-pit face.



Figure 104. Bauxite of agglomerate texture in the top part of the deposit, Mas de Fléchons, Alpilles, France. Photo by Bildgen (1971)

Finally, the author wishes to point out some terms which he feels are misused in a textural context. Terms of hardness (hard, soft or friable) are no textural terms; nor can an 'earthy bauxite' (Valeton 1972) be a textural type. The number and form of pores and cavities contained in a bauxite are not textural features, either. They belong to the category of space filling or packing. This is why, in the writer's opinion, 'vesicular texture' (Valeton 1972) is not a textural type, either. Some cavities in some bauxites may be due to the selective dissolution of certain textural elements. For instance, in certain outcrops, pisoids were dissolved out of the bauxite by a supergene process. This

is a peculiar 'negative texture', in which small spherical vesicles indicate the places occupied previously by the pisoids. This, however, is no separate textural type, either; it is precisely the vesicles that permit identification of the original pisoidic texture.

In describing a bauxite texture, only textural elements may be taken into account, and lithostructural features such as stratification, or physical features such as hardness, compactness or colour, should be disregarded if a tangled-up terminology fraught with internal contradictions is to be avoided.

4.3.3. ABUNDANCE OF TEXTURAL TYPES IN SPACE AND TIME

Using the principles outlined in the foregoing section, the writer has identified the principal textural types and their relative abundances in each bauxite deposit group (Appendix 2). He has determined the textural constitution of some deposit groups in the field; from others, he could procure typical bauxite samples whose textures he has identified at the laboratory. For the rest, he has evaluated the data found in the literature and the textural photographs annexed to papers. Proceeding in this manner, he could establish no relevant data for only 2% of all karst bauxites.

In addition to the 19 simple texture types, karst bauxites feature more than 50 composite ones, too. There are few deposit groups dominated by a single type of texture. Even within a single deposit, several textural types can be distinguished as a rule. The relative abundances of these, however, are widely different. The writer has referred the abundances of textural types to total karst bauxite tonnage. Of these, only the ten most frequent ones are presented below.

(1) Pelitomorphic texture	26%
(2) Ooido–pisoidic texture	10%
(3) Microclastic–arenitic texture	7%
(4) Conglomerato–arenitic texture	5%
(5) Pisoidic texture	3%
(6) Arenitic texture	3%
(7) Arenito–roundgrained texture	3%
(8) Microclastic texture	3%
(9) Microclasto–pelitic texture	2%
(10) Ooidic texture	2%

The abundance of the remaining 60 or so structural types is comparatively great (36%); the ten types enumerated encompass only 64% of total karst bauxite tonnage.

A more meaningful picture can be obtained by classing textural types into genetic groups.

(1) There are five textural types made up exclusively of matrix textures. Their aggregate abundance (32%) is greatest of all textural-type groups. The types of this group are distributed as follows: pelitomorphic textures, 26%; pelitomorphic–microgranular ones, 2%, microgranular, 1%; microclastic–panidiomorphic, 2%, and panidiomorphic, 1%.

(2) There are seven textural types made up of concentric textural elements. Their aggregate abundance is 16%, an overwhelming majority of which (15%) is constituted

by ooidic and pisoidic types. The remaining 1% is made up by macropisoidic, pisoidic–macropisoidic, ooidic–pisoidic and ooidic–pisoidic–macropisoidic textures. The only pure macropisoidic texture so far known is from the Arkalyk deposit group of Kazakhstan. Micro-ooids nowhere attain the abundance required for classing them as independent textural type.

(3) The writer has distinguished five textural types in which the matrix and the concentric textural elements occur in about equal abundance. This group is essentially transitional between Groups 1 and 2; its abundance is a mere 4%. Within it, ooidic–pelitomorphic and pisoidic–pelitomorphic textures are the most abundant. For an example, let us cite the Palaeocene deposit group of Kozina in Slovenia, Yugoslavia, with its ooidic–microgranular and ooidic–panidiomorphic textures: the striking increase in the grain size of its matrix is due to a diagenetic diasporisation.

(4) The aggregate abundance of all other authigenic textures is just 0·4%. This group includes collomorphic textures, which are most abundant in Arkalyk bauxite. Pseudoporphyric textures were identified by Šinkovec (1970) in the Triassic bauxite deposits of Lika County in Yugoslavia, but they are fairly rare even there. Pseudobreccia textures are known so far only in the South Timan bauxite USSR, thanks to Gulyanitsky (1969), but they are subordinate even there. Nodular textures have been observed by the writer in some bauxites of Vietnam and Slovenia, and in the Pepper area of St Elizabeth Plain, Jamaica. The abundance of nodes in the Pepper area bauxite is 20 to 60 wt. %. The other Jamaican bauxites also contain iron-rich nodes, but their abundance is less than 5% almost everywhere.

(5) Seventeen texture types are composed predominantly of clastic textural elements or combinations of such. Such textures have an aggregate abundance of 27% in karst bauxites. Finely clastic textures at 16% are more abundant than coarsely clastic ones (at 11%). In the finely clastic group, microclastic and arenitic textures predominate; those containing roundgrains are much rarer. Within the coarse type, conglomerates are most abundant; breccias are scarcer and agglomerates even more so. The only pure agglomerate texture in karst bauxite is known in the southern part of the French Alpilles. A bauxite agglomerate–conglomerate is fairly frequent in the Gánt deposit group of Hungary. Some large, scattered bauxite boulders are also found in the deposits of Szőc and Malomvölgy, Hungary.

Some types of texture are characterised by an extremely poor sorting of clastic elements (microclastic–arenitic conglomerate, conglomeratico-roundgrained microclastic texture). Both are rare, with an aggregate abundance of just 1%. This poor sorting is an important genetic feature enabling the inference of a peculiar turbidity type of transport.

(6) The writer has identified eight textural types in which the matrix and the clastic textural elements are of about equal abundance. This group is essentially a transition between Groups 1 and 5. Its aggregate abundance is 6%. Of the types in the group, the microclastic–pelitomorphic type is the most abundant.

(7) Twenty textural types combine textural elements of clastic and diagenetic origin. Their aggregate abundance is 9%. Genetically, this is the most remarkable group; it reveals a strong diagenetic rearrangement of material subsequent to a clastic transport. The most abundant types of this group are the ooidic–pisoidic–arenitic (2%) and the pisoidic–roundgrained (1%).

(8) The aggregate abundance of the granoblastic, porphyroblastic and lepidoblastic textures typical of metamorphosed bauxites is 3%.

Figure 105 representing the relative abundances of the above eight groups reveals clearly that the transitional and the genetically inhomogeneous groups are much less abundant than the genetically pure types.

An examination of Appendix 2 further reveals that several of the known textural elements form no independent textures either on their own or in combinations (that is, their abundance does not attain 25% in any bauxite). These include pseudomorphs, micro-ooids, pore fillings, segregations of irregular shape, nests, spheroidal concretions, veins and fissure fillings. Their abundance in bauxite is highly varied and will not be amenable to an evaluation until some fairly detailed quantitative assessments get made.

On the basis of examinations on numerous type samples and of an evaluation of the data in the literature, the writer has established the following order of decreasing abundances for the texture types of lateritic bauxites.

(1) Relict texture (Bushinsky's (1971) 'pseudomorphic' texture).
(2) Collomorphic and collomorphic–pisoidic texture.
(3) Macropisoidic and macropisoidic–pisoidic textures.
(4) Pelitomorphic textures.
(5) Concretionary textures.
(6) Tubular textures (vermicular).

Lateritic bauxites formed *in situ* may contain no clastic textural types in the first place: such types may only be produced by reworking. In comparison with the karst bauxites, the most remarkable difference is that relict textures, totally absent from karst bauxites, are the most abundant. The fair abundance of collomorphic textures and macropisoids, also rare in karst bauxites, is another striking difference. No concretionary structures are known from karst bauxites, either. Pelitomorphic textures are abundant in both, although much more so in karst bauxites. Ooidic and ooidic–pisoidic types are also more abundant in karst bauxites.

In the relative abundances of textural types, then, there is a marked difference between karst and lateritic bauxites, as a sequel to their different processes of formation. The geographic distribution and abundance of textural types is not

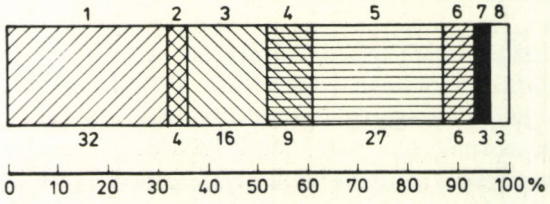


Figure 105. Percentage abundances of the textural types, grouped on a genetic basis

(1) textural types consisting of matrix only; (2) matrix and concentric elements in about equal abundance; (3) textural types composed of concentric elements; (4) concentric and clastic elements in about equal abundance; (5) textural types made up of clastic textural elements; (6) clastic textural elements and matrix in about equal abundance; (7) textural types of metamorphic bauxites; (8) bauxites unevaluated for lack of data

accidental, but a function of genesis. This is also confirmed by the distribution of textural types according to geological age and deposit type.

Young unburied deposits are predominantly pelitomorphic (Solomon Islands; Samar Island; Lifou, Maré in the Loyalty Group; Jamaica, Hispaniola). The bauxites of these deposits are so friable as to yield no clastic material even when redeposited. Instances of diagenetic rearrangement of some intensity (including e.g. the formation of concentric textural elements) are very rare. In deep-lying deposits close to the ground water table, the formation of iron-rich nodes may have begun (e.g. in the Pepper Area, Jamaica). The formation of ooids and pisoids seems to require a diagenesis of some intensity under an overburden, with the bauxite permanently saturated with ground water.

The older bauxite deposits exhibit no essential age-dependent textural features: even the oldest bauxites contain both the clastic and the principal diagenetic types of texture. The only difference is in the matrix: microgranular and panidiomorphic matrices are typical of the Triassic and older deposits above all, whereas the pelitomorphic matrix is typical of the post-Triassic ones. The growth of crystals up to a full crystallisation of the matrix seems to be a very slow process: it is inferred from the above facts to take more than 200 million years. The oldest deposits which still contain bauxites of pelitomorphic texture are found in the Timan Mountains, USSR (Lower Carboniferous) and in Missouri State, US (Middle Carboniferous). Both regions have been totally quiescent structurally: that is, the bauxite has not been exposed to any structural stresses accelerating its crystallisation. In the regions exposed to the strongest structural stresses, on the other hand, even the bauxites of the younger deposits have become microgranular. The acme of this process is attained in the bauxites fully recrystallised by regional metamorphism (the emeries).

As regards the textural relevance of depositological type, the writer has recognised the following relationships. In the most typical *Mediterranen-types* deposits, those of the pure carbonate formation, ooidic, ooidio–pisoidic and pelitomorphic textures are predominant (Yugoslavia, Greece). The Hungarian deposits, on the other hand, are characterised by a predominance of microclastic and arenitic texture types. Also in the Paleozoic bauxites of the Urals and of Central Asia, USSR, clastic types are the most abundant. In the Neogene and younger deposit groups, the pelitomorphic texture is dominant (Jamaica, Haiti etc).

In the bauxites of the *Timan-type deposits*, finely clastic and ooido–pisoidic texture types and those composed of a matrix only are usually found together. The overwhelming majority of these deposit groups being Paleozoic, the matrix is recrystallised, microgranular or panidiomorphic in most places. Only in the Timan Mountains proper are there bauxite types of pelitomorphic texture in any significant abundance.

The *Kazakhstan-type of deposits* is characterised by a predominance of clastic texture types and by the broad granulometry range of the clastic elements. Within one and the same deposit group (or deposit for that matter), all textural types from the microclastic to the conglomerate may occur. In the deposits where there was strong bauxitisation after deposition, macropisoidic, pisoidic and collomorphic textures much resembling those of the lateritic bauxites are also found, for instance, in the deposit groups of Arkalyk and the Western Turgay, USSR, and of the Coastal Plain and the Valley and Ridge Province, US. The 'relicto-clastic' textures described by Pastukhova (1975) are found in those deposits, too.

In *Ariège-type bauxite*, conformably to its *in situ* bauxitisation, it is likewise the pure pisoidic textures that are the most abundant, whereas the underlying, hardly bauxitised clay is predominantly pelitomorphic. Clastic textures are rare in this type: they are confined to the uppermost parts of the deposits which underwent some reworking.

The peculiar mode of formation of the *Tulsk type of deposit* is also reflected by its pelitomorphic–panidiomorphic textures.

The *Salento-type deposits* have no typical textural features of their own; they are made up of redeposited lumps of bauxite of larger or smaller size, in which the texture of the original bauxite is hardly modified if at all. The matrix itself is pelitomorphic.

In summary, any differences of depositological type are tendencies rather than hard and fast rules: there may be substantial differences even between the deposit groups of one and the same deposit type. The joint influence of geological age and overburden pressure seem to have had more control over the definitive textures than the depositiological features.

In the writer's opinion, the study of textural types may turn out to be much more helpful in the understanding of genetic processes than has been believed so far. The broader acceptance of systematic quantitative measurements may in future bring substantial advances in this field.

4.4. LITHOSTRUCTURES OF BAUXITE

As used by the writer, the term encompasses the spatial arrangement of textural types, their orientation, the larger-scale internal inhomogeneities of a deposit, and its larger-scale form elements commensurable with its overall size. Lithostructures are primary or secondary: the first had formed before the lithification of the bauxite; the second ones are subsequent to it. The group of primary lithostructural features includes stratification; allothigenic alien rock pebbles and interbeddings in bauxite; the group of secondary features includes joints, fissures, vesicles, cavities and incrustations at the bedrock contact.

4.4.1. STRATIFICATION, FOLIATION

Many authors consider karst bauxites to be unstratified, and, indeed, regard the absence of stratification as their most conspicuous lithological feature. This, however, is not true: many bauxite deposits are stratified to a greater or lesser extent. The writer's observations and the relevant data in literature have been included in Appendix 2. The following basic types of bauxite stratification can be distinguished.

4.4.1.1. UNSTRATIFIED BAUXITE

Within this group, the writer has identified two types of lithostructural constitution.

(i) *Homogeneous–massive lithostructure.* Bauxites of such structure are homogeneous in all directions, and so are their textures: if they change at all, they do so very slowly and gradually. In the thin sections of such bauxites, individual mineral grains are seen to exhibit no preferred orientation at all; their arrangement is entirely random. Such a

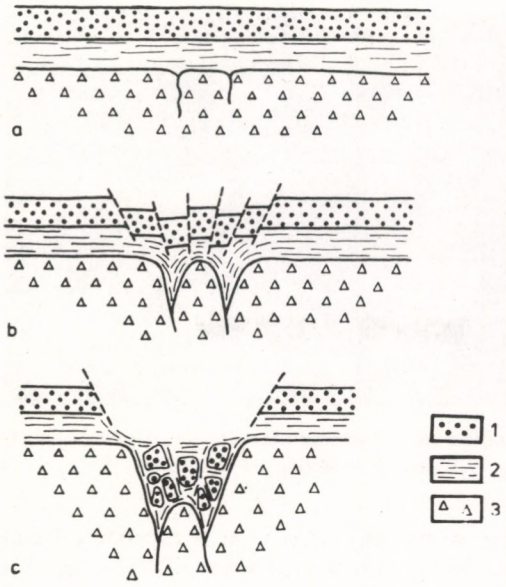


Figure 106. Scheme of evolution of a heterogeneous–chaotic lithostructure according to Combes (1969)
(1) pisoidic bauxite; (2) pelitomorphic bauxitic clay; (3) bedrock dolostone or limestone

structure may be formed by the very slow crystallisation of a gel-like pulp under very quiet conditions. The type includes the huge bauxite deposits of Central America above all, the deposits of the Pacific, and parts of the deposits of Yugoslavia and Greece.

(ii) *Heterogeneous–chaotic lithostructure.* Blocks of a bauxite of a given texture and constitution are embedded in an altogether different bauxite than that from which they are sharply separate. The blocks are as a rule, composed of hard and high-grade bauxite; the embedding bauxite is softer and more clayey. The blocks may be of 0·5 to 10 m^3 size. One of the most remarkable examples of this structure is at Bédarieux in Southern France. It is a secondary structure, engendered by subsequent karstification in the bedrock. Combes (1969) states the process to have been as follows (fig. 106).

Bedrock karstification under an originally stratiform deposit began in those places where permeability was highest (phase a). Bauxitic clay flowed, as it were, into the fissures and depressions thus formed. More rigid, high-grade bauxite slumped down after it, breaking up into blocks (phase b). The further deepening of the sinkhole thus formed gave the entire deposit a chaotic lithostructure (phase c). The writer observed similar lithostructures in some deposits of the Maestrazgo in Spain, of Nézsa in Hungary, and of Pădurea Craiului in Transylvania, Rumania.

4.4.1.2. STRATIFIED BAUXITE

The degree of stratification is highly varied. Its most rudimentary, least conspicuous form involves no beds or bedding planes in the bauxite: it is only the major textural elements that are arranged parallel to the plane of deposition (which would be the

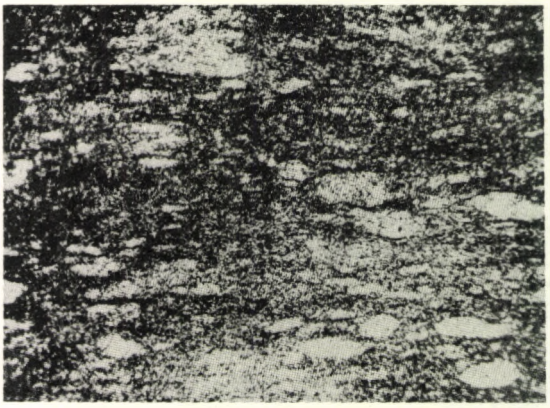


Figure 107. Stratification due to a parallel arrangement of textural elements. Gun District, Honan Province, China. Thin section. After Schüller (1957)

bedding plane, if there were any). This type is especially frequent in the Chinese bauxites (fig. 107) and in the primary deposits of Vietnam. A vague stratification, passing gradually into the underlying unstratified bauxite, is observed in the top few centimetres of the bauxite deposits of Central Greece.

Nicolas (1970) also observed arrangements of bauxite pebbles parallel to the plane of deposition to be frequent in the deposits of Brignoles, France. In the Halimba Basin, Hungary, bauxite in many places contains scattered dolostone detritus and pebbles. Most of these are arranged in strings parallel to the plane of deposition, but some are scattered entirely at random. Dolostone detritus is aggregated in places into layers 5 to 20 cm thick. There may be several such layers in one profile. Taken together, these layers trace phantom horizons of stratification all over the bauxite deposit. The writer has himself observed this sort of stratification in Halimba II mine.

In the next stage, there are still no bedding planes, but the texture of bauxite exhibits gradual changes perpendicular to the planes of deposition. For instance, at several points in the Gánt bauxite deposit group of Hungary, bauxite types of pelitomorphic, microclastic, arenitic and conglomerate texture alternate with no sharp limits between them (colour figure 9) forming a rudimentary stratification.

In other cases, there is a graded arrangement of clastic particles, with a gradual upward decrease of particle sizes, the finest particles at the top of one cycle overlain by the coarsest texture at the base of the next. These cycles may be repeated several times. The feature is observed in several places near the deposit tops in the deposit groups of Gánt, Halimba and Nézsa, Hungary ('graded bedding').

Most Jamaican bauxite deposits are unstratified. Traces of stratification can be seen only in the deposits of the plains and of the larger valleys. In one of the open bauxite pits near Meyersville in the St Elisabeth Plain, graded bedding was observed by the author in the central part of the deposit. The largest bauxite pebbles at the base of the cycle were 2 to 3 cm in size. Particles became gradually finer towards the top. The bauxite was unstratified both above and below the stratified section.

Bushinsky (1971) observed frequent graded bedding in the Devonian bauxite deposits of the Northern Urals (fig. 108). Abdullayev (1970) stated it to be frequent in the

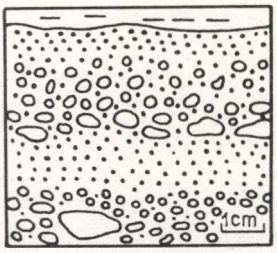


Figure 108. Graded bedding in a Devonian bauxite, Vagran deposit, Northern Urals, USSR. After Bushinsky (1971)

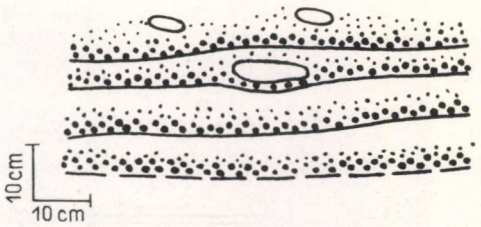


Figure 109. Repeated cycles of graded bedding in an Upper Cretaceous bauxite deposit. Mine de Tourves near Brignoles, France

Paleozoic bauxites of the Central Asian bauxite region, USSR as well. Kharin (1969) observed it in the Devonian bauxite of Berdsk–Obuchovsk, USSR. The writer observed a repetition of graded beds in the underground Mine de Tourves of the Brignoles deposit group. Cycles of 5 to 10 cm thickness, growing finer upward, occupy a zone about 3 m high. Some of the larger bauxite pebbles are seen there to have penetrated, under their own weight, the still soft, pelitomorphic bauxite underlying them (fig. 109). Graded bedding was confined to a rather restricted area: growing indistinct, it passed sidewise into unstratified bauxite. Bauxite is unstratified in the rest of this mine; elsewhere in the Brignoles deposit group, however, especially in the Mazaugues deposit, graded bedding was observed in several places by Nicolas (1968, 1970). In the Villeveyrac Basin, the writer observed graded bedding with a cycle thickness of 10 cm in a pisoidic–roundgrained bauxite. The pisoids exhibited signs of transport. Such phenomena are also frequent in the deposits of the Western Turgay, USSR (colour figure 10).

It is not justified to call a form of stratification a true bedding unless the beds are separated by well-defined bedding planes. True bedding is less frequent in karst bauxites than vague stratification. It occurs most frequently in the Devonian deposits of the Northern Urals. The beds are usually 0·5 to 4 cm thick, parallel, conspicuously distinct in colour and particle size. This type of stratification is called *banded bedding* (colour figure 11). There is also some finely bedded bauxite made up of layers 1 to 3 mm thick: it is more rare, though (colour figure 12). The Devonian deposits at Berdsk–Obuchovsk USSR, are also well-bedded in their majority. In them, *microstratified bauxite* made up of microscopic layerlets alternates with bauxite of banded bedding (Nagorsky 1958). The top parts of the Cambrian bauxite deposits at Boksonsk are also microstratified, with an alternation of layerlets of lighter and darker colour a tenth of a millimetre thick (Orlova 1958).

Combes (1973) observed types of finely stratified bauxite in those parts of the Bédarieux deposit where subsequent karstification was weak. The layers, 1 to 3 mm thick, are made visible by their lighter and darker shades. The layers underlying a bauxite pebble are seen in places to be slightly depressed by the weight of the pebble. At one place in the deposit called La Braunhe, Combes found a well-bedded alternation of layers, each a few millimetres thick, of clayey bauxite and finely clastic dolostone (fig. 110).

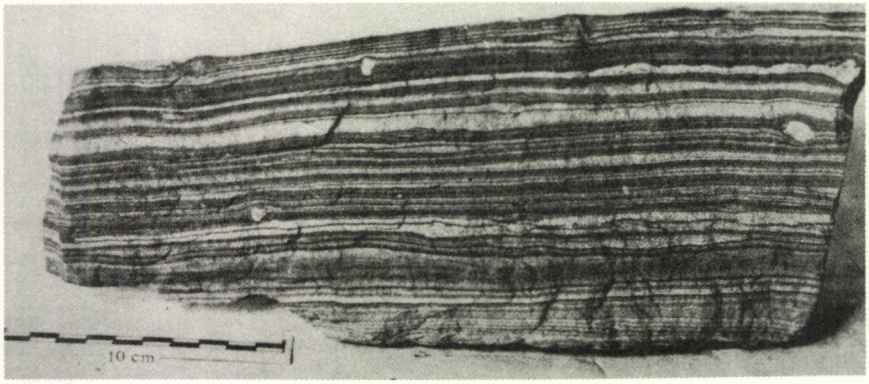


Figure 110. Alternating layers of clayey bauxite (dark) and of dolostone sand (light), in places engulfing dolostone pebbles of several millimetres in size. La Braunhe deposit, Bédarieux, France. Photo by Combes (1973)

In Slovenia near Hrast, the writer has observed a finely stratified bauxite section about 30 cm thick in the central part of an unstratified deposit. Subsequent beds differ in their colour (colour figure 13). This well-stratified section is made up of bauxite-like clay, grading into bauxite both upward and downward. Denishevich (1958) discovered, using a microscope, a very fine microstratification in the bauxite of Raho, Carpatho-Ukraine, USSR.

Cross-bedding is even rarer in bauxite. One case, observed by Bushinsky (1971) in a bauxite of arenitic texture in the deposit Krasnaya Shapochka of the Northern Urals (fig. 111), was ascribed by him to fluvial transport. The oblique beds include angles of some 30° with the lowest beds; they dipeast, indicating a sediment inflow from the west. Nicolas and Lecolle (1966) observed a cross-bedding ('stratification oblique') in beds made up of broken and worn ooids at the bottom of the Mazaugues deposit near Brignoles, France. The cross-bedded material grades upward into unstratified bauxite.

Some exceptional cases of *desiccation craks* have also been discovered in bauxite. Bushinsky (1971) described such forms in bauxite of banded stratification in certain Paleozoic deposits from the Northern Urals and from Central Asia, USSR, where beds of bauxite of pelitomorphic and arenitic texture alternate. Vertical fissures closed at the bottom can be observed in places in the pelitomorphic layers (fig. 112). In ground

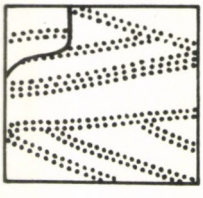


Figure 111. Cross-bedded Devonian bauxite. Vagran deposit, Northern Urals, USSR. After Bushinsky (1971)

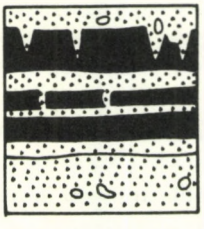


Figure 112. Desiccation cracks in stratified bauxite. Devonian. Vagran deposit, Northern Urals, USSR. After Bushinsky (1971)

Table 8. Age distribution of stratification in bauxite.

Geological age	Unstratified	Partly vaguely stratified	Partly well-stratified	No data	Altogether
Neogene and Quaternary	21	1	–	–	22
Paleogene	27	8	–	3	38
Mesozoic	57	30	1	8	96
Paleozoic	5	12	11	5	33
Altogether	110	51	12	16	189

plan, the fissures form a polygonal pattern. Maximum crack length is 5 cm, maximum depth 3 cm, and maximum gape at the top, 4 mm. The cracks are filled with the arenitic bauxite of the layer above. Desiccation cracks indicate the occasional drying-out of the basin of accumulation. The extreme rarity of this phenomenon in bauxites implies that basins of accumulation tended to dry out very seldom if at all in the course of bauxite deposition.

The data in Appendix 2 reveal among others that it is the bauxites with clastic textures that tend to be stratified, whereas pelitomorphic textures tend to be unstratified. Sedimentary clays, on the other hand, are usually stratified, and often very well at that, despite their similar particle size. The lack of stratification, then, must be attributed to peculiar conditions of sedimentation rather than to the fine particle size of the bauxites.

An even more important fact is recognised on inspecting the geological age distribution of stratification (table 8).

All the Neogene and Quaternary deposit groups are unstratified. So are the great majority of Paleogene ones, although some are weakly stratified. Among the Mesozoic deposit groups, a significant number includes weakly stratified bauxite in places. Finally, in the Paleozoic deposit groups, well-stratified and weakly stratified bauxites are predominant, and unstratified ones are subordinate.

Stratification is a primary lithologic feature. Structural stress or metamorphism may engender foliation or schistosity, which is easy to distinguish from primary stratification. The data in table 8, then, indicate a gradual change in the conditions of accumulation of bauxite sediment over the geological ages. It is accordingly wrong to view the older and younger karst bauxites in exactly the same light, at least as far as their conditions of accumulation are concerned. It is the Paleozoic deposit groups above all that differ in this respect from the younger bauxites.

Supergene processes may partly or entirely efface the original bedding of a bauxite. Cases in point have been observed by the writer in samples of the bauxites, reworked in the Tertiary, of Vietnam. The original grey Paleozoic bauxite is weakly stratified. The finely clastic bauxite particles are seen in thin sections to be arranged subparallel to the stratification. The reworked deposits are made up of boulders and pebbles of the primary bauxite. These were oxidised to a yellow or red colour on their outsides by supergene alteration, to a depth of 5 to 20 mm, with a sharp reaction interface. Wherever this surface had passed, it changed not only the colour but also the texture of the original bauxite, effacing its stratification. Such strong supergene weathering is connected with the surface oxidation of reduced bauxites above all.

Foliation (slaty cleavage) is a secondary structural feature, frequent in those bauxite deposits exposed to strong structural stresses or to some degree of metamorphism. It tends to be most conspicuous in clayey bauxite and bauxitic clay, as a result of the recrystallisation of the clay minerals in planes perpendicular to the vector of compression.

The deposit groups where foliation in bauxite is known have been specially marked in Appendix 2. Foliation in Chinese bauxites is the product of structural stress and overburden load almost exclusively, without the benefit of any significant metamorphism. The same holds for a small bauxite deposit near Mazin in Yugoslavia, where a strong structural stress imparted a very strong slaty cleavage to a clayey bauxite and bauxitic clay.

4.4.2. PEBBLES AND GRAVELS OF CARBONATE ROCKS IN BAUXITE

4.4.2.1. LIMESTONE AND DOLOSTONE PEBBLES

More or less worn limestone and dolostone pebbles are frequent at the bases of Mediterranean-type deposits, including the Hungarian ones. Their size varies from a few millimetres to half a metre.

In places, the pebbles consist of quite fresh, unaltered rock. Much more often, however, they are friable, weathered, pulverulent inside. They are frequently impregnated by greyish-black, finely dispersed manganese minerals, or indeed, coated with a black manganiferous film. The writer has observed such films in Greece near Itea, in France, on the rims of the deposit worked in the mine called Monplaisir in the Villeveyrac group, and at the bottom of the bauxite deposits of Le Revest, near Toulon. In the Ariège, the dolostone pebbles are coated by a thin pyritic–mercasitic–haematitic crust (Combes 1969). In places, the pebbles were fully dissolved, with only the empty cavities left behind. The present writer attributes pulverulency in the pebbles and their removal to epi- and supergene leaching.

In the deposits of Monti Maggiore, Monti Matese and Monti d'Ocre in Italy, and in the Itea area in Greece, the writer has observed the number of limestone pebbles in the bauxite to grow towards the deposit margins. Pebble numbers and sizes increase as the deposit thins out. As a result, the bauxite is gradually replaced by a limestone breccia of bauxitic cement. Its thickness is less than one metre as a rule; it seldom attains two metres. Scattered limestone pebbles are most abundant in the deposits of Milovići near Velimlje in Crna Gora (Montenegro). In the small open pits there, most of them worked out, large numbers of hard white limestone pebbles of 10 to 30 cm size were well visible all over the deposit. The most frequent pebble size range was from 5 to 20 cm.

Pebbles of limestone and dolostone may be abundant enough to form intercalated gravels in places in the bauxite. The large stratiform deposit of the Halimba Basin in Hungary also contains such intercalated gravels, besides scattered pebbles (fig. 113). The gravels consist largely of dolostone pebbles; limestone pebbles are less abundant. The pebbles of 0·5 to 10 cm size are embedded in a matrix made up of dolomite flour and clayey bauxite. Their most frequent size class is 1 to 5 cm. They are but slightly

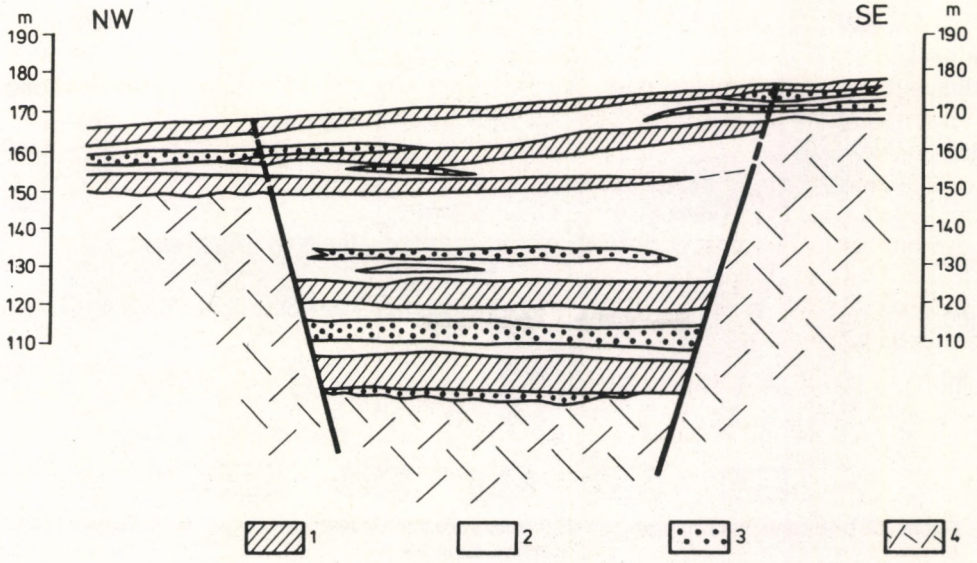


Figure 113. Intercalations of dolostone and limestone gravels in the Halimba Basin deposit, Hungary
(1) bauxite; (2) bauxite with scattered dolostone and limestone pebbles; (3) intercalations of dolostone and limestone gravels; (4) dolomite, Upper Triassic

worn. Some dolostone pebbles have undergone subsequent calcitisation: these are of a concentric-zoned texture. The gravels contain a very few pebbles of a few centimetres size of a dark red, hard, iron-rich bauxite. The pebble-to-matrix ratio varies widely. In places, the matrix is so sparse as to be hardly sufficient to hold the pebbles together. There is a copious hard matrix elsewhere, constituting a conglomerate with the pebbles.

Such intercalated gravels are most frequent on the southern and southwestern rims of the deposit group. Their thickness exceeds 10 m in places; they nevertheless cannot as a rule be traced for more than 100 to 200 m horizontally; 300 m is their maximum extent (fig. 114). In the central part of the deposit—in the Halimba III underground mine—their thickness, as observed by the writer in the roadways, is already reduced to a few centimetres; they very seldom attain 50 cm.

In addition to these intercalations, scattered pebbles of dolostone and, less abundantly, of limestone are frequent all over the stratiform Halimba deposit. They are entirely ungraded, with coarse pebbles next to the finest detritic material. Their abundance also varies capriciously. On an average over the Halimba III mine, the aggregate abundance of pebbles is 1 to 3%. It tends to be lower in the north than in the south and at the centre than on the peripheries. Near the cover, the pebbles are accompanied by finely dispersed authigenic dolomite grains, forming dolomitic bauxites and bauxitic dolostones. This varied dolomitic contamination of the Halimba bauxite is a phenomenon unparalleled in any other deposit group in Hungary or abroad.

The bauxite deposit called Vrace in Yugoslavia is known to contain intercalations of a red limestone conglomerate (Šinkovec 1970). One to eight metres thick, these are of

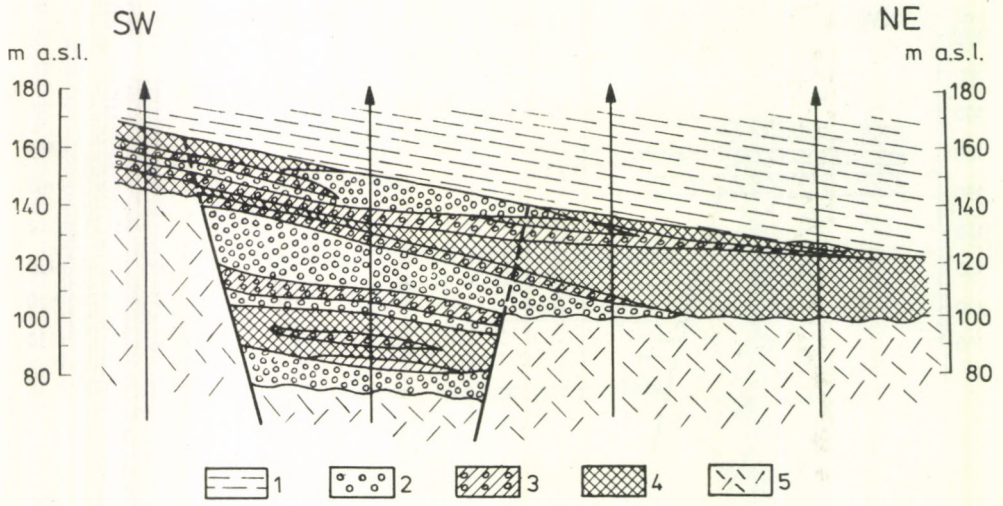


Figure 114. Intercalations of dolostone and limestone gravels in the southwestern part of the Halimba Basin deposit, Hungary

(1) Upper Cretaceous cover; (2) intercalated gravels of dolostone and limestone; (3) bauxite with scattered dolostone and limestone pebbles; (4) bauxite, clayey bauxite and bauxitic clay; (5) Hauptdolomite and Dachsteinkalk, Upper Triassic

laterally tapering, lenticular form (fig. 54). They tend to contain bauxite pebbles, broken and worn ooids and pisoids, especially in their bottom parts. An intercalation rich in dolostone detritus, 1 to 5 m thick, was found by Combes (1971, 1973) in the central part of a bauxite deposit called Uston (Bédarieux, France). In this layer, hardly worn dolostone pebbles are embedded in a fine dolomite sand. The bed, overlain in erosional unconformity by a violet, grey and black clay 1 to 3 m thick, is followed in places by another layer of dolostone detritus. Combes discovered in this intercalation one block of dolomitic limestone of $6 \times 4 \times 1$ m. This is a veritable *olistolite* derived from some bedrock prominence immediately to the south of the deposit. According to Combes, it travelled a distance of a few hundred metres at most to its present site, sliding on the wet clay. Remarkably enough, the bauxite of the deposit contains no scattered dolostone pebbles. The above mentioned intercalation wedges out rather abruptly: it does not traverse the entire bauxite deposit (fig. 115).

Özlü (1977) observed an intercalation of limestone breccia, 5 m thick on average, in the Upper Cretaceous bauxite deposits near Kizilgedik, Taurus Range, Turkey. The lumps of limestone, of 1 to 5 cm size, slightly worn, constitute graded beds in several places, embedded in a red bauxitic matrix. There are two similar layers of limestone breccia at the bottom and at the top of each deposit (fig. 116).

In Jamaica, as far as the writer could observe, there are limestone pebbles of 2 to 20 cm size in the top parts of the deposits, scattered in places, more abundant in others. They are derived from nearby prominences of limestone, having travelled in the taluses of those. They are confined to a zone half a metre thick below the surface, and are almost entirely absent from the deposits' interiors. It is only in a bauxite quarry on the St Elisabeth Plain, at the eastern feet of the Sta Cruz Mts, that the writer could observe in the upper part of the deposit an intercalation 0·5 to 1 m thick containg limestone

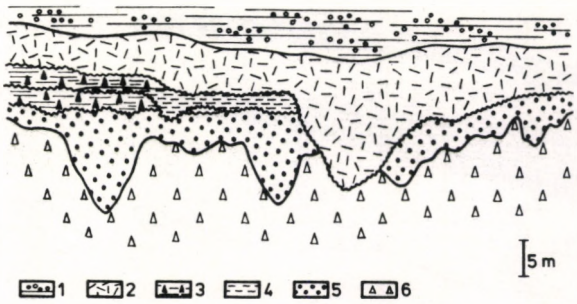


Figure 115. An intercalation of dolostone detrius and clay in bauxite. Uston deposit, Bédarieux, France. After Combes (1973)

(1) clays and conglomerates of the cover, Campanian; (2) Upper horizon of bauxite deposit; (3) intercalations of dolostone detritus; (4) intercalation of grey clay; (5) lower horizon of bauxite deposit; (6) dolostone, Jurassic

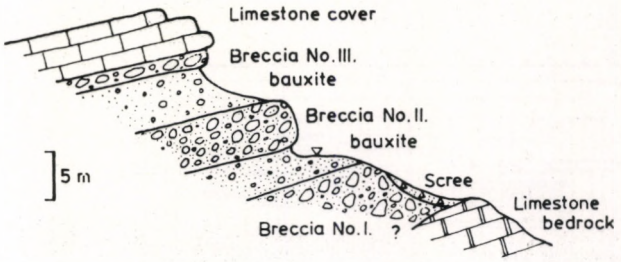


Figure 116. Profile of Kizilgedik bauxite deposit, Turkey. After Özlü (1977)

pebbles of 1 to 10 cm in size, slightly worn and embedded in a bauxitic matrix. The intercalation wedged out within distances of 20 to 50 m. The remarkable thing about this is that the writer could not, even by the most painstaking examination, detect any difference in colour or texture between the bauxite above and below the intercalation.

4.4.2.2. INTERCALATIONS OF AUTHIGENIC LIMESTONE

Such intercalations are known exclusively from Mediterranean and Ariège-type deposits. They were formed in the course of brief inundations which temporarily interrupted the accumulation of bauxite with a phase of carbonate sedimentation.

Limestone intercalations are frequent in greenish-greyish bauxite near the top of the Devonian deposits in the Northern Urals. The limestones are organogenic, comparatively rich in organics and in marine fauna, Amphipores in particular (Peive 1947; Gladkovsky 1948). A lens of bituminous limestone intercalated in red bauxite was described by Dementev and Rodchenko (1966) from the Krasnaya Shapochka deposit of this group (fig. 117).

The top third of the bauxite deposit near Dréveneuse in Switzerland also contains a limestone intercalation 0·3 to 0·5 m thick (Badoux and De Weisse 1959). In Montenegro (Crna Gora), in the Kosijerovo and Petrovići areas, Yugoslavia some

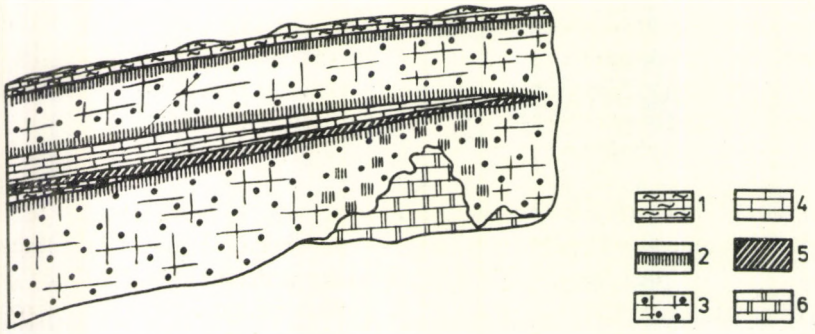


Figure 117. Intercalation of a bituminous limestone in Middle Devonian bauxite. Krasnaya Shapochka deposit, Northen Urals, USSR. After Dementev and Rodchenko (1966)

(1) dark grey limestone with Amphipores; (2) grey pyritic bauxite; (3) red bauxite; (4) grey bituminous limestone; (5) greyish-black bauxite; (6) pale pink reef limestone

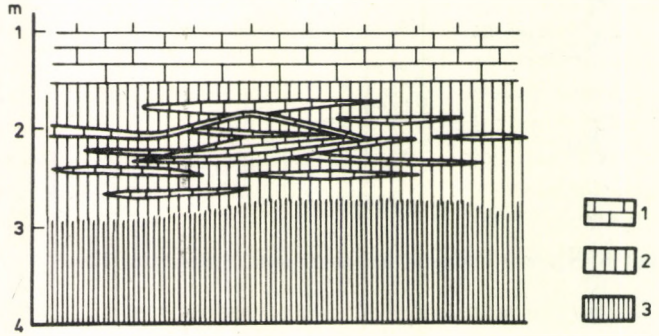


Figure 118. Limestone interbeddings in the top part of a bauxite deposit. Mine de Tourves (underground working), Brignoles, France

(1) dark grey limestone; (2) pale pink bauxite; (3) rust-red bauxite

deposits contain intercalations 0·2 to 0·5 m thick of marine limestone (Burić 1966). In one of the roadways of the underground Mine de Tourves near Brignoles, the writer observed the profile presented in figure 118: near the top of the deposit, the cover limestone of marine facies is anticipated by thin limestone intercalations. There is a sharp limit between the bauxite and the limestones everywhere; also, the bauxite, brick-red elsewhere, is much lighter in the zone where the intercalations are.

4.4.3. OTHER INTERCALATIONS AND LUMPS OF ALIEN ROCK

In the deposits of Kazakhstan type, there are frequent pebbles of alien rock and also clayey, carbonacous and sandy intercalations. In the Timan-type deposits, intercalations of coal are the most frequent. These were described in the depositological chapter. Such features are very rare in the Mediterranean- and Ariège-type deposits: those that exist are largely confined to the bases of the deposits.

The number of deposit groups where alien matter attains a significant abundance is fairly small. In the bottom parts of the Cambrian deposits of Boksonsk (USSR), lumps and pebbles of alumosilicate rocks are fairly frequent (Orlova 1958). In the bottom half of the small bauxite deposit near Markušovce in Czechoslovakia, there are numerous rounded pebbles of shale and more seldom of chlorite schicht (Konta and Kužvart 1962).

In the deposits transitional from the Mediterranean to the Kazakhstan type (marked Med > Kaz in Appendix 1), pebbles of alien rocks are even more abundant; they in fact form beds of gravel in a few cases. For example, some deposits of the Haţeg group in Rumania contain thin intercalations of clayey sandstone, sandstone and microconglomerate (Papiu *et al* 1970).

4.4.4. JOINTS, FISSURES, SLICKENSIDES

There are no joints at all in the young bauxites of friable, earthy or loamy consistency. A precondition for jointing is a certain compaction (mechanical strength) of the bauxite; most Paleogene, Mesozoic and Paleozoic bauxites satisfy that condition. In the undeformed bauxite deposits of structural quiescent regions, joints are few and far between; such fissure planes (diaclases) as exist are to be attributed to the diagenetic compaction of the rock.

Subsequent bedrock karstification triggered slumps and slides in numerous deposits. These produced slickensides with striae indicating the direction of motion. They are distinguished from fault planes by their shorter length and by the fact that they invariably cluster about depressions in the bedrock surface.

In the structurally undeformed Hungarian deposit groups, joints are found especially in the vicinity of faults. In structurally deformed regions, joints are much more abundant, spaced 10 to 20 cm apart in the bauxite. The principal sets of joints include angles of abot 120°. In Greece, in the Anthimos open pit near Distomon, the writer has seen parallel joints spaced about 5 to 10 cm apart in a very hard bauxite. In the deposit called Malia Zeza of the structurally even more deformed (folded and overthrust) Eleusis–Mandra bauxite region Greece, the hard bauxite is traversed by so many joints that, on winning, it falls apart into small particles. In places in the underground workings, these particles of a few mm size flowed, as it were, into the roadways, hampering both the extraction of the bauxite and roadway maintenance. This is, however, a fairly rare phenomenon in karst bauxites.

4.4.5. INCRUSTATIONS AND SEGREGATIONS AT THE BEDROCK SURFACE

In many a bauxite deposit of Mediterranean type, there is a crust a few millimetres to several centimetres thick at the bedrock contact. Its thickness (its presence or absence even) is rather capricious, even within one and the same deposit. Its composition is also highly varied.

Iron-rich incrustations are the most frequent. They are very widespread and well accessible to study in the worked-out open pits of the Gánt deposit group of Hungary. Their usual range of thickness is from 2 to 4 cm, attaining 10 cm in a few places. They

are dark red or dark violet (colour figure 14). Their limit towards the underlying dolostone is not clear-cut: the dolostone is impregnated and dyed violet to a depth of 10 to 40 cm. Their limit towards the bauxite, on the other hand, is sharp. On top of the crust, there are in places clusters of upward-projecting crystals. The principal mineral of the crust is haematite; it is accompanied by smaller amounts of goethite.

The Turonian deposits of the Kiona Mountains in Greece are often underlain by 3 to 10 cm of a liver-brown hard ferric crust composed largely of goethite. Ferruginous crusts of this type are also frequent under the deposits of Yugoslavia, Italy and France, but their usual thickness is but a few millimetres.

In places, the ferruginous crust is underlain by a black *manganiferous crust* several centimetre thick. Between the two crusts there are white nests of gibbsite, sometimes also an uninterrupted gibbsite layer. This triple layering was especially conspicuous under the bauxitic clay deposit on Pajtikatető near Cserszegtomaj, Hungary (Bárdossy 1961b).

The black crust usually consists of lithiophorite or todorokite or a mixture of the two minerals. In places, there is only a black manganiferous crust of this sort above the bedrock, e.g. in the Eleusis–Mandra bauxite region. In the open pit at Anthimos near Distomon, Greece, this black crust is several centimetres thick; it is of a porous and friable consistency. The manganese minerals impregnating the bedrock gave it a black colour to a depth of 10 to 30 cm. Manganiferous crusts are also frequent under the Yugoslavian and French deposits, but less so under the Jamaican ones.

Under some of the deposits of the Turonian bauxite deposits of the Kiona range, Greece, the writer has observed in places white, fibrous *segregations of gibbsite*, with both the iron- and the manganese-rich crust absent. Here and there, the gibbsite contains nests of a few centimetres in size of white halloysite of soapy consistency. More seldom, there are segregations of a yellow, porous and friable goethite above the gibbsite.

In places, the limestone and dolostone pebbles embedded in the bottom parts of the bauxite deposits are also coated by a manganiferous or ferro-manganiferous crust.

Numerous deposits contain *segregations of calcite* at the bedrock contact. Those underlying the deposit called Cabanes of the Bédarieux group, France, are among the finest. The calcite crystals attain the length of 10 cm, forming half-spherical or reniform radiated aggregates to a thickness of 30 cm in places (Combes 1973). At the bedrock contact in the Spinazzola deposit group, Italy, the writer observed secondary calcite crystals of several centimetres in length at several places. In the Halimba–Malomvölgy deposits, Hungary, the writer has found calcite crystals of several centimetres in size at the contact with the dolostone bedrock, forming radiated, spherical and rosette-like crystal aggregates. There are projecting calcite crystals of 1 to 3 cm in size in minor depressions of the dolostone surface.

Minerals other than these ubiquitous crusts and segregations are exceptional at the bedrock contact. Combes (1969), for example, observed a very hard brownish-black *pyritic-gypsum crust* on top of the bedrock dolostone near Péreille in the French Pyrenees. Under the deposit Phtheria-1 in the Kiona Mountains, Greece, a crust 1 to 4 mm thick of turquoise-blue *takovite* was found on top of the bedrock limestone by Bárdossy and Mack (1967). The writer found loose green segregations of *malachite* at the bedrock contact of the Nazda deposit, Vlasenica, Yugoslavia.

The writer holds the majority of these crusts and segregations to be epi- and supergene. His ideas concerning their origin are detailed in Chapter 5.4.3.

4.5. PHYSICAL PROPERTIES OF BAUXITE

4.5.1. THE FRACTURE OF BAUXITE

No objective method for the quantification of bauxite fracture is available. This is why literature abounds in terms not well enough defined referring to bauxite fracture. After a study of large numbers of bauxite samples, the writer has distinguished six types of fracture, as follows:

(i) *Uneven fracture:* the fracture surface is irregular.

(ii) *Smooth fracture:* the fracture surface is entirely smooth or exhibits minor irregularities only.

(iii) *Rhombohedral fracture:* the bauxite, on being struck, falls apart into parallelepipeds of larger or smaller size; the fracture surfaces are plane and smooth.

(iv) *Conchoidal fracture:* the fracture surface is made up of shallow form elements of 0·5 to 10 cm^2 surface area, similar in relief to the interior of a bivalve shell.

(v) *Foliate fracture:* the bauxite, on being struck, falls apart into thin sheets ('foils').

(vi) *Earthy fracture:* the fracture surface is granulate, earthy, resembling the fracture of a soil.

The abundances of these types of fracture are plotted in Appendix 2. The most widespread and most frequent type of fracture is the uneven type, followed by earthy fracture. Most of the bauxites in the huge Caribbean deposits are of earthy fracture. The bauxites of uneven fracture making up most of the deposits of the Transdanubian Hills in Hungary and those of Vlasenica in Yugoslavia are accompanied by less abundant bauxites of earthy fracture. Most Kazakhstan-type bauxites are of uneven, less often of earthy fracture. Most of the bauxite in the Timan-type deposits is of uneven fracture. The writer has classed the bauxites of the Pacific islands as earthy-fracture types, although they are so loose and friable that they cannot be said to possess a fracture at all in the strict sense of the term.

Conchoidal fracture is also fairly widespread, but usually of minor abundance within the deposits. Of the Hungarian deposit groups, bauxite of conchoidal fracture is most abundant at Nyirád and Alsópere. It occurs in some such bauxite in the French, Yugoslav and Greek deposits, and also in the Devonian deposits of the Urals.

The remaining three types of fracture are of subordinate abundance as compared with the above three. Bauxite of rhombohedral fracture is most abundant in the Eleusis–Mandra region of Greece. There are some fine examples of rhombohedral fracture in the Unterlaussa deposit, Austria, in the Paleocene deposits of Kozina, Slovenia, in the Malm bauxite horizon of the Parnassos and Kiona Mountains, Greece, and in the deposit group of Akseki–Seydisehir, Turkey.

Bauxites of smooth fracture are most abundant in the Triassic bauxite deposits of Slovenia, Yugoslavia, and in the Mesozoic bauxites of Greece. Of the Hungarian deposits, those at Nagyharsány contain the most abundant smooth-fractured bauxite.

The rarest type of fracture is the foliate one. Essentially confined to the bauxitic and bauxite-like clays accompanying bauxite, it is most frequent in the Ariège and in the Languedoc, France.

After a comparison of large numbers of bauxite samples, the writer has gained the conviction that bauxite fracture is a function of texture and hardness above all; influenced, however, also by packing and by the structural stresses to which the bauxite had been exposed.

Wherever a bauxite is made up of different textural elements of different hardness, its fracture tends to be uneven. The writer has observed conchoidal fracture primarily in bauxites of pelitomorphic, microgranular and collomorphic texture, provided they were also hard enough. Colour figure 15 shows a pelitomorphic bauxite sample of typical conchoidal fracture. Bauxites of such texture tend to be of earthy fracture if they are friable and porous. Earthy fracture is confined to friable bauxites; it is impossible in hard massive bauxite. Smooth fractures will develop in those harder bauxites which are heterogeneous as to texture, provided that the textural elements making them up are of equal hardness. If such a bauxite is broken, the pisoids, bauxite pebbles etc also break in two, and a fairly smooth fracture surface results.

Rhombohedral fracture has been observed by the writer exclusively in hard bauxite of strong structural deformation, traversed by systems of joints. Such bauxite tends to fracture along the joints and parallel to them. Bauxite of rhombohedral fracture tends to be pelitomorphic or microgranular (that is, homogeneous as to texture). Foliate fracture is a consequence of five stratifications in bauxitic or bauxite-like clay.

4.5.2. THE TOUCH OF BAUXITE

No means are available for quantifying this petrographic feature, either. On the basis of this own experience, the writer has distinguished six types:

(i) very rough to the touch,
(ii) rough,
(iii) slightly rough,
(iv) smooth,
(v) greasy-slick,
(vi) earthy.

Most bauxites are rough; bauxites slightly rough and earthy to the touch are less frequent. The other three groups are much rarer.

Touch also is a function of bauxite texture. The more inhomogeneous the texture, the rougher is the bauxite. Bauxites of pelitomorphic and collomorphic texture are usually smooth or slightly rough. Mineral composition also plays a role: a greasy-slick touch is largely confined in the writer's experience to samples of bauxitic clay and clayey bauxite, as a result of their higher clay minerals content. Hardness and compaction also play a role; the earthy touch is confined exclusively to friable-porous bauxite types of low hardness.

4.5.3. THE HARDNESS OF BAUXITE

By hardness, the writer means the resistance of bauxite to external mechanical efforts towards changing its form. The hardness of bauxite, then, is not simply the weighted average hardness of its rock-forming minerals, but a composite petrographic feature which, in addition to the hardness of the constituent minerals, depends also on texture, particle size and compactness.

In the absence of modern devices for the measurement of hardness, the writer was obliged to resort to the time-honoured Mohs scale applying it to several hundreds of samples, the writer has defined the following hardness groups for bauxite in terms of the Mohs scale:

Mohs hardness	Bauxite-hardness term
above 8·0	high-hardness
4·0 to 8·0	very hard
3·0 to 4·0	hard
2·0 to 3·0	medium-hard
1·0 to 2·0	low-hardness
0·5 to 1·0	friable-soft
0·1 to 0·5	unconsolidated

Appendix 2 also includes successions frequency of bauxites of different hardness. Weighting these by bauxite tonnage, the writer has found the following frequency distribution over world total karst bauxite tonnage:

above 8·0	1%
4·0 to 8·0	23%
3·0 to 4·0	22%
2·0 to 3·0	15%
1·0 to 2·0	20%
0·5 to 1·0	18%
0·1 to 0·5	1%
	100%

All rocks of bauxitic composition having hardness greater than 8 are emeries. Their extreme hardness is the result of metamorphism. Their abundance is insignificant, and so is that of unconsolidated bauxite. The latter is confined exclusively to the bauxite regions of the Pacific islands. Colour figure 16 is a photo of such an unconsolidated bauxite. The lumps of such a bauxite can easily be crumbled by hand. Remarkably enough, all the other classes of hardness are of about equal abundance. The result of a weighted averaging by geological periods is presented in table 9.

In a general way, the older the bauxite, the greater its hardness; the more time there was for the processes of hardening. Some of the processes in question are very slow, and may have taken millions of years.

There are, however, also manifold time-independent factors of hardening, as proved by the fact that, even within one and the same geological period, bauxite types covering the broadest possible range of hardness may be found. One of the time-independent factors is *overburden pressure* (load). The reason why Miocene, Pliocene and Quaternary bauxites are so uncompacted and friable is because they have never been exposed to overburden pressure, which would have resulted in a gradual compaction and hardening.

Another factor is *structural stress*. The more undeformed the bauxite, the lower its hardness as a rule. To quantify this feature, the author has subdivided the bauxite deposits of the Carboniferous and of the Upper Cretaceous into classes of structural

Table 9. Percentage distribution of bauxite hardness over the geological periods (Mohs degrees).

Geological age	8·0	4·0–8·0	3·0–4·0	2·0–3·0	1·0–2·0	0·5–1·0	1·0–0·5
Quaternary	–	–	–	–	–	–	100
Miocene–Pliocene	–	–	–	4	36	60	–
Oligocene	–	40	34	13	8	5	–
Paleocene	–	5	11	29	38	17	–
Upper Cretaceous	–	26	25	28	18	3	–
Lower Cretaceous	<1	14	34	26	15	11	–
Jurassic	–	63	34	2	1	–	–
Triassic	<1	40	53	3	2	1	–
Permian	22	53	25	–	–	–	–
Carboniferous	<1	54	35	5	6	–	–
Devonian	3	38	43	16	–	–	–
Silurian	–	–	–	–	–	–	–
Ordovician	40	60	–	–	–	–	–
Cambrian	–	60	40	–	–	–	–

Table 10. Hardness of Carboniferous bauxites as a function of structural deformation (percentages) (Mohs degrees).

Degree of deformation	>8·0	4·0–8·0	3·0–4·0	2·0–3·0	1·0–2·0
Undeformed deposits	–	13	15	46	26
Moderately deformed deposits	–	60	36	–	4
Strongly deformed deposits	5	40	55	–	–

Table 11. Hardness of Cretaceous bauxites as a function of structural deformation (percentages) (Mohs degrees).

Degree of deformation	>8·0	4·0–8·0	3·0–4·0	2·0–3·0	1·0–2·0	0·5–1·0
Undeformed deposits	–	–	19	22	40	19
Weakly deformed deposits	–	–	19	49	30	2
Strongly deformed deposits	–	48	30	17	5	–
Very strongly deformed deposits	–	17	56	27	–	–

deformation, and has calculated tonnage-weighted hardness distribution for them (tables 10 and 11).

Bauxites in the structurally undeformed Carboniferous deposits (e.g. in the Timan Mountains bauxite region) are predominantly of medium and low hardness, despite their old age. The medium-deformed bauxite deposits of China consist of very hard and hard bauxites; low-hardness bauxites are confined to the outcrops. In the strongly folded Turkestan Range (USSR), bauxites are very hard; there are even emeries of extreme hardness, produced by dynamic metamorphism.

In the structurally undeformed Cretaceous deposits at Chadobetsk, along the Yenisei River and on the Turgay Plateau, USSR low-hardness bauxite is most

abundant, and there is even a fair quantity of friable bauxite. In the slightly deformed Cretaceous deposits, including the Upper Cretaceous deposit groups of Hungary, medium-hard bauxite is the most abundant, the quantity of friable bauxite being insignificant. In the strongly deformed Cretaceous deposits, very hard bauxite is the most abundant (e.g. in Yugoslavia and Greece). Average hardness tends to be somewhat less in the very strongly dislocated deposits (e.g. those of Austria and Central Italy), but these contain no bauxite of less than medium hardness at all.

The situation is also similar in the other geological periods. This is especially striking in the Paleocene and Eocene deposits of the Transdanubian Hills, Hungary, which are weakly deformed in the main. Most of the bauxite contained in them is of low hardness: in fact, most of them also contain some friable bauxite. The bauxites of the deposits of identical age in Yugoslavia and Greece, on the other hand, are hard or medium-hard; they are accompanied by no friable bauxite at all. These latter deposits, however, are of middling or strong deformation.

Hardness is influenced considerably also by *mineral constitution*. This is why hardness in bauxite may vary substantially even within one and the same deposit. Of the rock-forming minerals of bauxite, gibbsite and boehmite are of about the same Mohs hardness (2·5 to 4). Shifts in the relative abundances of these two minerals do not therefore affect the hardness of the bauxite. Diaspore, on the other hand, is of 6·5 to 7 Mohs hardness. Accordingly, even within a given deposit, the bauxites rich in diaspore are invariably harder than the boehmitic ones. Haematite and goethite, of Mohs hardness 5 to 6, also augment the hardness of bauxite. The writer has observed in many of the bauxite deposits in Hungary that iron-rich bauxites are invariably harder than bauxities of average or low iron content in their vicinity. According to Konta and Kužvart (1962), the hardness of Markušovce bauxite (Czechoslovakia) is the greater, the more haematite it contains. Clay minerals decrese the hardness of bauxite, so that, within a given deposit, bauxitic clay and clayey bauxite are invariably softer than high grade ore.

Hardness is influenced also by the *compactness of bauxite*. For a given mineral constitution, bauxite of greater porosity is less hard than massive bauxite. The secondary processes that augment a bauxite's porosity by leading out some of its components may substantially also reduce its hardness.

In the majority of lateritic bauxites, hardness is a variable property. They are soft as a rule when in the ground, and harden a great deal after winning. This is what enables their use as a building stone in India (whence their name: 'later' means brick in Latin).

In karst bauxites, this phenomenon is less pronounced. Medium-hard, hard and very hard bauxites are as hard in the ground as after drying, and these make up 74% of all karst bauxites. A certain hardening will be observed in low-hardness and friable bauxites. The bauxites of the Transdanubian Hills are softer on average by 0·5 to 1·0 Mohs units when wet than after drying. (The data in Appendix 2 invariably refer to dried-out bauxite.)

Jamaican bauxite undergoes some hardening on drying. This is why, in earlier centuries, it could be used as a mortar or plaster in a mixture with slaked lime. The walls so plastered, of a characteristic red colour, are called 'Spanish walls' by the local inhabitants (Zans 1954). The friable, porous bauxite of Haiti hardens into a tough brick-like material if dried on the surface (Goldich and Bergquist 1948). In the

Dominican Republic, in the deposit group of the Sierra de Bahoruco, the outcrops of a bauxite that is quite friable otherwise consist of hardened blocks of several cubic metres in size (Goldich and Bergquist 1947). Also in the bauxite regions of the Southeastern US, bauxite in the ground is very soft: it hardens somewhat on drying. The large bauxite blocks of the outcrops are invariably much harder than freshly won bauxite (Bergquist and Overstreet 1965).

This hardening process is explained by a recrystallisation triggered by drying. It is the iron minerals above all, especially alumogoethite, which recrystallise on drying, thus contributing to the hardness of the bauxite.

4.5.4. THE COMPACTNESS OF BAUXITE

The compactness of bauxite can be quantified in terms of its *porosity*, using the formula

$$P = \frac{d - D}{d} \cdot 100\%,$$

where P is porosity, as a percentage,
d is specific gravity and
D is bulk density.

In the writer's experience it is also preferable to determine at the same time the apparent porosity

$$P_1 = \frac{W - V}{W - S} \cdot 100\%$$

of the bauxite,
where P_1 is apparent porosity, as a percentage,
W is the weight of water-saturated bauxite,
S is the same, as measured in water and
V is the dry weight of the lump of bauxite.

The expression was introduced by Allen (1952, 1955) to characterise bauxites and refractory clays. It expresses the abundance of open pores and micropores (those that can be filled with water). In the writer's experience, porosity in most bauxites equals apparent porosity or differs from it but very slightly.

Bauxite porosity may be widely different within even a given deposit. It is thus unjustified to calculate averages if the data base available is not rather large.

Tercinier (1971) established an average porosity of 75 to 76% for Lifou Island bauxite, Loyalty Groups. Zans (1954) stated 50% on average for the Jamaican bauxites; Goldich and Bergquist (1948) gave 44 to 48% on average for the Haitian ones. The porosity of most Hungarian bauxites is in the 15 to 30% range. The overwhelming majority of Triassic and Paleozoic bauxites, on the other hand, has less than 10% porosity. Within a given bauxite region, a bauxite tends to be the less porous, the older it is. For instance, in the Slovenian bauxite region, the following porosities have been determined by the writer and Komoróczy:

Oligocene bauxite horizon	10 to 34%
Paleocene bauxite horizon	8 to 12%
Turonian bauxite horizon	8 to 10%
Malm bauxite horizon	2 to 5%
Middle Triassic bauxite horizon	1·5 to 3%.

Within a given deposit, high-grade bauxite tends to be the most porous, more so than clayey bauxite and bauxitic clay. In the deposits of the Turonian bauxite horizon of Central Greece, Nia (1968) found the highest-porosity bauxites to be diasporic, those of medium porosity to be boehmitic, and those of lowest porosity to be kaolinitic–boehmitic. He attributed this pehomenon to diagenetic leaching and diasporisation, an issue to which we shall return in the chapter on mineralogy.

High-porosity bauxites are also of fair *permeability*. Jamaican bauxite, for example, is of excellent permeability. The deposits swallow the abundant tropical showers in a very short time, draining them towards the bedrock. This is why Jamaican bauxite is so resistant to erosion, being stable over decades in steep road cuts, without slumping or being washed down by the rains. According to Anderson (1971), this property can easily be destroyed by compaction. Tamping by hand, widely used in local agriculture, is quite sufficient to do away with it. A loss of permeability as a result of compaction under the tracks and wheels of shovels and heavy off-highway trucks was observed in the open pit workings.

In the petrographic aspect, the nature of *space-filling* is even more essential than the percentage of porosity. A bauxite may be made up of pores and cavities which are highly varied as to size and shape, regardless of whether percentage porosity is high or low. The writer has distinguished the following categories of space-filling in karst bauxites.

(i) *Microporous bauxite*. Pore size is less than 0·05 mm. These so-called micropores cannot be distinguished on a broken bauxite surface even with the help of a hand lens.

(ii) *Porous bauxite*. Pore size is from 0·05 to 0·5 mm; the pores can be distinguished using a hand lens.

(iii) *Vacuolar bauxite*. The vacuoles are of 0·5 to 5·0 mm diameter, of roughly round or oval cross section.

(iv) *Tubulo-vacuolar bauxite*. The bauxite contains straight or curved tubular cavities of 1 to 2 mm diameter and 1 to 2 cm length.

(v) *Vesicular bauxite*. The cavities are larger than 5 mm; they are isometric, more or less.

(vi) *Tubulo-vesicular bauxite*. It contains tubular cavities of more than 5 mm diameter and several centimetres length, straight or curved.

(vii) *Spongy bauxite*. Vacuoles and pores are very close-spaced, giving the bauxite a spongy appearance.

It takes much less effort to identify these seven basic types than to determine percentage porosity. The writer estimated their relative abundance for each deposit group (as stated in a separate column in Appendix 2). Their tonnage-weighted distribution over the total world karst bauxite tonnage, as derived from these figures, is as follows:

Microporous bauxite	56%
Porous bauxite	36%
Vacuolar bauxite	7%
Tubulo-vacuolar bauxite Vesicular bauxite Tubulo-vesicular bauxite Spongy bauxite	1%
	100%

The overwhelming majority of all karst bauxites, then, is microporous and porous. Vacuolar bauxite is fairly rare; the other four types are rarities. In lateritic bauxites, on the other hand, tubulo-vacuolar and tubulo-vesicular space-filling are rather frequent, the difference being attributed by the writer to the stronger processes of secondary leaching and retexturation acting upon the lateritic bauxites.

Tubulo-vacuolar and tubulo-vesicular bauxite types are of evanescent frequency in the Mediterranean type deposits. They are comparatively most frequent in the Kazakhstan type (e.g. in the deposit groups at Arkalyk, along the Yenisei River and in the Southern Urals, USSR). The tubular vacuoles and cavities are subvertical as a rule. Some are due to dissolution by downward-seeping ground water; others correspond to one-time roots.

Spongy bauxite is the product of secondary processes, of the dissolution of limestone and dolostone pebbles contained in the original bauxite above all. Sometimes they exhibit cellular forms. This is how the spongy Devonian bauxite sample from the Northern Urals shown in colour figure 17 came to exist. A pumice-like spongy bauxite due to supergene alteration was described by Knyazeva (1966) from the Ivdel deposit of the Northern Urals. This bauxite likewise resulted from the dissolution of the limestone lumps originally contained in it. There are also some rare spots of spongy bauxite in the deposit groups of the Southeastern US.

The cavities of vesicular bauxite are also due to secondary processes. Most of them mark the places of dissolved largish pebbles. The writer has discovered some typical vesicular bauxites in the Turonian bauxite horizon in the vicinity of Nadrt and in the Paleocene bauxite horizon of the Kozina area, both in Slovenia. They are of subordinate abundance even in those deposits, however. There is also a small quantity of vesicular bauxite in the Turonian bauxite horizon of the Parnassos and Kiona Mountains, Greece.

Secondary dissolution along faults may give rise to remarkably large caves in bauxite. Kiss (1955) e.g. found a cave several metres long and half a metre wide on average, with a crust of calcite and aragonite on its walls, along a fault plane in the northeastern part of Hosszúharasztos deposit, Gánt, Hungary.

Cavities even larger than these are produced by the epigene dissolution of a karstic bedrock if the bauxite and the beds covering it are of sufficient strength not to collapse. In one of the deposits of the Turonian bauxite horizon in the Kiona Mountains, Greece, the writer has observed a cylindrical cavern of about 20 m diameter and 30 m depth. Its upper end was flush with the cover top; its bottom end formed a sinkhole-like depression in the bedrock. The hard bauxite making up the deposit formed the vertical walls of the cavern, and its blocks fallen down out of what had become the cavity lay around at the bottom of the sinkhole.

Table 12. Age distribution of packing categories over the geological periods (percentages).

Geological period	Microporous	Porous	Vacuolar	Other
Quaternary	–	100	–	–
Miocene–Pliocene	–	82	18	–
Oligocene	77	22	1	–
Paleocene–Eocene	34	56	8	2
Upper Cretaceous	73	22	4	1
Lower Cretaceous	68	30	2	<1
Jurassic	99	<1	<1	–
Triassic	99	1	–	–
Permian	99	–	<1	<1
Carboniferous	93	6	1	–
Devonian	80	9	8	3
Ordovician	100	–	–	–
Cambrian	100	–	–	–

Vacuolar bauxite is also a product of secondary dissolution. Besides the dissolution of smallish pebbles of limestone and dolostone, it may be due also to the oxidation plus leaching of pyrite and marcasite grains. The latter was the case, as observed by the writer, with the grey bauxite types of the Turonian bauxite horizon in the Parnasssos and Kiona Mountains, Greece. In other cases, pisoids and roundgrains are dissolved, leaving spherical vesicles in the rock. Bauxites of this sort are frequent in the deposit groups of the Southeastern United States (colour figure 18). There are also vacuolar bauxites of this sort in the upper parts of the Ariège deposits in France.

The writer has calculated the frequency distribution of the space-filling categories so defined over the geological ages (table 12), as it is closely related to geological age. Miocene and younger bauxites are porous and vacuolar; Oligocene to Lower Cretaceous ones are alternately porous and microporous; their older ones are predominantly microporous. This state of facts may be due in addition to the time factor, also to contributions by overburden pressure and structural deformation.

The Hungarian bauxites fit well into this scheme: they are porous or microporous in alternation. In most deposit groups, small quantities of vacuolar bauxite are also found. The oldest bauxite, that of Nagyharsány, which is also the most structurally deformed, is microporous in its entirety.

Insight into the finest, submicroscopic particulars of space-filling in bauxites can be gained by means of the *scanning electron microscope*. In recent years, a large number of scanning electron micrographs of bauxites have been prepared at the laboratories of ALUTERV–FKI (Budapest), the engineering and research subsidiary of the Hungarian Aluminium Corporation (Bárdossy *et al* 1978). The most important results were as follows.

In karst bauxites, stacky, uniformly microporous, uniformly compact and webby space-filling types can be distinguished. In the lateritic bauxites examined so far, the stacky type has not been encountered. In addition to the other three types mentioned, lateritic bauxites also feature collomorphous spongy and framework type space-filling.

All the young karst bauxites with no overburden exhibit stacky space-filling. The loosest stacky space-filling was observed in bauxite from Maré Island in the Pacific

(Loyalty Group) (fig. 119). The crystallites of that bauxite form stacks of 20 to 300 μm size which are joined by column-like aggregates. It is this stacky–columnar microstructure which is responsible for the extreme porosity of these bauxites. The crystallites of Jamaican bauxites form stacks of 2 to 20 μm size, with micropores of a few micrometres in the interstices. These bauxites are also very highly porous (fig. 120).

Overburden pressure will compact but not efface the original stacky space-filling. This is how the Hungarian bauxites can still exhibit stacky space-filling; the aggregates, however, are only of 1 to 10 μm size, and the micropores in the interstices are also smaller (fig.121). The crystallites within the stacks are close-packed. Most French bauxites have similar space-filling features. In all these bauxites, the crystallites are isometric: regular crystal faces or edges are rarely observed. The stacks together constitute a comparatively strong three-dimensional grid which can stand up to the pressure of up to 400 m of overburden.

If overburden pressure presists over a very long time—e.g. in the case of Paleozoic bauxites — or is assisted by structural stress, then compaction advances to where stacky space-filling is changed into the uniformly microporous or uniformly compact type.The porosity of the bauxite is reduced hand in hand with this process. A fine example of uniformly compact space-filling is the Triassic bauxite of Podlipa, Slovenia, shown in figure 122. The micrograph reveals clearly how the clusters have disappeared, the space now being uniformly filled by the crystallites. In the micrographs of 3000 and 10 000 diameters magnification, a larger authigenic kaolinite crystal rising above the background is visible. The slightly curved kaolinite lamellae ('books') of the particle can be distinguished quite well.

Strong structural stress initiates the recrystallisation of the original crystallites; a crystalline-webby space-filling being the result. Tap-Na bauxite from the northern frontier of Vietnam is a case in point (fig. 123). The original crystallites are recrystallised into particles of 5 to 50 μm size. Most crystallites are bounded by well-developed crystal faces and edges. Together, they form a very densely packed web. Some of these bauxites were exposed in the Neogene, eroded and redeposited at the feet of the slopes as a secondary debris. Superficial oxidation and leaching increased the porosity of the debris by dissolving numerous micropassages (fig. 124). The flakes of 0·2 to 10 μm size seen in the micrograph are newly formed secondary minerals. Surface alteration, then, can greatly change the space-filling of older bauxites.

In summary, most karst bauxites differ in space-filling from lateritic bauxites. Within the karst bauxite group, space-filling and particle morphology reflect the secondary effects undergone by a bauxite or the lack of such. A systematic study of space-filling is very helpful in understanding the origin of bauxites.

4.5.5. THE COLOUR OF BAUXITE

There are many who believe that bauxites are red by definition, although in fact their colours cover a broad range (white, pale yellow, ochre, yellowish brown, brown pink, brick-red, rust-red, chocolate-brown, violet, green, pale grey and even black). Bauxites of uniform colour are accompanied frequently by multicoloured—motley or variegated—bauxites.

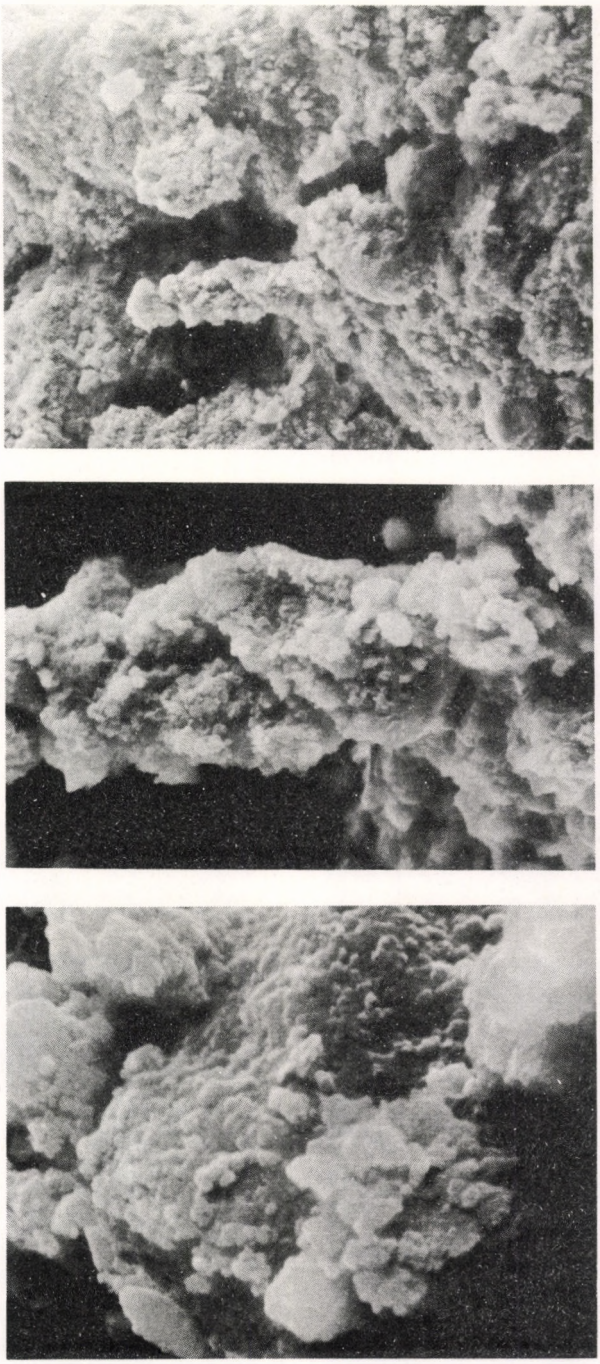


Figure 119. Scanning electron micrograph of recent bauxite. Maré Island, Loyalty Group, Western Pacific
(1) 300 × ; (2) 1000 × ; (3) 10 000 ×

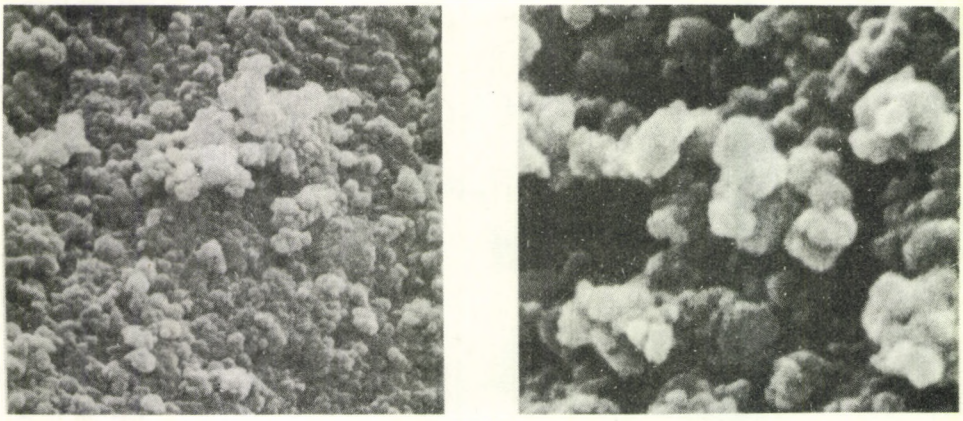
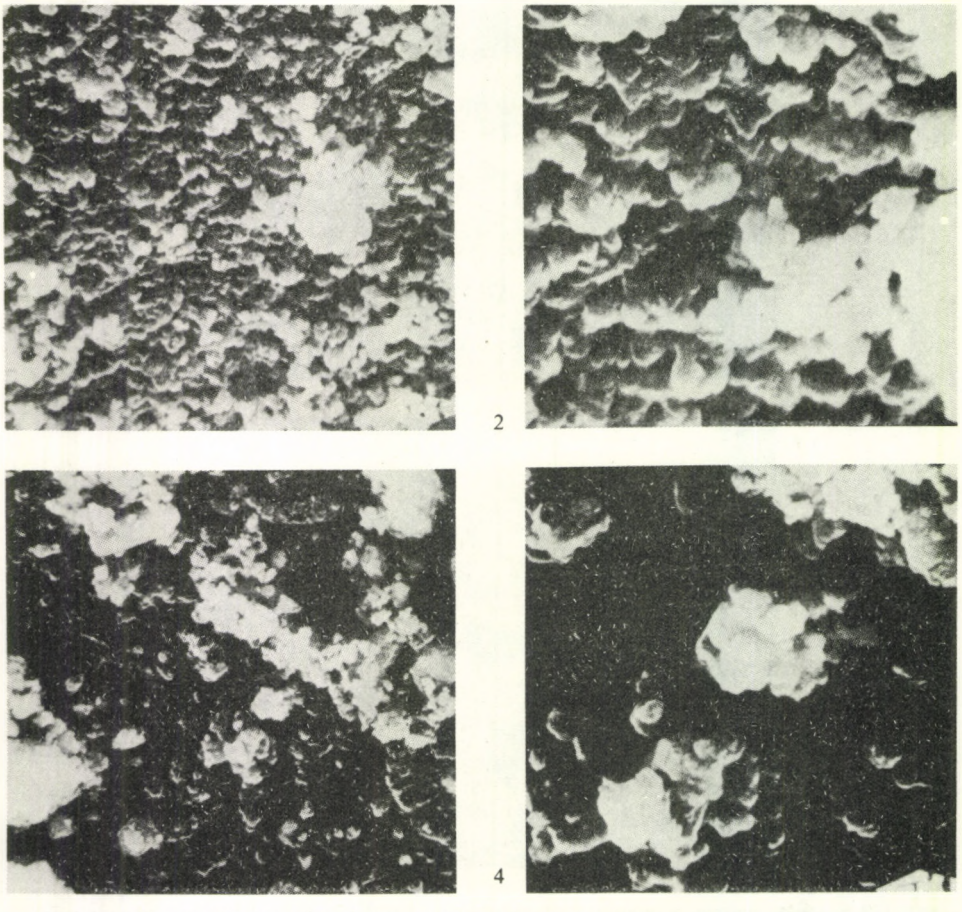


Figure 120. Scanning electron micrographs of typical Jamaican bauxites, South Manchester Plateau
(1) 10 000 × ; (2) 30 000 ×



1 2 3 4

Figure 121. Scanning electron micrographs of typical Hungarian bauxites
(1) red bauxite, Dűlt-Nyires VII lens, Nyirád, 4800 × ; (2) *idem*, 16 000 × ; (3) red bauxite, Malomvölgy XI lens, Halimba, 4800 × ; (4) *idem*, 16 000 ×



Figure 122. Scanning electron micrograph of a massive Middle Triassic bauxite. Podlipa, Slovenia
(1) 1000×; (2) 3000×; (3) 10 000×

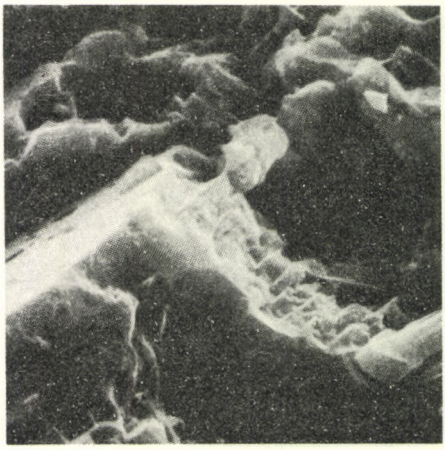


Figure 123. A crystalline-webbed packing of Permian bauxite. Scanning electron micrograph. Tap-Na, Vietnam (3000 ×)

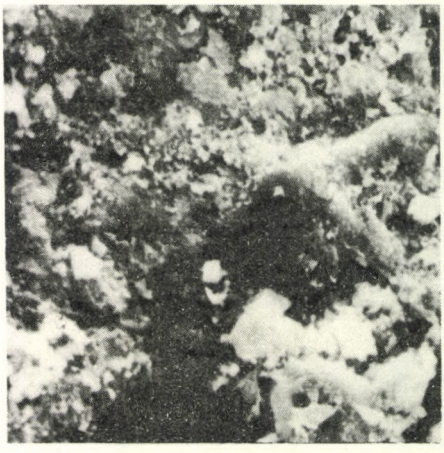


Figure 124. Packing characterised by secondary leaching of redeposited Permian bauxite. Scanning electron micrograph. Tap-Na, Vietnam (3000 ×)

The aluminium and titanium minerals of bauxite are white or colourless, and so are kaolinite and halloysite. The colour of a bauxite is therefore determined by its iron oxides and hydroxides, its iron-containing clay minerals and organic matter, and to a less extent also by its manganese minerals. Haematite gives a red, pink or violet colour, goethite a yellow or yellowish-brown one. Maghemite dyes a bauxite a fallow brown, ferrichamosites a dark brown. Finely dispersed pyrite and marcasite give a grey colour. Siderite gives a light grey tint. Green is due to chlorites, grey and black to organic matter and to manganese oxides and hydroxides. The greater the content of a chromophorous mineral in the bauxite, the darker the resulting shade. In fine

dispersion, even 0·5 to 1% of a chromophorous mineral may give a characteristic colour.

Red, violet, pink, yellow, yellowish brown and multicoloured bauxites indicate oxidative conditions. White bauxite indicates a strong leaching of iron, most often as a result of reduction. A weakly reductive medium is indicated by dark brown, ferrichamositic bauxites; a stronger one by green, grey and black bauxites. The colour of bauxite, then, is indicative not only of its mineralogical composition, but also of the diverse geochemical processes it was exposed to, tracing, as it were, their spatial course.

4.5.5.1. DISTRIBUTION AND ABUNDANCE OF DIFFERENT COLOURS IN BAUXITE

Colour abundances in karst bauxites as stated in Appendix 2 are derived from the field observations of the writer, from the samples collected and from the data he has found in literature. Using these data, the writer has calculated tonnage-weighted percentage abundances for the principal colour groups over world total karst bauxite tonnage, with the following results.

Red, pink, violet	53%
Grey	17%
Multicoloured	11%
Yellow, yellowish-brown	9%
Green, greyish-green	5%
Dark brown	2%
White	2%
Black	1%
Total:	100%

It is doubtless that the red and related colours are most frequent, but grey and multicoloured bauxites are not negligible, either. The subordinate abundances of dark brown, white, and black bauxites indicate unusual circumstances of their origin.

The writer has calculated the distribution of these colours also over the geological periods. In table 13, the colours are grouped so that those indicating oxidative environments are on the left, those indicating reductive environments on the right, and the white and dark brown colours indicating transitional conditions in between.

Colour abundance exhibits certain regularities as a function of geological age. Red and violet bauxites are most abundant in the Quaternary and in the Ordovician. They are fairly abundant also in the other periods, except in the Carboniferous and in the Permian. In these latter, grey and green bauxites are the most frequent. These two colours become increasingly rare in the Mesozoic, to disappear altogether within the Tertiary. Dark brown and black bauxites are also most frequent in the Permian. From the Jurassic on, dark brown bauxites occur in just a few tenths of a percent, and they are entirely absent from the Tertiary, apart from some reworked deposits. Even in these latter, the original bauxite is Permian and it is only the redeposition that took place in the Tertiary. The abundance of yellow and multicoloured bauxites is subordinate in the Paleozoic deposits: it increases throughout the Mesozoic, and is greatest in the Upper Cretaceous–Oligocene deposits. In the Miocene and Pliocene deposits, their

Table 13. Percentage age distribution of bauxite colours over the geological periods.

Geological period	Red, violet	Yellow	Multi-coloured	White	Dark brown	Green	Grey	Black
Quaternary	100	–	–	–	–	–	–	–
Miocene–Pliocene	76	10	9	4	<1	<1	<1	–
Oligocene	57	3	32	–	–	<1	7	–
Paleocene–Eocene	48	18	22	5	–	1	5	<1
Upper Cretaceous	60	13	18	2	<1	1	5	–
Lower Cretaceous	72	7	13	1	<1	2	4	<1
Jurassic	81	11	<1	<1	<1	4	3	–
Triassic	65	–	11	<1	7	8	8	–
Permian	8	–	5	–	10	27	43	7
Carboniferous	6	<1	2	1	2	18	70	<1
Devonian	50	7	5	–	5	12	16	5
Ordovician	100	–	–	–	–	–	–	–
Cambrian	83	–	–	–	–	15	–	2

Table 14. Percentage distribution of bauxite colours over the deposit types.

Deposit type	Red, violet	Yellow	Multi-coloured	White	Dark brown	Green	Grey	Black
Mediterranean	63	10	12	2	2	3	7	1
Timan	5	1	1	1	4	15	68	5
Kazakhstan	58	14	13	9	–	–	4	2
Ariège	70	–	19	–	–	5	6	–
Salento	60	2	27	–	2	1	8	–
Tulsk	–	10	–	60	–	–	30	–

abundance decreases, and they are absent from the Quaternary. White bauxites, whose geographical spread is fairly restricted, are likewise most frequent in the Paleocene and Eocene deposits.

In the geochemical respect, the most important feature is the great abundance of 'reduced' colours in the Carboniferous and Permian, and to a less extent also in the Devonian and Triassic. Bauxite deposition in those periods took place largely in reducing environment. In the remaining geological periods, it is the 'oxidised' colours that form an overwhelming majority; from the Neogene on, they become well-nigh exclusive.

According to the data presented in table 14, red and violet are the most abundant colours in the Mediterranean, Kazakhstan-, Ariège- and Salento-type deposits. All the other colours are also represented in greater or less abundance in the Mediterranean-type deposits, whereas one or more colours are absent from the other deposit types. The Timan-type deposits fundamentally differ in colouring from the above four types: grey bauxite is the most abundant in the Timan deposits, and green bauxite is also frequent. The aggregate abundance of the 'oxidised' colours, red, yellow and multicoloured, is a mere 7%. This circumstance also confirms the necessity of distinguishing this deposit type. The genetic separateness of the Tulsk-type deposits is also confirmed by their

peculiar colour composition: white bauxites are the most abundant, followed by light grey and pale yellow ones, with all the other colours missing.

There may be essential differences in colouring also within the individual deposit types, and indeed even between deposit groups identical by and large as to geological age. A case in point is the Transdanubian Hills region of Hungary; in its southwestern part (at Nyirád, Szőc and Halimba), bauxite is predominantly red, whereas in the Iszkaszentgyörgy and Gánt deposit groups yellowish-brown and multicoloured bauxites are the most abundant. No black or dark brown bauxite has been discovered so far in any of the Hungarian deposit groups; green bauxites are confined to Gánt and Nagyharsány, and very subordinate even there. White bauxite is also very rare; it is known from Gánt, Nézsa and Nagyharsány, in quantities not exceeding 5%.

In Jamaica, the following relationship is observed between bauxite colour and altitude. The deposits of the high plateaux are predominantly red, whereas in the tectonic grabens e.g. in the Porus Valley—and on the deeper-lying limestone plains—e.g. on the St Elisabeth Plain—red bauxite alternates with ochroid, russet and multicoloured bauxite. In the Pepper area deposit, situated in the deepest-lying part of the St Elisabeth Plain, the bottom part of the bauxite deposit is a light buff colour; this is the lightest-coloured bauxite on the island. It gradually passes into red bauxite upward.

The colour table of Appendix 2 enables one to find out about what colour is the most frequent and where. According to the relevant literature, several bauxite deposit groups consist entirely of red bauxite. A high organics content lends a characteristic fallow brownish-red colour to the bauxites of the Pacific islands. A single small deposit group is known which is composed entirely of yellow bauxite: it is Nadrt in Slovenia. Yellow bauxite attains substantial abundances on Velika Planina in Slovenia (60%), and in the Nurra deposit group of Sardinia (50%). Multicoloured bauxite is found in greatest abundance at Kislőd in Hungary (50%), at Maktesh Ramon, Israel, (40%) and Sisophon–Battambang, Campuchea, also (40%).

White bauxite is most frequent in the Missouri deposit group, US (60%) and in the small deposits of the Tulsk and Ryazansk areas, USSR (also 60%). White bauxitic clay is also fairly abundant at Bijele Poljana, Montenegro (40%). Dark brown bauxite is most abundant in the two bauxite horizons of Jajarm in Northern Iran (50% in each), and at Yazd also in Iran (30%). Green varieties of bauxite are most frequent at Berdsk–Obukhovsk in the USSR (50%), at Kozina in Slovenia and in the Kandahar deposit group, Afganistan (40% each). Grey bauxite is most frequent in China. According to the data in the relevant literature, the Liaoning deposit group consists entirely of such bauxite, whereas in the rest of the deposit groups its average abundance is 70%. The bauxite of the Southern Timan deposits group is also predominantly grey (70%). Black bauxite is most widespread among the more or less metamorphic bauxites and emeries, e.g. on Naxos Island, Greece (30%), and in the Menderes Mountains, Turkey (10%). It does not exceed 10% in any other deposit group.

4.5.5.2. MULTICOLOURED VARIETIES OF BAUXITE

Multicoloured bauxite varieties exhibit several types of arrangement of colours and forms, the most frequent of which are the following.

(i) *Light veins on a dark background.* By the term '*veins*' the writer means elongate

stripes of bauxite distinct in colour from their surroundings, regardless of texture. They should not be confused with fissure fillings, which are textural elements. These veins are most often of 0·5 to 5 cm diameter and 10 to 200 cm long; their outlines are usually sharp, more seldom indistinct. They are predominantly vertical or near-vertical, downward-tapering. In Hungary, for example, they are frequent in the upper parts of the deposits at Nyirád, Szőc, Halimba, Kislőd, and Iszkaszentgyörgy. The veins are yellowish white, pale yellow or pink against a background that is predominantly red.

In quite a few places, a regular succession of colours is observed: the central part of the vein is a pale violet or pink; it is surrounded by pale yellow bauxite; there is a gradual transition between the two. The surrounding brick-red bauxite is darker over a width of 0·5 to 1·0 cm about the veins than elsewhere. This colour pattern has been observed by the writer over lengths of several tens of metres in the roadways; indeed, it is repeated in deposits several kilometres apart. In the writer's opinion, these vertical veins of a late diagenetic and epigene origin are due to the action of downward-seeping ground water of acidic pH. Along the veins, most of the original iron content was leached out and removed laterally and downward. A small part got precipitated along the veins, causing the above-mentioned darker red coloration.

(ii) *Lighter spots of irregular outline on a darker backround.* The spots are of 1 to 30 cm diameter. Their outlines are distinct as a rule, more seldom hazy. This colour pattern is likewise independent of the bauxite's textural elements. Such bauxite is especially frequent at Gánt and Iszkaszentgyörgy in Hungary; it repeats the colour succession described in connection with the veins (colour figure 19). In the writer's opinion, these spots are also due to diagenetic and epigene iron leaching, with the difference that here the seepage of ground water was not strictly unidirectional.

(iii) *Round or oval light spots of sharp outline on a darker background.* These are almost exactly circular in places. Their diameter ranges from 1 mm to 3 cm. They are independent of the textural elements. Most of them are red, pale yelow or off-white. There may be a slightly darker spot at the centre. Yellowish or pink round spots of this type with grey or dark grey centres are frequent e.g. in the rust-red bauxite of the Middle Triassic bauxite horizon of Slovenia (colour figure 20). The outlines of the spots are entirely independent of the densely ooidic texture of the bauxite. Such spots are also frequent at the bottom of the deposit in the Halimba III mine, Hungary.

These spots may also be due to a secondary leaching of iron. Their regular outline and isolation, however, precludes their attribution to seeping ground water. The action of micro-organisms creating local centres of reduction, facilitating the leaching away of iron after its reduction to the ferrous form, seems more plausible.

(iv) *Darker spots on a lighter background.* These tend to be attached to textural elements. It is the matrix that tends to be light, and the ooids, pisoids, roundgrains, bauxite pebbles and fragments that tend to be darker. This type of multicoloured bauxite is especially frequent in bauxites having a conglomerate or breccia type texture. Such are the colour patterns of the roundgrained bauxite in colour figure 7 and of the pisoidic–roundgrained bauxite in colour figure 10.

This type of bauxite is especially frequent in the top part of deposits of the Kislőd deposit group, Hungary. As to texture, it is a conglomerate or breccia with a light yellow or ochroid matrix embedding hard, massive brick-red lumps of bauxite. This is a

case of epigene deferrification which removed most of the iron from the less massive, more permeable matrix, leaving unaffected the more massive, less permeable bauxite pebbles and lumps.

(v) *Light ooids, pisoids and macro-pisoids in a darker matrix.* A fairly rare colour pattern. In the upper part of the bauxite deposit near Lescalé, in the Ariège (France), the writer has observed a brick-red bauxite containing yellowish-white pisoids. In the Missouri deposit group, US, the writer has seen a few cases of a grey bauxite of ooidic texture some of whose ooids where off-white or pure white.

The mechanism responsible for this colour pattern is not yet understood: it is presumably another case of iron leaching. Pyrite and marcasite may conceivably have concentrated in certain ooids and pisoids, and subsequent oxidation may have leached out only those.

(vi) *Striped or banded bauxite.* Lighter and darker stripes in near-parallel, concentric or capriciously swinging patterns independent of bauxite texture. Stripe width ranges from 1 mm to 2 cm. The colour pattern is similar to that of banded agate. It is usually confined to small areas, a few square metres at most. It is usual to call the tortuous bands '*liesegang rings*' after Liesegang who explained the phenomenon early in this century, proving by model experiments that the ions of an ionic solution diffusing through a finely dispersed medium will precipitate rhythmically (periodically), in patterns resembling those of banded agate. It is the finely disperse texture of bauxite that permits this phenomenon to operate. The solutions that gave rise to the lighter and darker bands usually contained ferric ions, more seldom manganese ions.

The writer has found the liesegang rings in bauxite to be largely due to pyrite oxidation. In Hungary, e.g. in the Félix II open pit at Szőc, a light and dark violet striped bauxite in a thickness of 0·5 to 1 m occurs in places where a fully oxidised pyritic bauxite is usually found. In the underground workings in Dűlt-Nyires V lens at Nyirád, liesegang rings are connected with near-vertical root traces in a light grey, pyritic bauxite. Wherever the bauxite was not exposed to no subsequent oxidation, these vestiges are conspicuous for a dark grey colour and a higher pyrite content. In places, downward-seeping ground water would progress fastest along the root traces oxidising the pyrite in the process. Mobilised iron reprecipitated in four to six concentric liesegang rings about the roots (colour figure 21). In Dűlt-Nyires VII lens, a liesegang-striped bauxite surrounding a grey pyritic bauxite was found. In the lenses at Nyirád–Izamajor also, a direct connection between grey pyritic bauxite and bauxite with liesegang rings was repeatedly observed (colour figure 22). The phenomenon has been observed also at Iszkaszentgyörgy, although it is rarer there.

The liesegang-striped bauxite of the Trubelja deposit, Crna Gora (Montenegro), called 'leopardite' by Dragović (1965), exhibits an alternation of light grey stripes and bands with darker grey and greenish brown ones. The term is a misnomer, though, as the leopard is spotted not striped.

According to Bushinsky (1971), liesegang-striped bauxite can also be found in the Devonian deposits of the Northern Urals, USSR. It is especially frequent in the open pits of the Krasnaya Shapochka, showing predominantly reddish and brownish, ferruginous bands, but more seldom also black, manganifereous ones. Bushinsky attributes them to the near-surface oxidation of piritic bauxite.

In summary, multicoloured bauxites are the products of a considerable variety of secondary processes, controlled by the texture and permeability of the bauxite and by the composition of the solutions permeating it. Micro-organisms may also have contributed their influences.

4.5.5.3. THE STAIN OF BAUXITE

Some bauxites stain the hand; others do not. This property has been used in the USSR for decades in the field description of bauxites ('markiy, nemarkiy boksit'). The writer deems it necessary to distinguish the following categories of staining:

(i) strongly staining bauxite,
(ii) staining bauxite,
(iii) weakly staining bauxite,
(iv) non-staining bauxite.

In the writer's experience, the harder the bauxite, and the less its porosity, the less it stains the hand. It is the high-porosity, uncompacted bauxites that stain the most. The intensity of staining is independent of bauxite colour: white bauxite may also give a strong stain.

The group of strongly staining bauxites includes among others those of Jamaica and Hispaniola. Most bauxites of the Transdanubian Hills in Hungary are of the staining variety, some very strongly so. Nagyharsány bauxite is weakly staining. Most of the hard Paleozoic bauxites are non-staining, although weakly and averagely staining types have also been described from the Devonian deposits of the Urals.

Staining intensity tends to be uniform by and large over a given deposit, although types giving a stronger and a weaker stain may coexist in certain deposits. Observations as to the lie and distribution of these types may be helpful in tracing a variety of mineralogical and petrographic processes.

4.6. THE PETROGRAPHIC ARRANGEMENT OF THE BAUXITE DEPOSITS

Bauxite deposits tend to exhibit certain regularities in petrographic constitution and in the spatial arrangement of their petrographic features. Let us review these features deposit type by deposit type.

In the *Mediterranean-type bauxites*, ore grade tends to be best at the middle. Clay content increases upward and downward as well as towards the deposit margins. More seldom, the deposit consists predominantly of high-grade ore, surrounded by a thin shell 0·5 to 1 m wide of clayey bauxite and bauxitic clay. This latter group includes some deposits from Jamaica; Dalmatia, Crna Gora (Montenegro) in Yugoslavia; and Central Greece. Deposits of the same sort of concentric build but containing about as much clayey bauxite and bauxitic clay as bauxite in the strict sense are much more common. There are, however, also deposits in which bauxite s.s. is confined to the top third or fourth of the deposit, the rest consisting of clayey bauxite, bauxitic clay and bauxite-like clay. Some deposits of the Szőc Basin in Hungary belong to this type.

Slides and slumps due to subsequent karstification may upset the original regular arrangement, giving rise to a chaotic structure in the entire deposit or in parts of it. This is what happened in several deposits of the Bédarieux group, France (fig. 110).

The Devonian bauxites of the Urals, USSR, often exhibit a peculiar two-phase arrangement: a thicker bottom unit made up of reddish-brown high-grade bauxite underlain by clayey bauxite and bauxitic clay, and a top unit, 1 to 2 m thick, made up predominantly of a greenish or greyish bauxite, mostly clayey, and of bauxitic clay (fig. 29). As a rule, the bottom unit is weakly bedded; the top unit is well-bedded. This, then, is a primary, syngenetic difference, due to a change in the medium of deposition: the bottom unit was deposited largely on land, the top unit already under lagoon and neritic conditions (Gutkin 1978).

In those Mediterranean-type deposits where primary reworking played an important role and immediate burial prevented any further bauxitisation, bauxitic rock types of different composition tend to form irregular successions of thick banks. This is the case e.g. in the extensive stratiform deposit of the Halimba Basin, which even contains layers of dolostone debris intercalated in the bauxite (figs 113 and 114).

Several bauxite deposits exhibit internal erosional disconformities. Combes (1971, 1973) observed three surfaces of disconformity within the deposit called Uston at Bédarieux, France, in those places where subsequent bedrock karstification was weak (fig. 115). Wherever subsequent karstification was strong, on the other hand, this primary deposit structure was effaced by slides and slumps.

Of the *Timan-type deposits*, those of the Southern Timan, USSR, also contain the highest-grade bauxite in the central parts of the ore bodies (Gulyanitsky 1969). It is surrounded first by clayey bauxite, then by bauxitic clay and bauxite-like clay. The Chinese deposit groups are also similar in deposit structure.

The bauxitic complex of the *Kazakhstan-type deposits* contains ore bodies in one or several horizons, the petrographic constitution of which exhibits no overriding regularity.

Texture as well as petrographic composition may vary within a deposit. In bauxites of an allogenic–clastic texture type, repeated changes of particle size within the deposit profile are predominant. The phenomenon has been described above in connection with the phenomenon of graded bedding, but even in simply bedded bauxite the beds tend to alternate between clastic texture types of finer and coarser particle size.

In authigenic textures, generally a gradual increase in the size of textural elements can be observed from the bottom of the deposit up. In the clayey bottom parts, the texture is pelitomorphic as a rule. Ooids appear some way up. They grow bigger and bigger, until the texture becomes pisoidic. The abundance of ooids and pisoids per unit surface area also increases in step. Macropisoids tend to be confined to the top parts of the deposit, e.g. in the Arkalyk deposit group, USSR, where the bauxite, uncompacted or low-hardness elsewhere, is cemented to hard blocks of 1 to 3 m^2 in size. These hard blocks form strings that can be traced over 20 to 100 m of lateral extent; they grow gradually smaller outward and then disappear. They have a pisoidic–macropisoidic texture, and resemble in appearance the ferricrete crusts ('cuirasses') of lateritic bauxites. Their origin also is presumably due to similar processes. They are called '*stony bauxite*' ('kamenisty boksit') in Soviet literature.

Secondary processes imparted a great variety to bauxite profiles also in the petrographic sense. The most frequent phenomenon is the fading of bauxite near the cover, in a thickness of 0·5 to 3 m, most often as a result of diagenetic or epigene deferrification. This zone tends to include those bauxite types the pisoids and bauxite pebbles and lumps included in which have remained darker thanks to their greater compactness and lower permeability. This is a very frequent phenomenon in the bauxite deposits of France, Italy, Yugoslavia and Greece.

In those deposits of the Transdanubian Hills, Hungary, where the immediate cover is a swamp-facies carbonaceous clay or coal, it is the top 0·5 to 2 m of bauxite that got reduced. Part of the iron content was leached out, but some of it was re-precipitated in the form of pyrite grains. The bauxite, which turned a light or dark grey in the process, also contains coalified roots. In the outcrops, and wherever the depth of burial is less than 50 m, the grey bauxite underwent a supergene oxidation which changed its colour to a pale violet, pink, orange or light cream. Iron was re-mobilised and re-precipitated in concretions, nests and crusts within the zone or at its bottom. Reprecipitated iron in places formed liesegang stripe or ring patterns. This upper zone rather conspicuously differs in colour from the less affected middle part of the deposit profile. The interface between the two is smooth as a rule, irregular more rarely and quite indistinct occasionally.

At the top of the middle zone there is most often a rust- or brick-red bauxite with yellow veins of the type described in Section 1.5.5.2. It is most often 0·5 to 2 m, exceptionally up to 3 m thick. Most of the veins are perpendicular to the deposit top (that is, vertical if the deposit is undisturbed). Structural dislocation may have tilted them into oblique positions, but even then they are perpendicular to the deposit top. The veins, gradually tapering downward, pass into spots of decreasing size. This yellow-veined bauxite can be found beneath both the grey bauxite and its supergene products of oxidation; that is, it cannot be a product of pyrite oxidation. It was presumably the downward-seeping reducing solutions of humic acid which in the phase of pyritisation dissolved the iron content out of a bauxite which had been a uniform red previously.

Figure 125, a collection of type profiles from the most important deposit groups of the Transdanubian Hills, shows these groups to also exhibit some individual traits in addition to the common features just described.

In the deposits which have remained unburied ever since the accumulation of bauxite, alterations by secondary processes were much more insignificant. This category includes the deposit groups of Jamaica, Hispaniola and the Pacific. Their deposits are mostly uniform as to both colour and texture, and only exhibit the smooth changes in ore grade described at the beginning of this section.

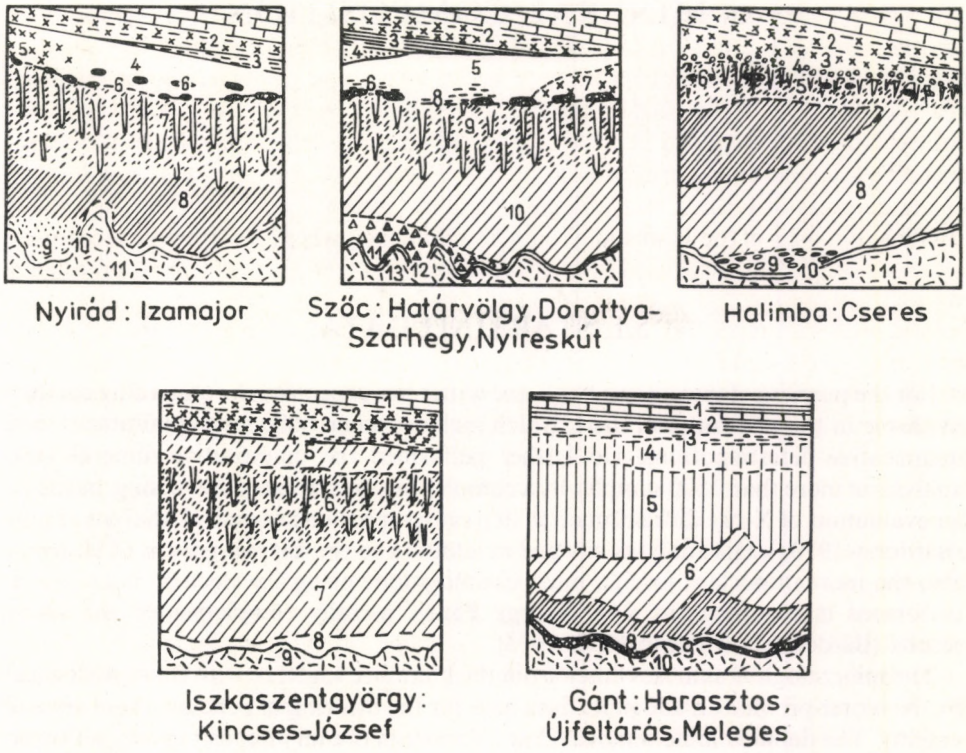


Figure 125. Typical lithological deposit profiles of the principal deposit groups of the Transdanubian Hills, Hungary

Nyirád: Izamajor

(1) yellow and grey miliolid limestone; (2) grey carbonaceous, pyritic clay; (3) multicoloured, yellow lamellar clay; (4) orange and light meat-red clayey bauxite; (5) light grey pyritic bauxitic clay; (6) dark rust-red ferruginous crust and nodes; (7) rust-red bauxite with yellow veins; (8) rust-red bauxite and clayey bauxite; (9) pink and pale violet bauxitic clay; (10) ochroid bauxitic clay; (11) dolostone

Szőc: Határvölgy. Dorottya-Szárhegy, Nyireskút

(1) brownish yellow miliolid limestone; (2) grey and brown carbonaceous clay; (3) multicoloured lamellar clay; (4) light cream bauxite; (5) brownish yellow and orange bauxite and clayey bauxite; (6) violet clayey bauxite; (7) light grey pyritic clayey bauxite; (8) dark red ferruginous crust; (9) brick-red bauxite with yellow veins; (10) rust-red clayey bauxite and bauxitic clay; (11) yellow, yellowish brown bauxitic clay; (12) ochroid bauxitic clay; (13) dolostone

Halimba: Cseres

(1) grey and yellow miliolid limestone; (2) grey pyritic, carbonaceous clay; (3) grey pyritic, clayey bauxite; (4) multicoloured breccia of clayey bauxite; (5) rust-red bauxite with yellow veins; (6) ooidic-pisoidic clayey bauxite; (7) rust-red bauxite; (8) rust-brown clayey bauxite; (9) red bauxitic clay with dolostone and limestone pebbles; (10) yellow and off-white bauxitic clay; (11) dolostone and Dachsteinkalk

Iszkaszentgyörgy: Kincses–József

(1) yellow and grey miliolid limestone; (2) dark grey carbonaceous, pyritic clay; (3) dark grey pyritic, clayey bauxite; (4) violet pyritic, clayey bauxite with ferruginous nodes; (5) yellow, brownish-yellow bauxite; (6) yellowish brown bauxite with yellow veins; (7) brick-red and reddish-brown bauxite; (8) brick-red bauxite and bauxitic clay, russet in its lower parts, (9) dolostone

Gánt: Harasztos, Újfeltárás, Meleges

(1) yellow and grey Melania limestone; (2) brownish yellow and grey clay; (3) red and multicoloured bauxitic clay; (4) multicoloured 'bauxite agglomerate' with pisoids; (5) rust-red and yellowish-brown bauxite; (6) ochroid and orange, massive bauxite; (7) reddish-brown clayey bauxite; (8) yellow low-iron clayey bauxite; (9) reddish-brown ferruginous crust; (10) Hauptdolomite

5. MINERAL COMPOSITION

5.1. STARTING DATA

For the purpose of the present book, the writer has processed the mineralogical data available in the relevant literature, which includes some 6000 semi-quantitative and quantitative determinations. He further performed the quantitative mineralogical analysis of more than 1500 samples by a complex method of his own devising, based on an evaluation of X-ray diffractometric, derivatographic and chemical analysis results (Bárdossy 1970a). His analyses covered in addition to the deposit groups of Hungary also the more important foreign ones. Examinations using the electron microprobe, performed in co-operation with György Pantó, usefully complemented the above results (Bárdossy and Pantó 1970, 1973).

The mineralogical analyses taken from the literature together with those performed by the writer provide an adequate data base for the drawing of conclusions of general validity. The detailed mineralogical characterisation of every deposit group, however, will require thousands of further analyses.

The total number of mineral species so far identified in karst bauxites is 160. Of these, only 20 are known to be represented in more than 10 deposit groups in abundances exceeding one percent over the deposit as a whole. These minerals which thus constitute the bulk of karst bauxites are called rock-forming bauxite minerals by the writer. In what follows, let us first survey the distribution, abundance and associations of these; we shall pass on to a description of all the minerals known from karst bauxites afterwards. Finally, the conditions of origin of all these minerals shall be discussed.

5.2. DISTRIBUTION AND ASSOCIATIONS OF THE ROCK-FORMING MINERALS

5.2.1. DISTRIBUTION AND AVERAGE ABUNDANCES OF THE ROCK-FORMING MINERALS

From the data at his disposal, the writer determined the number of deposits in which these minerals are represented and calculated average abundances for each of the deposits in question. He then prepared the histograms shown in figure 126. The columns are defined by the percentage limits 0·1 to 1%, 1 to 5%, 5 to 10%, 10 to 20%, 20 to 40%, 40 to 60% and 60 to 70%. Column height is proportional to the number of deposit groups which contain the mineral in the abundance range indicated. A

percentage calibration of the ordinate has been avoided deliberately in order to give the reader a direct insight into deposit group numbers. The numbers in circles state the total number of deposit groups in which the mineral in question is known to be present. No mineralogical data were available for seven deposit groups. The aggregate bauxite tonnage of these is less than ten million tons, giving less than 0·1% of world total karst bauxite tonnage.

The distribution of a mineral does not correlate with its average abundance. Rutile and anatase, for example are among the most widespread minerals of karst bauxites, despite their low average abundances. Boehmite, diaspore and gibbsite, whose average abundances are greatest, are each known to occur in a much smaller number of deposits.

Of the aluminium minerals, *boehmite* is the most widespread. It tends to be abundant wherever it occurs (its most frequent abundance range is 40 to 60%, cf its histogram in fig. 126). *Diaspore* is much less widespread, but abundant wherever it is present (40 to 60% and 20 to 40% being its most frequent abundance ranges). *Gibbsite* resembles diaspore in this respect, but its average abundance in the deposit groups is lower, most often in the 20 to 40% range. *Corundum* is much less widespread than the above three minerals. It has two frequency peaks, one at 20 to 60% in metamorphic bauxites, the other at 0·5 to 10%, in non-metamorphic ones. In the first group, corundum is invariably macrocrystalline; in the second, it is crypto- and microcrystalline.

Of the iron minerals, *haematite* is both the most widespread and the most abundant in terms of deposit-group averages. Its most frequent abundance range is from 10 to 20%. *Goethite* is not much less widespread than haematite, but its most frequent abundance range is 1 to 5% only. The histograms show these two to be the most widespread and most abundant iron minerals.

Of the silicate minerals, *kaolinite* is the most widespread by far, identified so far in 150 deposit groups. Its abundance is in the 20 to 40% range in 72 deposit groups, in the 40 to 60% range in 11, and in the 60 to 70% range in 3. The other phyllosilicates are subordinate to kaolinite, in terms of both distribution and average abundance.

The two most important titanium minerals are *anatase* and *rutile*. Their distribution is near-identical: they are known from an overwhelming majority of all deposit groups. Their most frequent abundances are fairly low, however (1 to 5% for anatase and 0·1 to 1% for rutile).

It follows from the above findings that bauxites are essentially made up of eight minerals, represented in an overwhelming majority of all the deposit groups. These minerals are boehmite, diaspore, gibbsite, haematite, goethite, kaolinite, anatase and rutile.

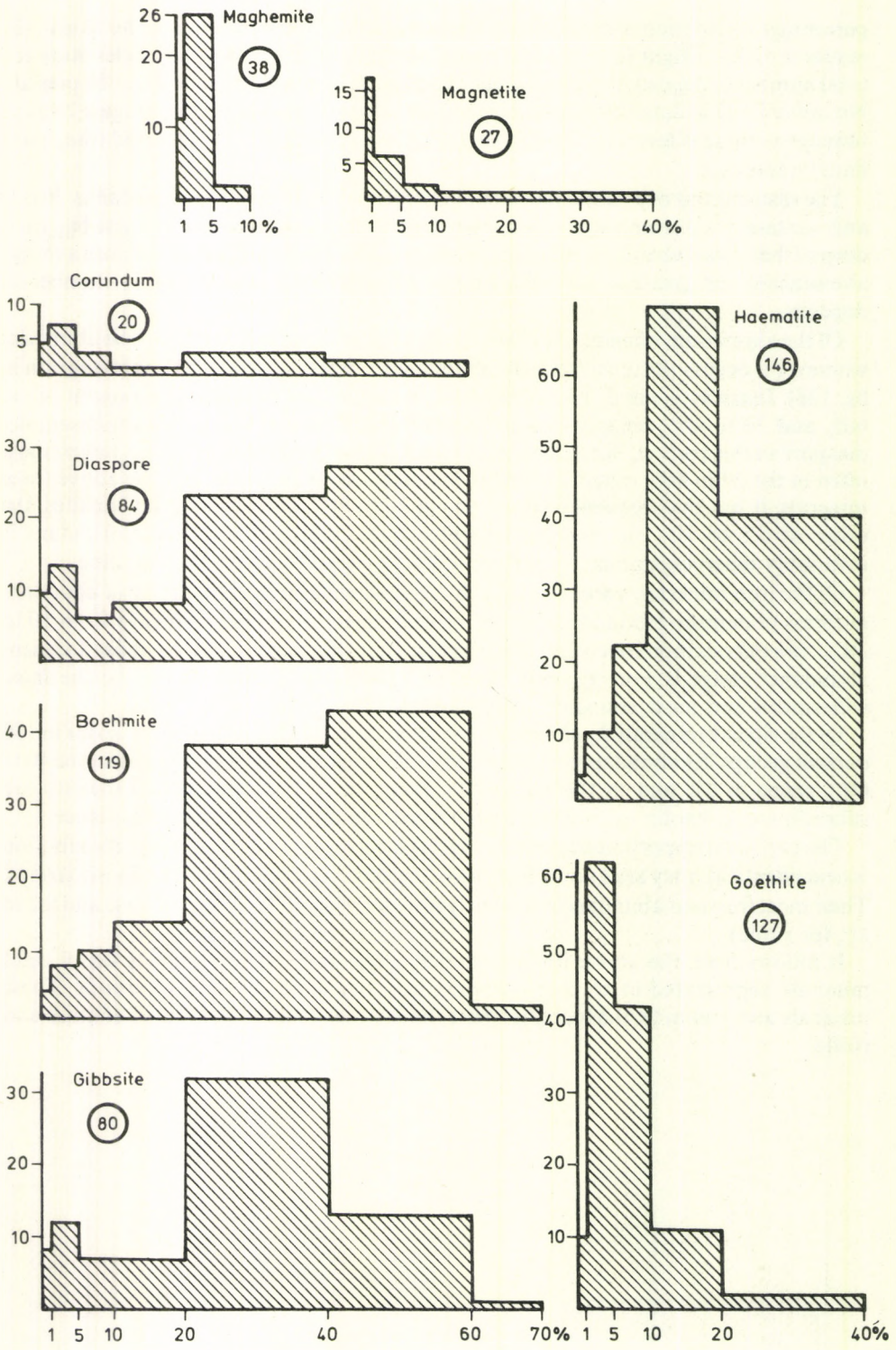
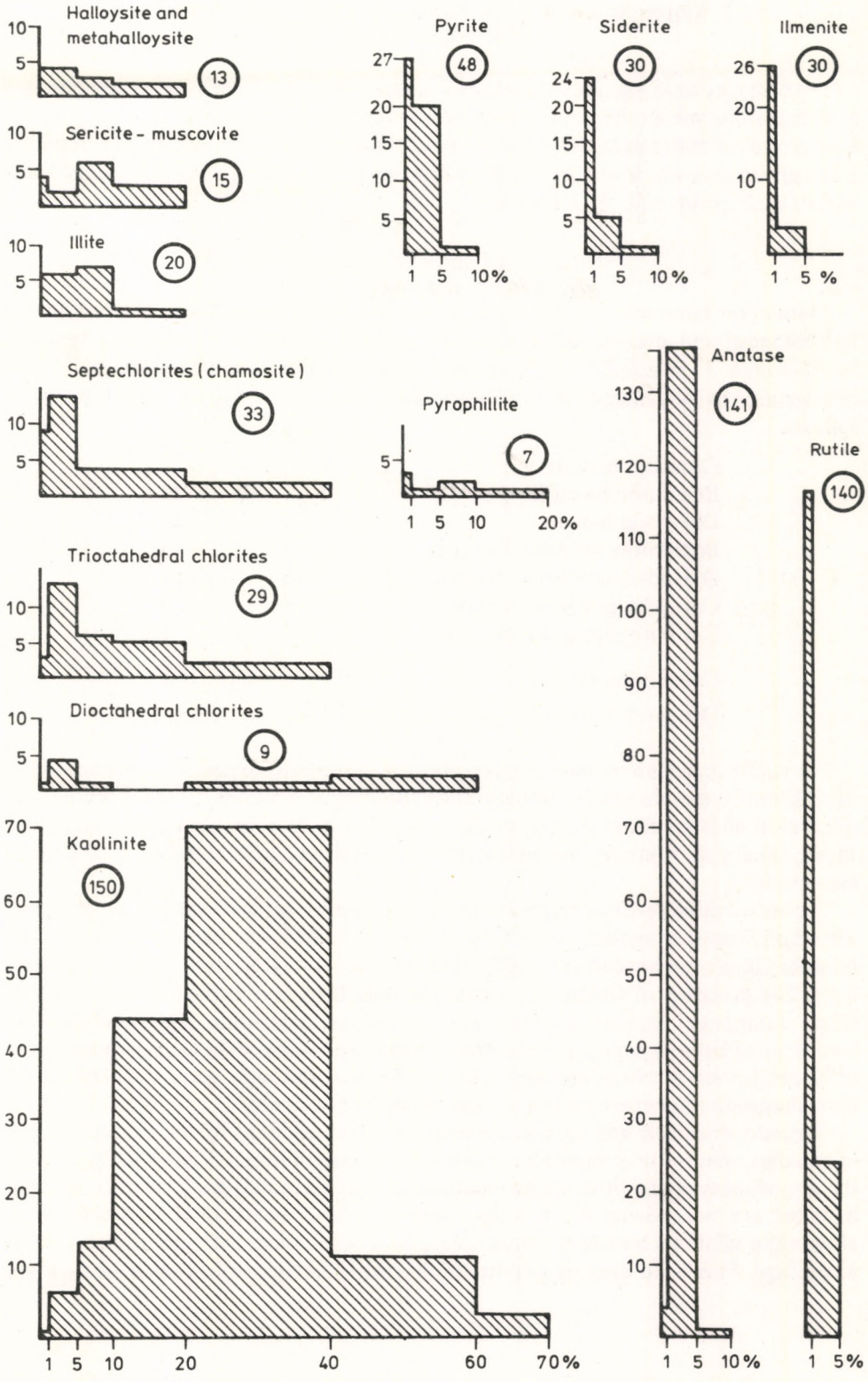


Figure 126. Histograms of the distribution of rock-forming bauxite minerals. The number in the circle is th number of deposit groups in which the mineral is found. The ordinate axis is calibrated in deposit-grou numbers, the abscissa axis in the average abundance of the mineral in question


Halloysite and metahalloysite
13
Sericite - muscovite
15
Illite
20
Pyrite
48
Siderite
30
Ilmenite
30
Septechlorites (chamosite)
33
Pyrophillite
7
Trioctahedral chlorites
29
Dioctahedral chlorites
9
Kaolinite
150
Anatase
141
Rutile
140
1 5 10 20 40 60 70%
1 5 10%
1 5%

5.2.2. RELATIVE ABUNDANCES WITHIN THE PRINCIPAL MINERAL GROUPS

The best way to recognise mineral associations is by examining relative abundances and their shifts within the principal mineral groups. In the following, we shall examine four groups of minerals in this fashion, those of aluminium, iron, silicon and titanium. Each group as a whole will be taken to represent 100%, and percentage abundances within each group will be established.

5.2.2.1. ALUMINIUM MINERALS

Aluminium minerals in karst bauxites constitute nine associations. The estimated percentage abundances of each of these are presented deposit group by deposit group in Appendix 3. Weighted with deposit-group tonnages, the abundances of these associations as percentages of total world karst bauxite tonnage are found to be as follows.

Gibbsitic bauxite	24%
Boehmitic bauxite	23%
Diasporic bauxite	20%
Boehmitio–gibbsitic bauxite	15%
Diasporo–boehmitic bauxite	14%
Corundo–gibbsitic bauxite	2%
Corundo–diasporic bauxite	1%
Corundous bauxite Diasporo–boehmito–gibbsitic bauxite	1%
	100%

The writer regards a bauxite as pure gibbsitic, boehmitic, diasporic or corundous if, in addition to the aluminium mineral thus specified, it contains no other aluminium mineral in an abundance of 10% or more (as referred to total aluminium minerals). In the binary and ternary associations, each of the minerals specified in the name exceeds 10%.

The above table reveals first of all that single component associations are the most abundant, together making up 67, 6% of world total karst bauxite tonnage. The binaries taken together end up to 32%; the single ternary association represents a mere 0·3%. The processes of aluminium-mineral genesis, then, favour the formation of pure single component systems representing equilibrium situations. The conditions of formation of binary systems are less often satisfied, and those of the ternary system only quite exceptionally. No quaternary system is known, although, given the existence of four aluminium minerals, such a system is not inconceivable.

Corundo–diasporic and corundous bauxites were produced by the metamorphism of bauxites, that is under highly specialised conditions of origin. Remarkably enough, the abundances of the three single component systems in non-metamorphic, normal bauxites are near-identical. In karst bauxites, then, the three non-metamorphic aluminium minerals had about equal chances of developing, with gibbsite at a slight advantage. This is an apparent contradiction with the fact of boehmite being more

Table 15. Weight percent distribution of aluminium-mineral associations over the deposit types.

Deposit type	Gibbsitic	Gibbsito–boehmitic	Gibbsito–boehmito–diasporic	Gibbsito-corundous	Boehmitic	Boehmito–diasporic	Dia-sporic	Diasporo–corundous	Corundous
Mediterranean	24	21	0·3	–	30	11	12	1	<1
Timan	6	3	0·2	–	4	25	62	–	–
Kazakhstan	70	<1	–	24	1	1	3	–	–
Ariège	–	<1	–	–	10	60	30	–	–
Salento	2	3	–	–	11	36	48	–	–
Tulsk	100	–	–	–	–	–	–	–	–

widespread. In evaluating the spread, however, it was the number of deposit groups which served as a basis (cf fig. 126) regardless of whether a group contained one million or one thousand million tons of bauxite. The above abundances, on the other hand, are tonnage-weighted, which eliminates the contradiction.

The abundances of the above mineral associations in lateritic bauxites have not yet been quantified. Enough is known, however, to suggest that gibbsitic bauxites are the most abundant; boehmitic ones are much less so; and diasporic and diasporo–boehmitic ones are quite subordinate. More recently, a gibbsite–corundum association has also been identified in laterite bauxites: the crypto- and microcrystalline corundum that they contain was formed under Earth's surface pressure and temperature. These differences are due partly to a difference between the processes of karstic and lateritic bauxite formation, and partly to the greater diversity of the secondary processes that had acted upon the karst bauxites.

Table 15 reveals essential differences in mineral composition between the depositological types. The Mediterranean type has the most varied composition; it encompasses all aluminium-mineral associations, with the single exception of the corundo–gibbsitic one. Boehmitic bauxites are the most abundant. The more or less metamorphosed bauxites with corondous composition all belong to this deposit type. The Timan type is characterised by the predominance of diasporic and diasporo–boehmitic bauxites. In the Ariège- and Salento-type deposits, too, these two associations are the most abundant. The Kazakhstan-type deposits, on the other hand, are predominantly gibbsitic. This is the only deposit type to contain corundo-gibbsitic bauxite, and in quite high percentages at that. The mineral composition of this depositological type is closest to that of the lateritic bauxites. The writer attributes this to the fact that the processes acting upon these deposits, after their accumulation, were very laterite-like. The Tulsk-type stands apart also in the mineralogical respect, consisting as it does of gibbsitic bauxite exclusively.

There is a distinct relationship between the frequencies of aluminium-mineral associations and their geological age (table 16); the abundance succession gibbsite–boehmite–diaspore corresponding to increasing age. Tertiary and Quaternary bauxites are predominantly gibbsitic, Mesozoic ones predominantly boehmitic, and Paleozoic ones predominantly diasporic. Pure gibbsitic bauxites are largely confined to the post-Lower Cretaceous deposits, and so are the gibbsito–boehmitic ones. Their occurrences in some of the older bauxites are sporadic and insignificant; they are

Table 16. Percentage distribution of the karst bauxites' aluminium-mineral associations over the geological periods.

Geological period	Gibbsitic	Gibbsito–boehmitic	Gibbsito–boehmito–diasporic	Gibbsito–corundous	Boehmitic	Boehmito–diasporic	Dia-sporic	Diasporo–corundous	Corundous
Quaternary	83	17	–	–	–	–	–	–	–
Miocene–Pliocene	65	34	–	–	1	–	–	–	–
Oligocene	6	17	–	–	6	29	42	–	–
Paleocene–Eocene	29	34	<1	10	21	3	3	–	–
Upper Cretaceous	11	9	<1	4	50	13	13	–	–
Lower Cretaceous	14	1	2	1	61	9	11	<1	–
Jurassic	–	–	–	–	60	37	3	–	–
Triassic	4	–	1	–	27	41	26	1	10
Permian	–	–	–	–	–	24	54	12	10
Carboniferous	3	2	–	–	2	27	64	1	<1
Devonian	–	–	–	–	18	34	43	3	2
Ordovician	–	–	–	–	–	–	60	40	–
Cambrian	–	–	–	–	–	50	50	–	–

entirely unknown from any pre-Carboniferous deposits. Boehmito-gibbsitic types exist even in the Quaternary, albeit in rather low abundances. This implies that some boehmite could form together with gibbsite even in the youngest karst bauxites. Pure boehmitic bauxite, on the other hand, first appears in an insignificant percentage in the Paleocene–Eocene deposits only, and has its peak abundance in the Lower Cretaceous and Jurassic, to decrease gradually in the older bauxites. The oldest pure boehmitic bauxites are Devonian.

Diasporo–boehmito–gibbsitic bauxite is confined to the interval from the Triassic to the Eocene, in rather insignificant abundances throughout. Corundo-gibbsitic bauxites, restricted, as we have seen, to the Kazakhstan-type deposits, are confined to the interval from the Lower Cretaceous through the Eocene, which is quite a limited temporal range. The youngest diasporo-boehmitic and diasporic bauxites are Oligocene. It should be added that the geological age of some of the Oligocene diasporic bauxites, sparse as they are, is controversial and presumably older (for example in Slovenia). The actual percentage of Oligocene diasporo–boehmitic and diasporic bauxites, then, is presumably much less than stated in the table. Diasporo–boehmitic bauxite becomes enriched to a significant extent in the pre-Jurassic deposits only; its abundance is constant by and large down to the Cambrian. Diasporic bauxite in turn is abundant from the Cambrian to the Permian only. A time shift is thus also apparent in these two associations.

The bulk of corundo–diasporic and coroundous bauxites produced by metamorphism are found in the Paleozoic deposit groups. No such bauxites are known from any post-Lower Cretaceous deposits.

Age, however, is not the only controlling factor of mineralogy, as within any given geological period several associations may occur. *Structural (tectonic) deformation* was another important controlling factor. In order to pinpoint its importance, the

Table 17. Percentage distribution of aluminium-mineral associations in Carboniferous and Palaocene–Eocene karst bauxites as a function of structural deformation.

Degree of deformation	Gibbsitic	Gibbsito–boehmitic	Gibbsito–boehmito–diasporic	Gibbsito–corundous	Boehmitic	Boehmito–diasporic	Dia-sporic	Diasporo–corundous	Corundous
Carboniferous									
Undeformed	30	23	–	–	24	6	17	–	–
Weak	–	–	–	–	–	–	–	–	–
Moderate	–	–	–	–	–	29	71	–	–
Strong	–	–	–	–	–	25	45	20	10
Paleocene–Eocene									
Undeformed	67	–	–	29	4	–	–	–	–
Weak	9	59	–	–	32	–	–	–	–
Moderate	–	–	–	–	–	–	–	–	–
Strong	2	38	3	–	41	11	5	–	–
Very strong	38	62	–	–	–	–	–	–	–

writer has calculated for each geological period the shifts in percentage abundances of aluminium-mineral associations as a function of structural deformation. The calculations referring to the Carboniferous and the Paleocene–Eocene are presented as examples in table 17.

In the Carboniferous, the relationship is fairly unequivocal: gibbsitic and boehmito–gibbsitic bauxites are confined to the undeformed deposits (Timan Mountains). In the deposits of medium and strong deformation, diasporic bauxites predominate. The strongly deformed group includes some metamorphic bauxites also, of pure corundous or corundo–diasporic composition.

In the Paleocene–Eocene, the picture is not so clear-cut. True, gibbsitic bauxite continues to be the most abundant in the undeformed deposits, but there is also some of it in the strongly and even in the very strongly deformed deposits. Boehmitic, diasporo–boehmitic and diasporic bauxites are most abundant in the strongly deformed deposits. It is the composition of the very strongly deformed deposits of Obrovac–Novigrad–Drniš, Yugoslavia that is farthest from expectation, inasmuch as they consist of predominantly boehmito–gibbsitic bauxites accompanied by less abundant gibbsitic ones. It would therefore seem that in certain cases bauxite can remain gibbsitic and boehmito-gibbsitic despite strong structural deformation. The circumstances of and the causes for this state of facts are not well understood. Elucidating them would require a more detailed structural analysis of these deposit groups.

The corundo–gibbsitic association repeatedly referred to—and presented in table 16—is most abundant in the Palaeocene–Eocene deposits: it is, however, restricted to the structurally undeformed ones.

In summary, the control of the aluminium-mineral associations is shared between geological age, structural stress and depositological type. There are some other minor factors which can be pinpointed by analyses of deposit profiles and individual bauxite samples. For a more detailed discussion, the reader is referred to Section 5.2.4. and 5.2.5.

5.2.2.2. IRON MINERALS

In bauxite, the associations of iron minerals are much more manifold than those of the aluminium minerals, the reason being that, whereas all the aluminium minerals are oxides and hydroxides, iron minerals also include sulphides, carbonates and even iron-rich silicates in rock-forming abundance. The writer has distinguished 22 mineral associations, in the following tonnage-weighted abundances:

haematito–goethitic bauxite	42%
haematitic bauxite	23%
chamosito–leptochloritic bauxite	12%
goethitic bauxite	8%
pyrito–leptochloritic bauxite	3%
chamosito–goethitic bauxite	2%
sideritic bauxite	2%
pyritic bauxite	2%
haematito–goethito–maghemitic bauxite	1%
haematito–magnetitic bauxite	1%
haematito–leptochloritic bauxite	1%
haematito–sideritic bauxite magnetitic bauxite haematito–magnetitic bauxite pyrito–goethitic bauxite chamosito–haematito–goethitic bauxite	2%
goethito–maghemitic bauxite pyrito–goethito–haematitic bauxite pyrito–haematitic bauxite siderito–leptochlorito–pyritic bauxite leptochlorito–sideritic bauxite pyrito–sideritic bauxite	1%
	100%

The definitions are the same as with the aluminium minerals: in the single component systems, no iron mineral except the eponym attains ten relative percent. In the binary and ternary systems, each of the minerals named exceeds ten relative percent.

Although the iron minerals form such a wealth of associations, 73% of all karst bauxites belong to those of haematite, goethite and their binary combination. The only other significant association is the chamosito–leptochloritic.

As distinct form the aluminium associations, a binary system is the most abundant here, and by a broad margin at that. Binary systems are numerous, and ternary ones are also more frequent than among the aluminium associations. This is a result of the manifold secondary alterations undergone by the iron minerals. The writer has found no association of four or more minerals. They seem to be precluded by the rules of physico-chemical equilibrium.

Table 18. Weight percentage distribution of iron-mineral associations in karst bauxites as a function of deposit type.

Deposit type	Haematite–goethite–maghemite	Goethite–maghemite	Goethite	Goethite–haematite	Haematite	Haematite–maghemite	Leptochlorite–haematite	Chamosite–goethite	Leptochlorite–chamosite	Pyrite–leptochlorite	Pyrite	Siderite
Mediterranean	1	<1	9	50	26	2	1	3	2	2	2	<1
Timan	–	–	<1	3	4	<1	–	–	71	8	4	9
Kazakhstan	4	–	11	48	26	<1	3	–	–	–	2	5
Ariège	–	–	1	41	48	–	–	4	<1	4	<1	<1
Salento	–	–	1	57	10	–	–	–	–	–	–	–
Tulsk	–	–	100	–	–	–	–	–	–	–	–	–

Table 19. Percentage distribution of iron-mineral associations over the geological periods.

Geological age	Haematite–goethite–maghemite	Goethite–maghemite	Goethite	Goethite–haematite	Haematite	Haematite–maghemite	Leptochlorite–haematite	Chamosite–Goethite	Leptochlorite–chamosite	Pyrite–leptochlorite	Pyrite	Siderite
Quaternary	82	15	1	2	–	–	–	–	–	–	–	–
Miocene–Pliocene	1	–	10	88	1	–	–	–	–	–	–	–
Oligocene	<1	<1	5	76	11	–	–	–	<1	<1	1	–
Paleocene–Eocene	<1	<1	18	62	13	1	–	–	–	–	1	–
Upper Cretaceous	2	–	10	36	41	1	<1	2	<1	<1	4	–
Lower Cretaceous	<1	–	7	37	45	4	1	1	<1	1	2	<1
Jurassic	–	–	2	17	74	–	1	<1	3	–	–	–
Triassic	–	–	–	24	52	<1	8	3	5	3	<1	1
Permian	–	–	–	–	13	17	–	24	24	12	–	–
Carboniferous	–	–	<1	2	6	–	2	–	68	8	4	8
Devonian	–	–	–	7	48	3	7	–	13	17	–	3
Ordovician	–	–	–	–	100	–	–	–	–	–	–	–
Cambrian	–	–	–	–	83	–	–	–	–	17	–	–

Of the principal iron minerals, maghemite is the only one to form no single component system. It is presumably its metastable structure that prevents its accumulation in substantial quantities.

The last 11 associations of the list add up to an aggregate abundance of 3% only. Confined to a few deposit groups, they are indicative of special conditions of origin. They have been left out of our further calculations.

In the lateritic bauxites, too, the haematito–goethitic association is the most abundant. Pure goethitic bauxite, though, is much more abundant than in the karst bauxites. The associations containing leptochlorite, chamosite, pyrite and siderite, on the other hand, are much scarcer. The writer attributes this to the fact that lateritic bauxites were exposed much less often to reducing influences, both during bauxite accumulation and thereafter.

In table 18, the associations are grouped so that those containing ferric iron are on the left, those containing ferrous iron are on the right, and those containing both are in the middle. The fact that the sum do not add up to 100% is due to the omission of the rare types referred to above.

The Mediterranean, Kazakhstan- and Ariège-type deposits are rather similar among them ,with the haematito–goethitic association being the most abundant in all three; it is followed by the pure haematitic one. All three deposit types are predominantly ferric, with the aggregate abundance of ferri-ferrous and pure ferrous associations equalling about 10%. The Timan type, whose most important minerals are the leptochlorites and ferrochamosites, is conspicuously different. Pyritic and sideritic bauxites are also most abundant there. This type contains only 8% ferric associations. In the Salento type, too, haematito–goethitic bauxite is the most abundant. The reworked Vietnamese bauxites belonging to this deposit type very often contain in addition a chamosito–haematito–goethitic association, which is very rare everywhere else. (This is why it does not figure in the table.) The only iron mineral of the Tulsk-type deposits is goethite.

There is a conspicuous relationship between the distribution of iron minerals and geological age (table 19). In Quaternary bauxites, haematitic and haematito–goethitic associations containing accessory maghemite are the most abundant. In the older deposits, maghemite grows scarcer; it disappears in the Lower Cretaceous.

In the Tertiary bauxites, the haematito–goethitic association is the most important. It is accompanied by pure goethitic and pure haematitic bauxite in substantially lower percentages. The very low abundance of ferri–ferrous associations is striking.

In the Mesozoic, the percentages of pure goethitic and haematito–goethitic bauxites decrease with increasing age. They are replaced by significant increases in the percentages of haematitic bauxite. The importance of ferri–ferrous and pure ferrous associations is on an increase, although none yet attains an abundance of 10%. It is in the Mesozoic that leptochlorito–haematitic and chamosito–goethitic associations turn up first.

The percentage distribution of iron mineral associations undergoes a fundamental change in the Late Paleozoic. The abundance of chamosito–leptochloritic bauxites makes some abrupt increases, especially in the Carboniferous deposits. Chamosito–goethitic and pyrito–leptochloritic bauxites are also significant. Pyritic and sideritic bauxites are also enriched. The abundance of haematitic and haematito-goethitic bauxites decreases in step. From the Devonian and before, haematitic bauxite is

enriched once more, while the percentages of ferri–ferrous and pure ferrous associations decrease. The only pure goethitic bauxite is known from the Carboniferous, albeit in an insignificant quantity. The role of haematito–goethitic bauxite is also much reduced against the Tertiary and the Mesozoic. Metamorphic bauxites are characterised by magnetitic and magnetito–haematitic associations.

These differences become even more striking if the aggregate percentages of ferric, ferri–ferrous and ferrous mineral associations are presented period by period (table 20).

Table 20. Percentage distribution of oxidised and reduced iron-mineral associations over the geological periods.

Geological period	Ferric associations	Ferrous and ferri–ferrous associations
Quaternary	100	–
Miocene–Pliocene	100	–
Oligocene	93	7
Paleocene–Eocene	94	6
Upper Cretaceous	89	11
Lower Cretaceous	90	10
Jurassic	93	7
Triassic	77	23
Permian	13	87
Carboniferous	8	92
Devonian	54	46
Ordovician	100	–
Cambrian	83	17

The Quaternary, Pliocene and Miocene deposits contain exclusively ferric mineral associations. Ferri–ferrous and ferrous associations first appear in the Oligocene, but hardly exceed 10% up to the Jurassic. It is in the Triassic that their percentage starts to increase, attaining a peak in the Permian and Carboniferous. It decreases again from the Devonian to the Cambrian. The Ordovician figure refers to a single deposit group of insignificant tonnage. It will presumably be modified by the discovery of any new Ordovician deposits.

Most of the ferri–ferrous and ferrous minerals in bauxites are of syngenetic and diagenetic origin. The fundamental mineralogical differences revealed by table 20, then, must be explained in terms of the conditions of sedimentation at the given time rather than of processes acting in later periods. The conditions of sedimentation are thus revealed to have been much more reductive, especially in the Carboniferous and in the Permian, than in the subsequent periods.

In summary, the principal role in the shaping of iron mineral associations was played by the oxidation–reduction conditions of the environment. Depositological type was a minor controlling factor. The influence of tectonic deformations was less than that of the aluminium associations. In areas of stronger structural stress the haematite/-goethite ratio tended to be displaced in favour of haematite. Some other factors influencing the iron mineral associations will be reviewed in Sections 5.2.4. and 5.2.5.

5.2.2.3. SILICATE MINERALS

The writer has observed 11 silicate-mineral associations in karst bauxites. Their tonnage-weighted percentage abundances are as follows.

kaolinitic bauxite	65%
kaolino–septechloritic bauxite	14%
kaolino–halloysitic bauxite	10%
kaolino–illitic bauxite	4%
kaolino–chloritic bauxite	3%
septechloritic bauxite	1%
chloritic bauxite kaolino–pyrophyllitic bauxite pyrophyllitic bauxite halloysitic bauxite	1%
metamorphic bauxites with a chloritoid, margarite, disthene, straurolite, etc. association	2%
	100%

In distinguishing mineral associations, the percentages and percentage limits used were the same as with the aluminium and iron minerals. Dickite, an occasional accessory mineral, was placed into the kaolinitic group; metahalloysite into the halloysitic group. All the diverse silicate associations of the metamorphic bauxites were lumped together.

Kaolinite is the most important silicate mineral of karst bauxites, present in most of them either on its own or in binary association with some other silicate mineral. Here, too, a single component system is seen to be the most frequent. There are no ternary or higher-order systems.

As concerns lateritic bauxite, the writer estimates that the abundance in those of the pure kaolinitic association does not exceed 50 to 60%: the percentage of kaolino-halloysitic bauxite is higher, and pure halloysitic bauxite is also significant. The abundance of kaolino–illitic bauxite is much the same as in the karst bauxites, but all the other silicate associations are insignificant. For another important difference, many lateritic bauxites contain quartz grains which have survived the lateritisation process thanks to their resistance to alteration.

The percentage abundances of the silicate-mineral associations vary rather widely from one type of deposit to the next (table 21). Kaolinitic bauxite is most abundant in the Kazakhstan deposit type, and somewhat less so in the Mediterranean, Ariège and Salento types. The Timan deposit group differs substantially, dominated as it is by kaolino–septechloritic bauxites associated with frequent kaolino–illitic ones, whereas pure kaolinitic bauxite is quite subordinate. The Tulsk type differs completely from all the others with its predominant halloysitic and rather less its kaolino–halloysitic association. In this deposit type, halloysite tends to be accompanied by some allophane.

Table 21. Percentage abundance of silicate-mineral associations in the depositological types of karst bauxites.

Deposit type	Halloysitic	Halloysito–kaolinitic	Kaolinitic	Kaolinito–illitic	Kaolinito–septe-chloritic	Kaolinito–chloritic	Chloritic	Septe-chloritic	Kaolinito–pyrophillitic	Pyrophil-litic	Metamorphic bauxites
Mediterranean	0·1	14	73	1	4	4	1	1	<1	–	2
Timan	–	–	11	18	68	2	1	–	–	–	–
Kazakhstan	–	<1	95	2	2	<1	–	–	–	–	–
Ariège	–	–	78	5	5	2	–	5	–	5	–
Salento	–	–	68	–	24	8	–	–	–	–	–
Tulsk	80	20	–	–	–	–	–	–	–	–	–

Table 22. Percentage distribution of silicate-mineral associations in karst bauxites over the geological periods.

Geological period	Halloysitic	Halloysito–kaolinitic	Kaolinitic	Kaolinito–illitic	Kaolinito–septe-chloritic	Kaolinito–chleritic	Chloritic	Septe-chloritic	Kaolinito–pyrophillitic	Pyrophyl-litic	Metamorphic bauxites
Quaternary	2	98	–	–	–	–	–	–	–	–	
Miocene–Pliocene	<1	38	60	2	–	–	–	–	–	–	–
Oligocene	–	–	78	–	14	7	<1	—	—	—	—
Paleocene–Eocene	–	<1	93	<1	1	4	–	<1	–	–	–
Upper Cretaceous	–	<1	95	<1	2	<1	0·2	1	<1	–	–
Lower Cretaceous	–	–	95	1	<1	2	2	<1	–	–	–
Jurassic	–	–	95	<1	<1	<1	3	<1	–	–	<1
Triassic	–	–	71	3	15	4	–	3	–	3	2
Permian	–	–	12	–	24	12	–	12	–	–	40
Carboniferous	–	–	13	17	65	3	<1	–	<1	–	1
Devonian	–	–	46	–	21	20	5	3	–	–	6
Ordovician	–	–	100	–	–	–	–	–	–	–	–
Cambrian	–	–	53	30	–	–	17	–	–	–	–

The data in table 22 reveal the halloysitic and kaolino–halloysitic bauxites to be most abundant in the Quaternary deposits, with lower percentages in the Mio–Pliocene ones, and traces only in the older ones. From the Miocene down to the Triassic, kaolinitic bauxites are predominant. In the Triassic, their percentage starts to decrease, though, becoming subordinate in the Permian and Carboniferous deposits. It is replaced by substantial percentages of kaolino–septechloritic bauxite, and lesser ones of kaolino–chloritic and kaolino–illitic ones. Kaolinite, then, is present also in these associations, albeit not as a single phase. Pure chloritic and septechloritic bauxite types are also most abundant in the Paleozoic, but they are insignificant even there. No pyrophyllitic and kaolino–pyrophyllitic bauxites have so far been identified outside the Paleozoic and Mesozoic deposits.

The above distribution is to be attributed partly to secondary processes, and partly to differences in the original conditions of sedimentation. The disappearance of halloysite from the pre-Miocene deposits is due in all probability to its recrystallisation into kaolinite due to loss of water. Septechlorites and chlorites in bauxites are of syn- and diagenetic origin in their overwhelming bulk and are therefore closely associated in age with the age of the deposits in which they occur. It must therefore be assumed that the above-outlined changes in the silicate–mineral associations were brought about by changes in the conditions of sedimentation.

Structural (tectonic) influences have not affected the silicate–mineral associations significantly. Metamorphism, on the other hand, generated entirely new associations also in the silicate–mineral group.

5.2.2.4. TITANIUM MINERALS

The titanium-mineral associations of karst bauxites are much less diverse than the above groups. The writer has observed seven associations in all, the tonnage-weighted abundance of which is as follows.

anataso–rutilic bauxite	90%
rutilo–anataso-ilmenitic bauxite	4%
rutilic bauxite	2%
anatasic bauxite	2%
rutilo–anatasic bauxite	1%
rutilo–ilmenitic bauxite ulvito–ilmenito–rutilo–anatasic bauxite	1%
	100%

Here too, the abundance limits are outlined in the same way as with the other mineral associations. Anataso–rutilic bauxite, then, is head and shoulders above all the others with regard to abundance. In this binary association, anatase is the predominant mineral; its percentage is two to five times that of rutile. Karst bauxites in which rutile is dominant are few and far between (and figure under the heading of rutilo–anatasic bauxite in the above list). The single phase systems are much less abundant than in any one of the preceding mineral groups. A further remarkable feature is the presence of one

Table 23. Percentage distribution of the karst bauxites' titanium-mineral associations over the geological periods.

Geological period	Anatase	Anatase > rutile	Anatase–rutile–ilmenite	Rutile > anatase	Rutile	Rutile–ilmenite	Ulvite–ilmenite–anatase–rutile
Quaternary	100	–	–	–	–	–	–
Miocene	1	96	3	–	–	–	–
Oligocene	9	25	66	–	–	–	–
Paleocene–Eocene	2	95	3	–	–	–	–
Upper Cretaceous	<1	95	4	<1	–	–	–
Lower Cretaceous	–	97	3	–	–	–	–
Jurassic	<1	98	–	<1	<1	–	–
Triassic	–	92	–	3	5	–	–
Permian	–	–	60	<1	19	21	–
Carboniferous	–	95	–	–	5	–	–
Devonian	–	78	–	10	4	–	8
Ordovician	–	–	–	–	–	–	–
Cambrian	–	–	–	–	100	–	–

ternary and one quaternary system. This is to be attributed in part to the fact that ilmenite and ulvite enter the bauxite largely in clastic form, and do not therefore represent physico-chemical equilibria.

The writer estimates the abundance of pure anatasic bauxite among lateritic bauxites at 30 to 40%. The rest is largely anataso–rutilic, with anatase predominant. Rutilic and rutilo–anatasic bauxites are negligible.

Calculating the abundance distribution of the titanium mineral associations over the types of deposits, the writer could discern no noteworthy deviations, and has therefore omitted the relevant table. The only exception is the Tulsk deposit type, characterised by an almost total absence of titanium.

Despite the great predominance of anataso–rutilic bauxite, there is a distinct relationship between the abundances of the mineral associations and the geological age of the deposits (table 23). The youngest deposits are pure anatasic, whereas the pre-Paleocene deposits contain pure anatasic bauxite in traces only. The older the bauxite, the more important an accessory to anatase is rutile. Pure rutilic bauxite is present in significant amounts only from the Triassic and before; the only bauxites containing no anatase at all are found in the Cambrian. All the metamorphic bauxites are pure rutilic. Anatase, the original titanium mineral of bauxite, then, gradually recrystallised into rutile as geohistoric time passed. This process was presumably promoted, in addition to the time factor, also by overburden pressure, and by metamorphism, if any.

The distribution over geological periods of the associations containing ilmenite and ulvite is irregular. As the bulk of these minerals was introduced into the bauxite in clastic form, their accumulation in this or that period is likely to have been a function of the local conditions of erosion.

Structural (tectonic) deformation had little influence upon the titanium mineral associations. It can be stated, however, that structural stress also tended to displace the anatase/rutile ratio in favour of rutile.

5.2.2.5. COMPARISON OF THE PRINCIPAL MINERAL GROUPS

Of the four principal mineral groups, the iron minerals are seen to be the most varied, forming as they do 22 associations as against 11 for the silicates, 9 for the aluminium and 7 for the titanium minerals. Of this multitude, however, only a few associations are found in noteworthy abundance. This is made clear by the percentage of the most abundant association in each group, which is 90% (for anatase–rutile) in the titanium group, 65% (for kaolinite) in the silicate group, 45% (for haematite–goethite) in the iron group, and 24% (for gibbsite) in the aluminium group. In the last group, boehmitic and diasporic bauxites are but slightly less abundant. This is in a fair agreement with the finding, based on the histograms in figure 126, that the overwhelming bulk of karst bauxites is composed of just eight minerals: boehmite, gibbsite, diaspore, haematite, goethite, kaolinite, anatase and rutile.

In the aluminium and silicate groups, single phase systems are the most abundant (67·6 and 66·9%, respectively). In the iron and titanium groups, on the other hand, binary systems are more abundant (62·5 and 91·3%, respectively). This fact is due partly to physico-chemical causes (the striving towards an equilibrium), and partly to secondary processes.

The types of deposits were found to differ not only in their features as such, but also in mineralogical composition. The differences are greatest in the aluminium-mineral associations. The differences in iron- and silicate-mineral associations between the Mediterranean, Kazakhstan and Ariège types are not great, but the Timan type differs substantially from the rest also in this respect. It is only as regards the titanium-mineral associations that no essential differences were found between the types of deposits.

The mineralogical composition of the Salento type is accidental as a result of secondary reworking, and cannot therefore be regarded as a characteristic feature. Also mineralogically, it is the Tulsk type that differs most from all the others, in further confirmation of its different origin.

The distribution of all mineral groups was found to be in obvious relation to geological age. This is due partly to the influence of the time factor, and partly to subsequent recrystallisation (for example the conversion of anatase into rutile, of maghemite into haematite and of halloysite into metahalloysite). There are, however, certain mineral associations the enrichment of which in this or that geological period is the result of primary or at most diagenetic influences, such as the remarkable enrichment of ferri–ferrous and ferrous associations during the Carboniferous and Permian, and to a less degree in the Devonian and Triassic. This seems to imply peculiar changes in the Eh, the pH and the medium of deposition.

The influence of structural (tectonic) deformations is clear-cut in the case of the aluminium minerals, but much more indistinct in the case of the other three groups.

All the above-outlined calculations were aimed at the elucidation of mineralogical relationships of the most general scope, and therefore encompassed the totality of karst bauxites. In the following, we shall focus upon the regularities governing the mineral compositions of smaller and smaller units (whole bauxite regions or deposit groups first, then individual deposits, and textural elements last).

5.2.3. VARIATION OF MINERAL COMPOSITION WITHIN A BAUXITE REGION OR DEPOSIT GROUP

Mineral composition within a given bauxite region is not uniform as a rule. If the region includes several bauxite horizons, then the differences tend to be less within any horizon than between any two horizons. For an example, let us consider the Dalmatian bauxite region (table 24).

Table 24. Percentage distribution of aluminium minerals in the six bauxite horizons of the Dalmatian bauxite region, Yugoslavia.

Geological period	Gibbsite	Gibbsite > boehmite	Boehmite > gibbsite	Boehmite	Boehmite > diaspore	Diaspore > boehmite	Diaspore
Oligocene	10	90	–	–	–	–	–
Middle Eocene	30	50	20	–	–	–	–
Paleocene	–	–	70	30	–	–	–
Senonian	–	–	–	90	10	–	–
Barremian	–	–	–	100	–	–	–
Middle Triassic	–	–	–	10	30	30	30

Mineral composition is rather uniform within each of the horizons, but each horizon is characterised by the emergence of some new association or associations and the disappearance of one or several others. The relationship found for karst bauxites as a whole, notably that the succession gibbsite–boehmite–diaspore corresponds to increasing geological age, is confirmed here within a single bauxite region. There are, however, some exceptions to this rule, e.g. in central Greece (table 25).

Table 25. Percentage distribution of aluminium minerals in the four bauxite horizons in the bauxite region of Central Greece.

Geological period	Boehmite	Boehmite ≷ Diaspore	Diaspore
Senonian	–	–	100
Turonian	25	35	40
Upper Malm	60	40	–
Lower Malm	30	70	–

In central Greece, it is the youngest bauxite horizon that is the richest in diaspore. The region was exposed to more or less uniform structural stresses, so that this factor may be eliminated as a cause of differences. Nia (1968) suggested that a strong diagenetic diasporisation in the deposits of the Turonian bauxite horizon is the most probable explanation.

In a majority of bauxite regions, the physico-chemical and paleogeographic factors controlling mineral composition were relatively uniform within one bauxite horizon. By the time the next horizon came to be formed, these factors had changed and affected the bauxite differently.

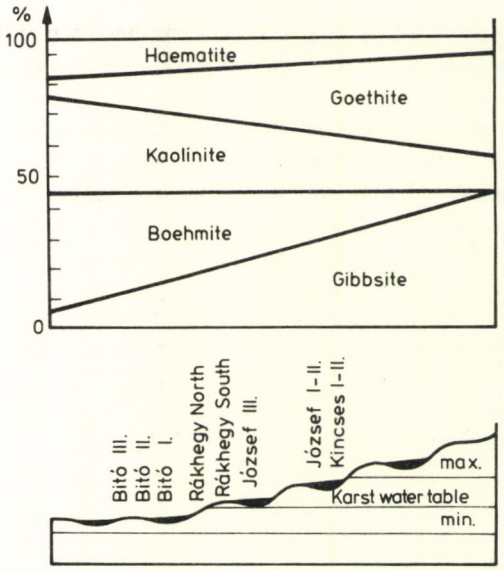


Figure 127. Average abundance of rock-forming minerals in the different ore bodies at Iszkaszentgyörgy, Hungary, according to Komlóssy (1970). The lower diagram indicates the stipulated altitudes of the ore bodies at the time of bauxite formation

In bauxite regions where all the bauxite was formed at about the same time, the mineral composition may exhibit one of the following distributions:

(i) uniform all over the bauxite region,
(ii) deposit averages changing monotonically in some direction,
(iii) composition varying so rapidly and profoundly in each deposit as to frustrate the proving of a coherent trend over the region as a whole.

The credible demonstration of such relationships requires hundreds of mineralogical analyses per bauxite region. Analyses in such numbers are available for a few bauxite regions only, and so the best we can do is to consider a few typical examples of each of the three types enumerated above.

The example presented for the first group is the central Italian bauxite region, where bauxite is of identical constitution throughout both the Abbruzzi and the Campania (boehmite–haematite–anatase–rutile). The individual deposits differ only in the average abundances of the minerals. The Chinese bauxite regions are also of remarkably uniform mineral composition, despite their vast extent.

Our example of the second group is the Transdanubian Hills region of Hungary. The writer in his earlier work (Bárdossy 1961a) found the higher-lying deposits of this region to be richer in gibbsite and goethite than the lower-lying ones which are richer in boehmite and haematite. He attributed this arrangement to the fact that the higher-lying deposits are above the main (regional) karst water table, whereas the lower-lying ones are below it. The dominantly gibbsitic deposits give higher Fe_2O_3/FeO values than the more boehmitic ones. The writer therefore conjectured that a higher Eh gave rise to a gibbsito–goethitic association, and vice versa.

The higher-lying Malomvölgy deposits on the southern rim of the Halimba Basin are gibbsitic with subordinate boehmite. Their average boehmite/gibbsite ratio is 16/84%. In the deeper lying Cseres deposits, the boehmite/gibbsite ratio is much higher at 95/5%. Finally, the deepest-lying deposit, the extensive stratiform one in the centre of the Halimba basin, is pure boehmitic.

Similar differences were observed by Komlóssy (1970) in the Iszkaszentgyörgy deposit group: these, however, are not controlled by the present topography. Komlóssy therefore correctly conjectured that the mineral distribution reflects the topography as it used to be at the time of bauxite accumulation (fig. 127). This would imply that, at the time of accumulation, the Kincses and József ore bodies lay highest (partly or entirely above the karst water table). It is these deposits that were leached most thoroughly by the prolonged percolation of rainwater, resulting in the lowest average kaolin content. The deeper were the deposits, the less intense was the percolation; the more kaolin was left behind. In the higher lying deposits, there is more gibbsite and goethite; in the lower lying ones, there is more boehmite and haematite. According to Komlóssy, the Eh was higher in the higher lying deposits, and the pH was slightly acidic. In the lower lying deposits, where ground water migration was slower, the Eh was probably lower and the pH higher. Komlóssy, then, assumed the mineralogical composition as we see it today to have developed at the time of bauxite accumulation, by the crystallisation of a semi-amorphous initial mass as controlled by the Eh and pH of the medium. Komlóssy does not exclude the further alteration of the minerals thus formed, e.g. by the gibbsitisation of boehmite if the bauxite came to be reworked in a higher lie. Komlóssy's interpretation differs from the present writer's only in that he restricts the evolution of the mineral composition as it is observed today to the deposits' period of accumulation, whereas the writer admits certain changes under the influence of subsequent mineralogical processes continuing in part up to this day.

In Hungary, the only purely diasporic deposit group is the one at Nézsa. It contains some gibbsite and boehmite in places, but nowhere more than 3%. In geological age and structural deformation, it does not differ from the rest of the deposit groups in the Transdanubian Hills, and even the environment of bauxite accumulation may have been the same. In a previous paper (Bárdossy 1961a), the writer attributed diaspore enrichment partly to structural stresses and partly to regional warming by nearby Miocene volcanism. More recent geological research has not confirmed this hypothesis. The writer therefore deems it more likely today that diasporic Nézsa bauxite is a product of redeposition of a wholly eroded older bauxite deposit, an assumption corroborated by the coarsely clastic conglomerate-breccia texture of this bauxite.

The Nagyharsány deposit in southern Hungary is also enriched in diaspore. The writer has calculated the deposit group to include 10% diasporic, 50% diaspore–boehmitic and 40% boehmitic bauxite. This deposit group is older than any other in Hungary; it is also the most deformed structurally (cf the relevant data in Appendix 1). The accumulation facies also differs from that of the Transdanubian Hills bauxites: Nagyharsány bauxite probably accumulated in the proximity of the ancient seashore, or in lagoons along it, under more reducing conditions. The bauxite of the deposit group in the Pădurea Craiului in Transylvania, Rumania, similar in age, deposit

features and facies, is also predominantly diasporic. The bulk of diaspore seems to be syn- or diagenetic in both deposit groups. Subsequent compressive stresses may also have contributed to an enhancement of diasporisation.

In France, in the bauxite region of the Provence, the average composition of the deposit groups changes gradually in a northeastern direction: Le Revest is pure boehmitic; Brignoles contains 80% boehmitic and 20% boehmito–gibbsitic bauxite; in the latter type, boehmite is predominant. Farther to the northeast, the deposits of the Haut-Var contain only 10% boehmitic and 80% boehmito–gibbsitic bauxite; the remaining 10% is pure gibbsitic. True, the present hilly relief gradually rises northeastward, but the deposits are strongly dislocated structurally: some are lifted high and others are downfaulted to some depth. The present altitude distribution, then provides no explanation, and the original one is well-nigh impossible to reconstruct. On the other hand, the stratigraphic hiatus in which the deposits are inserted is shortest at Le Revest and longest in the Haut-Var. The hiatus between the bauxite and its cover is non-existent or very short at Le Revest and Brignoles. In the Haut-Var, on the other hand, there is a distinct hiatus everywhere. These features provide circumstantial evidence that, at the time of deposition, it was the Haut-Var deposits that lay highest, and those of Le Revest deepest. The Eh–pH gradient invoked in the Transdanubian Hills, then, may have also been operative here, controlling the mineral distribution outlined.

According to Goldich and Bergquist (1947, 1948), the deposits of the Island of Hispaniola contain the less kaolinite and halloysite, the higher their altitude above the sea. The higher lying ones are richer in gibbsite and goethite: the deeper lying ones in boehmite and hematite. Goldich and Bergquist assumed the parent material of bauxite to have been uniform on the whole island of Hispaniola, as the bedrock limestone is itself uniform, and so were the conditions of bauxite accumulation. The differences referred to must therefore be due to altitude-dependent external factors acting subsequent to bauxite accumulation, such as average annual precipitation, which is the more copious, the higher the altitude. The relative height of the deposits above the ground water table may also have contributed; according to the authors previously cited, it likewise increases with altitude above sea level. In the higher lies, then, rains were more abundant and could better percolate the deposits. This resulted in the leaching of some of the kaolinite and halloysite and in the removal of their silica content. Some of the boehmite was converted to gibbsite, and some of the heamatite to goethite. These processes were interpreted by the authors cited in terms of a simple process of hydration, but in the light of the above examples the Eh–pH gradient may also have intervened.

The data of Anderson (1971) and Hill (1973) and the writer's personal observations imply that the mineralogical composition of Jamaican bauxite is controlled by the relief above all. Bauxite on the high plateaux contains little kaolinite, and its predominant iron mineral is haematite. In the grabens and on the low lying limestone plateaux, haematite in bauxite is gradually replaced by goethite. This is visible even to the naked eye: the red colour of the bauxite gradually passes into an ochroid yellow. Below about 200 m above the sea, bauxite grows increasingly clayey, and in the altitude range of 50 to 100 m, clayey bauxite is replaced by bauxitic clay. According to Lee (1964), this 'clay line' is at an altitude of 200 ft (about 60 m) in the St Elisabeth Plain. In addition to the

relief, however, bauxite composition was also influenced by the permeability of the underlying limestone. On top of the marly-clayey 'Montpelier Limestone', there is bauxitic clay rather than bauxite everywhere.

In the Boksonsk bauxite region of the USSR, the deposits in the northern, strongly dislocated area are diasporic and dipping 80 to 90° according to Terentyeva (1958). In the central and southern areas, where the deposits are less dislocated (dipping only 30 to 40°), the bauxite is diasporo–boehmitic, with diaspore dominant in some places and boehmite in others. This mineralogical gradient prevails over some 6 to 8 km. According to Terenteva, the more strongly dislocated deposits are also richer in diaspore in the Devonian deposit group of the Southern Urals, USSR, whereas the less dislocated ones are more boehmitic. In these two bauxite regions, then, a gradual change in mineralogy was controlled by a tectonic dislocation gradient.

As an example for the third group, let us consider the Turonian bauxite horizon of Central Greece. Nia (1968) states most of the deposits of this horizon to contain diasporic, mixed diasporo–boehmitic and boehmitic bauxites side by side. No gradual mineralogical changes acting over the entire bauxite region can be detected here.

5.2.4. CHANGES OF MINERALOGICAL COMPOSITION WITHIN THE DEPOSIT

In most bauxite deposits, changes in mineral composition can be observed both vertically, from the bedrock towards the cover, and also laterally. In the writer's experience, the most important difference tends to prevail between the oxidised and reduced parts of a deposit. Let us therefore compare these parts first of all.

5.2.4.1. A COMPARISON OF THE OXIDISED AND REDUCED PARTS OF A DEPOSIT

The reduced deposit portions are readily distinguished from the oxidised ones by their colour: the former are grey, green, black or off-white, whereas the latter are red, yellow, violet, or multicoloured.

The Cambrian deposits in the bauxite region of *Boksonsk*, USSR are red in their lower parts and greenish-grey in their upper parts. Both parts are diasporo–boehmitic in equal proportions; it is only the aggregate percentage of these two minerals that is lower in the reduced part of the deposit. In the oxidised part, the principal silicate mineral is kaolinite; it is accompanied by minor amounts of sericite and muscovite. The principal iron mineral is haematite. In the reduced part, there is no kaolinite or haematite; both are replaced by a leptochlorite accompanied by minor amounts of pyrite. In both parts of the deposit, rutile is the only titanium mineral. The essence of the change, then, is that the independent iron and silicate minerals of the oxidised part had combined to form ferrous silicates in the reduced parts. According to Orlova (1958), the red bauxite accumulated in fresh water under oxidising conditions. As a result of a slowly advancing ingression, the greenish-grey part of the deposit had already accumulated in a very shallow sea, but comparatively far offshore, under reducing conditions. The bauxitic material washed into the sea was reduced immediately on deposition; that is, the reduced minerals are syngenetic.

In the Devonian bauxite deposits of the *Northern Urals*, USSR, the lower, bulkier part of each deposit is red; the upper part is green, grey and black. The principal minerals in the oxidised part of the deposit are diaspore, boehmite, haematite, kaolinite, anatase and rutile. The reduced part of the deposit is lacking kaolinite and haematite also here. Those minerals are replaced by leptochlorites accompanied by siderite and minor amounts of pyrite. There are local enrichments of calcite or organic matter. The aggregate percentage of diaspore, boehmite, anatase and rutile is reduced, but no systematic change can be detected in their relative abundances. The changes, then, are similar to those at Boksonsk, and can be interpreted in a similar way. Bushinsky (1971) states the red bauxite to have deposited on land, under oxidising conditions in temporary fresh-water basins. Advancing transgression made the upper part of the deposit accumulate in the shore lagoons of an advancing sea. The bauxitic material that the lagoons received from the land was also red to start with: it was, however, soon reduced in the sea water under the influence of organic matter. The mineral association of the reduced deposit part, then, is syngenetic and early diagenetic.

In the Upper Devonian deposits of the *Southern Urals*, USSR, a similar situation can be observed. The only difference is that there the oxidised part of the deposit contains ferrichamosite, which is replaced by ferrochamosite in the reduced part. The conditions of origin were similar to those in the Northern Urals (Kirpal' and Tenyakov 1974).

Of the Triassic deposits, the one at *Vrace, Lika County*, Yugoslavia, also features some reduced bauxite, which is, however, at the bottom of the deposit (fig. 54). This greyish-green bauxite is diasporo–boehmitic in varying proportions. Its principal clay minerals are ferrichamosite and clinochlore, accompanied in places by minor quantities of kaolinite and illite. In the overlying oxidised part of the deposit, the aggregate abundance of boehmite and diaspore is higher; though their proportions keep fluctuating. Chamosite and clinochlore are both replaced by kaolinite and haematite accompanied by some goethite. Illite also disappears. In both parts of the deposit, anatase with somewhat less rutile represent the titanium association. According to Šinkovec (1970), the gentle karst depressions of the relief were filled with a clayey colloidal mud early on in the course of sedimentation. This gave rise to a lacustrine–swampy environment, in which the reduced part of the deposit could develop. The red part of the deposit overlying it came to be deposited under more oxidizing conditions to start with, and was further oxidized by an *in situ* phase of lateritization. Here, then, the process was the opposite of the above described case in Cambrian and Devonian deposits. The reduced mineral association is syn- and diagenetic in both cases.

The Lower Malm bauxite horizon near *Distomon* in the bauxite region of central Greece is made up differently. The deposits are elongate lenticular, two metres thick at most. Their middle and lower parts are dark red, passing gradually into a yellowish brown and then into a brownish grey upward and toward the deposit rims. Their mineral composition as established by the writer is presented in table 26.

Here, too, haematite and kaolinite are replaced by ferrochamosite and some small amount of pyrite in the reduced part of the deposit. As regards the aluminium minerals, however, there is an unmistakable enrichment of boehmite in the oxidised part and of diaspore in the reduced part of the deposit. The oxidised part was deposited on dry land, the reduced one was already in a lagoon. This can be inferred from the

Table 26. Mineralogical composition of the oxidised and reduced deposit parts in the Distomon deposits, Greece (weight percentages).

Part of deposit	Boehmite	Diaspore	Haematite	Goethite	Ferrochamosite	Kaolinite	Pyrite	Anatase	Rutile
Red	50·4	6·6	28·0	–	–	11·5	–	1·6	1·1
Yellowish-brown	46·2	–	18·6	6·0	–	25·9	–	1·2	1·1
Brownish-grey	6·3	43·2	–	0·8	40·0	5·3	0·5	2·4	0·6

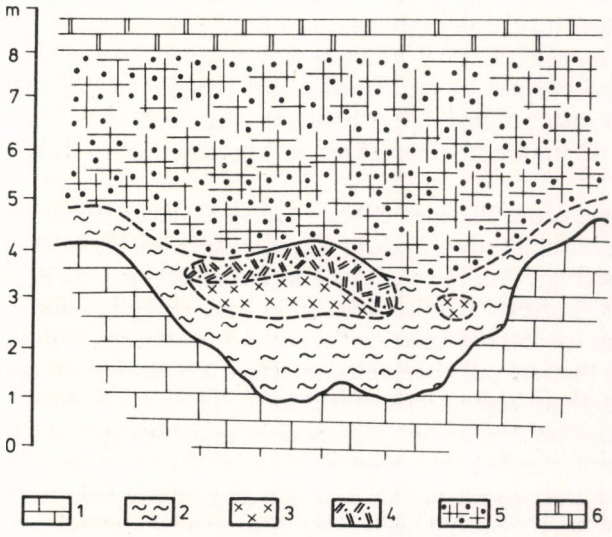


Figure 128. Geological profile of a bauxite deposit. Turonian bauxite horizon, Kiona Range, Grecee
(1) bedrock limestone; (2) brown, light brown bauxite; (3) grey pyritic marcasitic bauxite; (4) green, chamositic bauxite; (5) red, reddish-brown bauxite; (6) cover limestone

stratification of the reduced part. The immediate cover is a light grey marine limestone. The role of organic matter must have been less pronounced here. The writer holds the reduced mineral association to be also syngenetic here, early diagenetic at the latest.

In the same bauxite region, the overwhelming majority of deposits in the Turonian bauxite horizon are oxidised. In their middle and lower parts, the writer observed grey and green coloured oval spots in places of 0·5 to 3 m diameter (fig. 128). The surrounding red bauxite is diasporo–boehmitic, with now the one, now the other aluminium mineral predominant. The associated minerals are haematite, kaolinite, goethite, anatase, and rutile. Haematite and kaolinite are replaced by ferrochamosite in the green bauxite; but ferrochamosite disappears from the grey bauxite, to be replaced by pyrite, marcasite and reappearing kaolinite. The boehmite/diaspore ratio is varied; it exhibits no clear-cut regularity. The formation of these reduced spots was explained by Bárdossy and Mack (1967) on the assumption of local foci of reduction acting during the course of bauxite deposition, presumably as a result of enhanced microbial activity. These could reduce the bauxite when it was still soft: the process, then, could be early diagenetic. In the outer parts of such a zone, haematite and kaolinite united

into ferrochamosite. In those places, on the other hand, where the sulphurous proteins of the microorganisms gave rise to a high enough concentration of H_2S, pyrite and marcasite were formed preferentially, and kaolinite was unaffected.

In the top parts of these same lenses, there are also some rounded or oval light grey spots of 0·2 to 2 m diameter. Their reduced nature is indicated by 0·5 to 2% pyrite and a few tenths of a percent of organic matter. They contain a mere 1 to 2% haematite plus goethite, and just a few percent kaolinite. They are remarkably enriched in diaspore, however (up to 90%), and also in anatase and rutile. In the writer's opinion, reduction in these places was due to the influence of carbonaceous clay and organically rich limestone in the immediate cover, resulting in a complete removal of the reduced iron minerals. This is confirmed by the extraordinary porosity of these bauxites and their porous–vacuolar space-filling. These processes may well have been late diagenetic and/or epigene. There are no chamosites or leptochlorites in these reduced deposit parts.

The Lower Cretaceous bauxites of the *Pădurea Craiului*, Rumania consist in their bulk of oxidised, red and brown bauxite. There is also some greenish, less often grey or black reduced bauxite, one metre thick as a maximum at top of the deposits. According to Papiu *et al* (1970a, b), the oxidised part of the bauxite is predominantly diasporic, in places with somewhat less boehmite. In the reduced parts, there is more boehmite than diaspore; in fact, they are pure boehmitic in places. Kaolinite disappears; haematite is much reduced; both are replaced by leptochlorite. There is, on the other hand, a slight enrichment of goethite. The writer's own mineralogic investigations have revealed the presence of ferrochamosite in the greenish bauxites. The grey and black bauxites are enriched in pyrite and organic matter, and contain no ferric minerals at all. According to Papiu *et al* (1970a, b) reduced parts of the deposit were accumulated in a lacustrine, reductive environment as secondary conversion products of the initial red bauxitic material. The immediate cover is mostly a light grey, yellowish or reddish marine limestone. The present writer does not consider it probable for the red bauxite to have been reduced secondarily under the influence of this cover. He is of the opinion that, as transgression proceeded, the bauxitic material washed in from the land came to be deposited in a lagoon, to be reduced immediately by the copious organic matter available there. The writer, then, holds the reduced association to be syn- or early diagenetic.

The majority of the deposits in the *Transdanubian Hills* is made up of oxidised bauxite types. It is only at the top of the deposits overlain by Eocene strata that a grey reduced bauxite 0·5 to 2 m thick is found in places. Its immediate cover is a Lower or Middle Eocene carbonaceous–pyritic clay. There is a hiatus between it and the bauxite. Haematite and goethite, prevailing in the oxidised part of the deposit, are replaced in their bulk by pyrite and minor amounts of marcasite. The aggregate percentage of iron minerals decreases; that of kaolinite and of organic matter increases. There are no systematic changes of relative abundances within the aluminium-mineral group, but the aggregate percentage of the group is reduced. The overall abundance of the titanium minerals is unchanged, and so are their abundance ratios. No leptochlorite or chamosite had formed anywhere.

The reduction of bauxite was initiated by the deposition of the overlying swampy clay deposits as a late diagenetic or early epigene process. The reducing ground water of

the swamps, enriched in humic acid, dissolved the bulk of the haematite and goethite out of the bauxite. Part of the ferrous iron was removed in solution; the rest combined with the H_2S generated by the decomposing organic matter into pyrite or marcasite.

In the course of underground mine development in recent years, the writer has observed a reduced grey bauxite also near the bedrock in several deposits of the Nyirád group, Hungary (e.g. Dűlt-Nyires II and VII; Izamajor IV deposits). Its thickness is just 20 to 50 cm as a rule, very seldom attaining one metre. Its horizontal extent is also much restricted, to less than 10 m in most instances. In places, it overlies the bedrock dolomite directly. Elsewhere, it is separated from it by 10 to 50 cm of yellow or meat-red clayey bauxite or bauxitic clay. These grey bauxites agree in mineral composition with the grey bauxite at the top of the deposits.

In the Paleocene and Eocene bauxite deposits of the *Istrian Peninsula*, the writer has observed similar alterations wherever the immediate cover is a carbonaceous–pyritic clay or lignite. Haematite and goethite were partly reduced to pyrite and marcasite and partly removed in solution. Kaolinite and organic contents increased. No chlorite was formed here, either. The process here, a late diagenetic to early epigene reduction under a swamp, was analogous in all particulars to that which can be observed in the Transdanubian Hills.

A similar alteration can be observed in the Paleocene deposits at *Kruja in Albania*. There, too, a late diagenetic–early epigene reduction took place under the influence of swampy deposits in the immediate cover.

No reduced deposit parts are found in Neogene or Quaternary deposits. These have never been covered by swampy deposits: indeed, some of them have remained uncovered up to this day.

The above data warrant the following conclusions:

(i) Where bauxite has remained uncovered to the present, no reduced bauxite types have come to exist.

(ii) The initial bauxite washed in was reduced syngenetically and by early diagenesis wherever the accumulation of bauxite continued in lagoons and brackish embayments were formed by an incoming transgression.

(iii) Late diagenetic to early epigene reduction took place in the top parts of the bauxite deposits wherever the bauxite came to be overlain directly by swampy sediments (most often with a hiatus between them and the bauxite).

(iv) Local foci of reduction formed exceptionally also within the oxidised part of a deposit. The writer attributes these to reductive action by micro-organisms.

(v) Reduction affected the minerals of iron and the silicates first of all. Syngenetic and early diagenetic reduction combined haematite, goethite and kaolinite into chamosite and leptochlorite. This process may have been promoted by the weakly alkaline or neutral pH of the lagoons and by the ions dissolved in sea water.

Late diagenetic and epigene reduction could not break up the kaolinite lattice as it was fully crystalline already: nor was the acidic pH of the marshes favourable to this. Kaolinite could thus remain unaltered, so that no chlorites were formed. The high concentration of H_2S in the swampy environment, on the other hand, promoted the formation of pyrite and marcasite. The humic acid content of the solutions and their overall acidic pH may have had a significant role in these processes.

(vi) In the reduced deposit parts, iron minerals are less abundant, but clay minerals more so than in the neighboring oxidised parts. This can be explained by invoking a removal of iron in ferrous form simultaneously with the reduction and an import of silicic acid in solution (a resilification).

(vii) The aluminium minerals seem to have undergone no clear-cut changes. Their relative abundances among them were found to be invariant in several deposits. A diasporo–boehmitic composition was shifted in favour of diaspore in certain deposits, and in favour of boehmite in others. The issue is not fully understood and requires further detailed investigations.

(viii) Titanium minerals were not affected by the reductive processes.

5.2.4.2. CHANGES IN MINERAL COMPOSITION WITHIN THE OXIDISED DEPOSIT PARTS

Changes can be explored best by vertical and/or horizontal profiles or grid sampling. Systematic investigations of this nature have so far been few and far between. Let us review their findings in geochronological succession.

In the Cambrian deposits at *Boksonsk*, USSR, Orlova (1958) states bauxite to be of highest-grade above the depressions of the bedrock, where the deposit swells to 15 or 20 m thickness from an average of 4 to 5 m. The present writer attributes this to the fact that ground water percolation was strongest where the bedrock was deepest, and so was, in consequence, also the leaching of silica. Bauxite lowest in kaolinite is found at the bottom of the deposit; the kaolinite content gradually increases upward. The high grade ore at the bottom is diasporo–haematitic. Above it, in the more clayey bauxite, some boehmite appears besides the diaspore.

The red lower parts of the Devonian deposits in the *Northern Urals*, USSR, are pure diasporic, diasporo–boehmitic or boehmitic. The boehmitic bauxite is massive, compacted, jasper-like. The diasporo–boehmitic bauxite is also massive, of uneven fracture; it does not stain the hand. Diasporic bauxite is comparatively more porous and does stain the hand. Boehmitic bauxite is largely confined to the marginal parts of the deposit; diasporic bauxite is in the middle. Diasporo–boehmitic bauxite is found in between the two, in comparatively minor quantities, or at the top of the red deposit zone (fig. 129). The spatial arrangement of the three types of bauxite is completely independent of the textural and lithostructural features of the bauxite deposit. Bushinsky (1971) states that the passages from boehmitic to diasporic bauxite in the stratified parts of the deposits, from the rims towards the centres of the dolines are invisible to the eye. According to Beneslavsky (1963), the diasporic part is lowest in silica but comparatively rich in calcite; its diaspore is well-crystallised. He attributes this mineral association to a diagenetic process which had the alkaline–hydrocarbonate solution seeping through the deposit for its vector. Seepage was strongest above the sinkholes, where the bedrock is most strongly fissured and karsted.

Relationships very similar to these were demonstrated by Nia (1968) in the Turonian bauxite horizon of *Central Greece*. He laid sampling grids over whole bauxite faces, two in the Parnassos Mountains and one in the Helicon Mountains. He found the bauxite to be diasporic in the central, thickest part of the deposits and boehmitic nearer to the margins (fig. 130). Transitional, diasporo–boehmitic bauxite is comparatively scarce as

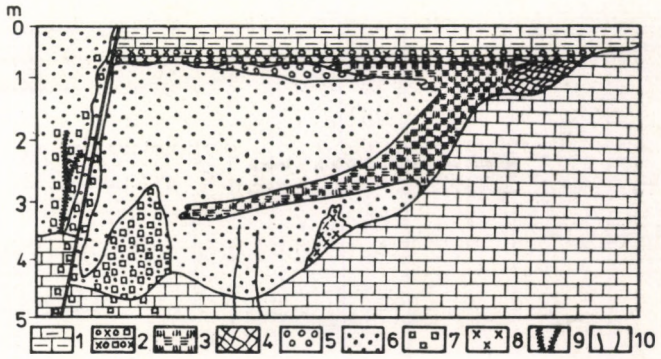


Figure 129. Lithological profile of a Devonian bauxite deposit in the Northern Urals, USSR. After Bushinsky (1971)

(1) immediate cover: black limestone and marl; (2) greenish-grey bauxite; (3) red, boehmitic bauxite; (4) multicoloured, boehmitic, clayey bauxite; (5) red boehmitic–diasporic bauxite; (6) red diasporic bauxite; (7) pyrite segregation; (8) chloritisation; (9) calcite veins; (10) sphalerite veins

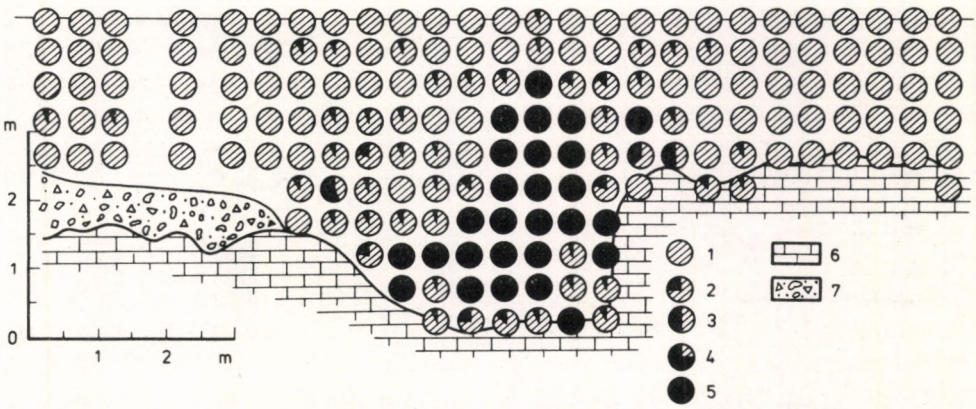


Figure 130. Relative abundance of diaspore and boehmite in the Kethro open pit, Turonian bauxite horizon, Parnassos Range, Grecee, according to Nia (1968)

(1) 100% boehmite; (2) 75% boehmite, 25% diaspore; (3) 50%boehmite, 50% diaspore; (4) 25%boehmite, 75% diaposre; (5) 100% diaspore; (6) bedrock limestone; (7) scree

compared with the two principal single phase-types. On passing from the boehmitic to the diasporic bauxite, the diaspore content gradually increases up to 56%, and then abruptly to 100%. In none of the diasporo–boehmitic bauxites studied by Nia was the diaspore content between 56 and 100%. He presumed the gradual diasporisation of boehmitic bauxite, postulating the reaction to continue spontaneously up to where all the boehmite had changed into diaspore once the diaspore/boehmite ratio had passed a threshold of 56/44%.

Both the pure boehmitic bauxite types and the diasporo–boehmitic ones are haematitic, but the diasporic ones are slightly richer in haematite. In the uppermost patrs of the deposits, bauxite turns goethitic, the total iron content diminished (fig. 131).

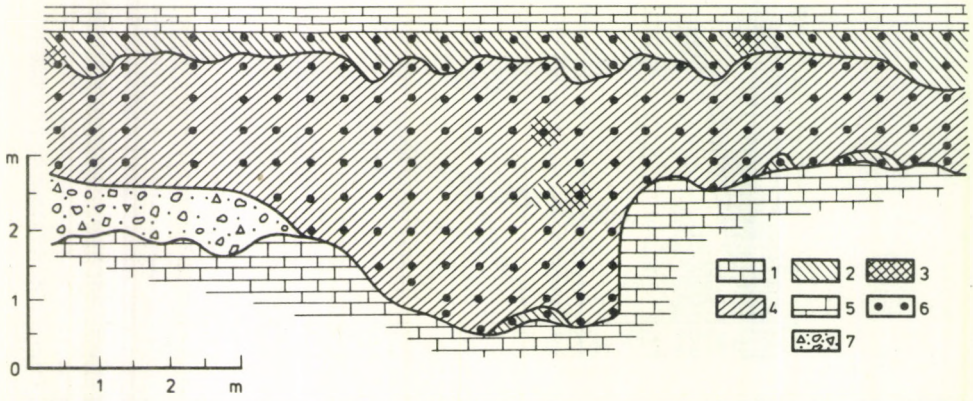


Figure 131. Distribution of iron minerals in Kethro open pit, Turonian bauxite horizon, Parnassos Range, Greece, according to Nia (1968)

(1) cover limestone; (2) goethitic bauxite; (3) goethito–haematitic bauxite; (4) haematitic bauxite; (5) bedrock limestone; (6) sampling locality; (7) scree

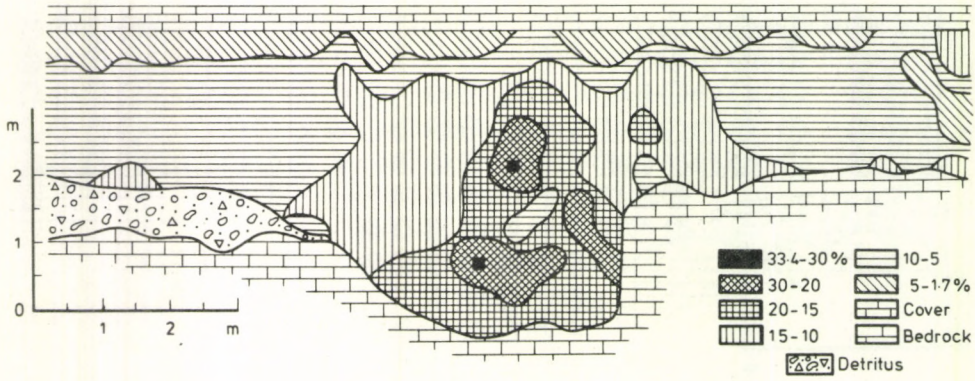


Figure 132. Porosity of bauxite in Kethro open pit, Turonian bauxite horizon, Parnassos Range, Greece, according to Nia (1968)

There is some goethitic bauxite also at the deposit bottoms. The diasporic part of the deposits is lowest in kaolinite; kaolinite content increases upward and towards the margins, to attain a peak at the deposit top. The porosity of bauxite is a function of mineral composition (fig. 132). It is 15 to 30% in the diasporic part, it is reduced to 10 to 15% in the diaspore–boehmitic and to 5 to 10% in the boehmitic part; finally, the clayey boehmitic bauxite of the deposit top has a porosity of 2 to 5% only. According to Nia, ground water seepage was strongest in the central parts of the deposits. It resulted in the diagenetic dissolution of part of the kaolinite and in the recrystallisation of boehmite into diaspore. Leaching resulted in the formation of micropores. The diaspore crystals and the recrystallised haematite together constituted a grid strong enough to maintain this increased porosity up to this day. In the top part of the deposit, haematite recrystallised into goethite before burial. The present writer, however,

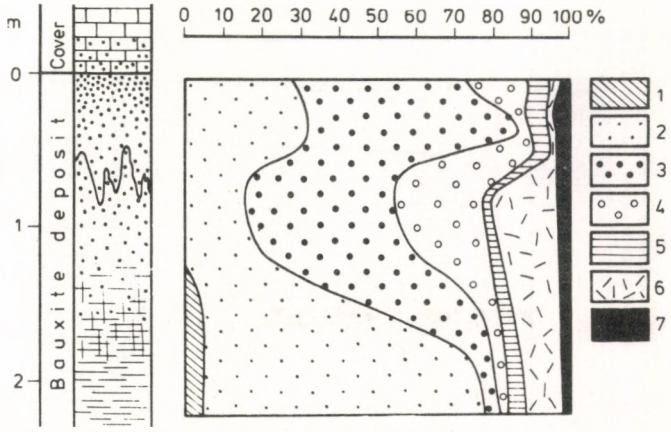


Figure 133. Lithologic and mineralogical profile of the Unjat II bauxite deposit, Ariège, France, according to Combes (1969)

(1) dioctahedral chlorite; (2) kaolinite; (3) boehmite; (4) diaspore; (5) goethite; (6) haematite; (7) anatase and rutile

prefers the explanation that this was an epigene process subsequent to burial. Enough is known to be certain that the process was unconnected with diasporisation.

The changes in the mineral composition and in the porosity of these bauxites exhibit a remarkable similarity to those in the Northern Urals deposits; they were most probably due to similar processes.

Changes of a different nature have been observed by the writer in a bedrock-to-cover profile of a Triassic bauxite deposit near *Kerman in Iran*, in samples put at his disposal by J. Sabot. His findings can be summarised in the following:

On top: more boehmite than diaspore, haematite, kaolinite, more anatase than rutile.

In the middle: more diaspore than boehmite, haematite, more chamosite than kaolinite, more anatase than rutile.

At the bottom: more boehmite than diaspore, haematite, more kaolinite than illite, more anatase than rutile.

Aluminium and titanium minerals are most abundant in the middle part of the deposit. Bauxite is dark red throughout; its chamosite is a ferrichamosite. The appearance of ferrichamosite in the middle section was accompanied by diaspore enrichment. This association is, in the writer's opinion, the product of syngenetic or early diagenetic processes. A slight drop in Eh may be assumed in the middle portions.

Combes (1969) studied the mineralogical composition of the Barremian–Albian deposits of the *Ariège, France* in a number of vertical sampling profiles. Those deposits developed from sedimentary clays, less seldom from clay marls, by *in situ* bauxitisation. The bottom of the deposits is kaolinitic; the kaolinite is accompanied by minor amounts of dioctahedral chlorite and illite. The clay mineral content gradually decreases upward, hand in hand with increasing contents of aluminium, iron and titanium. A comprehensive study of several profiles seems to imply that the diaspore/boehmite ratio is highest where the aggregate concentration of aluminium minerals is highest and kaolinite content is lowest (fig. 133). There are also some purely

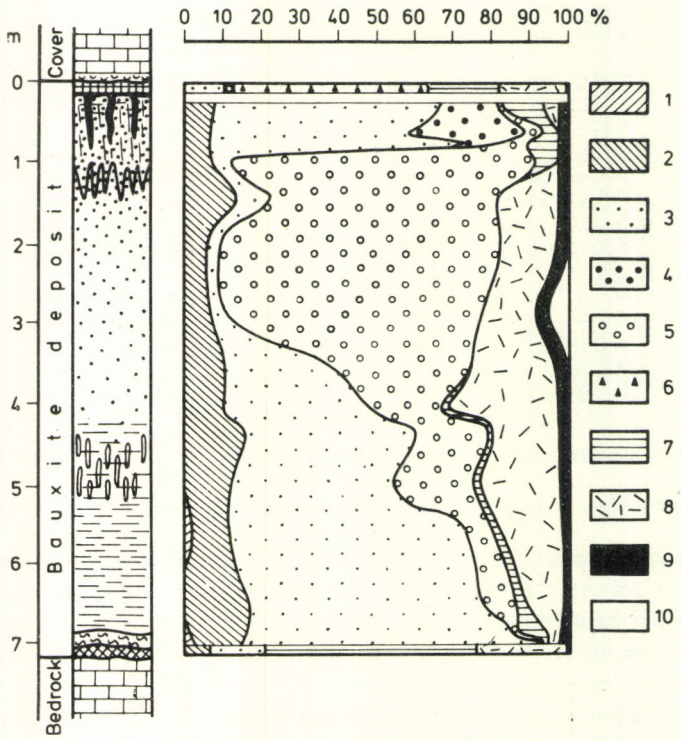


Figure 134. Lithologic and mineralogical profile of the 'Lescalé F' bauxite deposit, Ariège, France, according to Combes (1969)

(1) illite; (2) dioctahedral chlorite; (3) kaolinite; (4) boehmite; (5) diaspore; (6) gibbsite; (7) goethite; (8) haematite; (9) anatase and rutile; (10) calcite

diasporic deposits, containing a little boehmite only at the top and bottom, where the bauxite is more clayey. Goethite is enriched in the same places (fig. 134). Also in this case, then, the strongest desilication was most conductive to diaspore enrichment.

From the *Cazouls-les-Béziers* deposit of Languedoc, France, Combes (1969) described a peculiar case of threefold repetition in a deposit profile (fig. 135). The bottom of the profile is kaolinitic clay with haematite and goethite. The content of kaolinite gradually decreases upward, that of haematite, as gradually, increases. There is some boehmite at the top of this cycle. All this is repeated in the second cycle. Bauxitisation attains full development in the third cycle, as indicated by substantial concentrations of gibbsite, boehmite and haematite. The lowest contents of kaolinite tend to be accompanied by a few percent of diaspore, too. The profile top is entirely argillised; all aluminium minerals disappear, and goethite takes the lead among the iron minerals. Combes interpreted this threefold repetition in terms of three successive phases of bauxite accumulation and bauxitisation. A correlation, presumably controlled by diagenetic processes, between desilication and diasporisation is also observed in this instance.

The *Hungarian* bauxite deposits have been investigated by the present writer in several bedrock-to-cover profiles. In the Southwestern Bakony Hills, it is the deposit

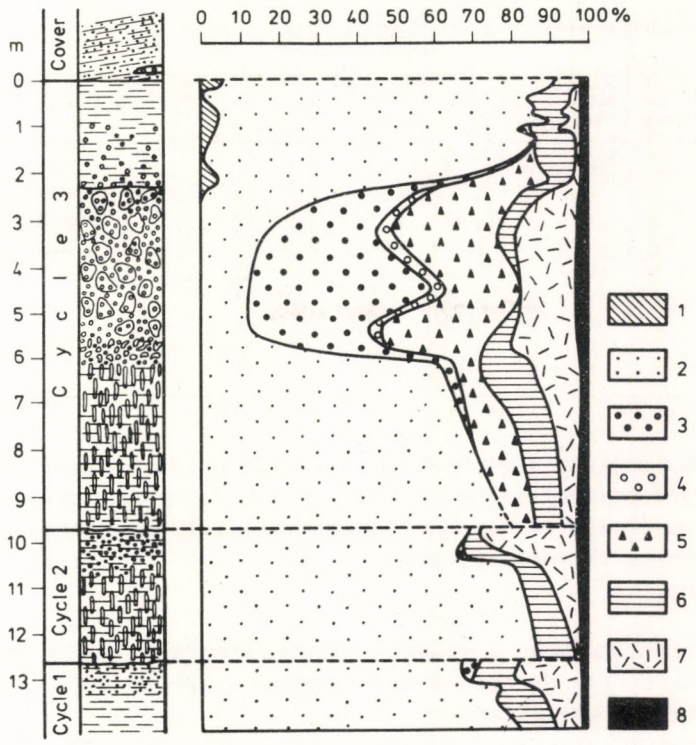


Figure 135. Lithologic and mineralogical profile of the Cazouls-les-Béziers bauxite deposit, Languedoc, France, according to Combes (1969)

(1) dioctahedral chlorite; (2) kaolinite; (3) boehmite; (4) diaspore; (5) gibbsite; (6) goethite; (7) haematite; (8) anatase and rutile

bottoms that are richest in clay minerals. The kaolinite content gradually decreases, and the contents of the aluminium, iron titanium minerals increase upward. Near the deposit top, the kaolinite content increases again. In the Nyirád deposit group, the lower deposit parts tend to be pure boehmitic; gibbsite appears in the middle and grows more abundant upward, but even at the top, boehmite is predominant. In the Szőc and Malomvölgy deposits, the bauxite is boehmito–gibbsitic at the bottom; it gradually changes into a pure gibbsitic type upward. In the Kislőd deposit, too, the gibbsite/boehmite ratio increases upward.

The changes in mineral composition from bottom to the top of the main deposit at Alsópere are presented in table 27, as based on the samples of a borehole denoted P-1. At the bottom of the profile, there is some illite accompanying the kaolinite, but it soon disappears upward. Here, too, gibbsite appears beside boehmite in the upper part of the deposit, and so does a small quantity of diaspore. Iron minerals are most abundant in the middle portion, which is also the least clayey. Near the top, haematite markedly decreases; goethite remains unchanged, and a little maghemite appears. A remarkable feature is the ferrichamosite identified in two of the samples from near the deposit top: this is something of a rarity in Hungarian bauxites. Near the deposit top, the bauxite matrix is impregnated with epigenetic calcite.

Table 27. Changes of mineral composition in the main deposit at Alsópere, Hungary, from bedrock to cover (weight percentages).

Depth interval m	Diaspore	Boehmite	Gibbsite	Haematite	Goethite	Maghemite	Kaolinite	Chamosite	Illite	Calcite	Anatase	Rutile
109·7–109·9	2	35	8	6	4	2	24	–	–	16	2	0·5
110·0–110·1	1	34	13	10	2	2	14	6	–	14	1·5	0·5
110·5–110·7	2	35	14	14	1	2	13	7	–	9	2	0·5
112·0–112·5	2	50	4	20	3	2	15	–	–	–	2	0·5
113·0–113·5	2	45	–	23	4	–	22	–	–	–	2	0·5
114·0–114·5	–	46	–	21	3	–	27	–	–	–	2	0·5
115·0–115·5	1	40	–	18	1	–	34	–	3	–	1·5	0·5
116·0–116·5	–	38	–	17	2	–	36	–	3	–	1·5	0·5
116·5–117·0	–	23	–	15	3	–	31	–	25	–	1·5	0·5

Goldlich and Bergquist (1948) state the deposits of *Haiti* to be more goethitic in their upper parts and more haematitic in their lower parts. According to them, downward-filtrating ground water rich in organic matter dissolved some of the haematite and reprecipitated it in the form of goethite.

In summary,

(i) Clay minerals are least abundant in the central parts of the deposits; they grow richer towards the top, the bottom, and the margins. Illite besides kaolinite appears near the bottom in several deposits.

(ii) The aggregate percentage of aluminium, iron and titanium minerals is as a rule contravariant to the clay mineral content (a maximum in the middle, decreasing both laterally and vertically).

(iii) The shifts in the aluminium-mineral associations within the deposits are regular rather than accidental. The regularity may differ from one deposit to the next; a more detailed clarification of abundance relations would require systematic profile and grid sampling and mineralogical analyses in each deposit group.

The analyses performed so far on diasporo–boehmitic deposits seem to imply an enrichment of diaspore in the strongly desilicified parts of those.

In the boehmito–gibbsitic deposits, gibbsite concentration tends to increase from the bottom upward, to the detriment of boehmite. In the lateritic bauxite deposits of similar composition, it is generally the quantity of boehmite that tends to increase upward, for example in Guinea (Balkay and Bárdossy 1967), and also in Australia, in the Weipa, Gove and Gippsland deposits (Grubb 1973).

(iv) In several deposit groups, the gibbsitic parts of the deposits tend to be richer in goethite; the boehmitic and diasporic ones tend to be richer in haematite.

(v) Near the deposit top, the total iron-mineral content declines as a rule. This is accompanied by a shift of the haematite/goethite ratio in favour of goethite.

(vi) The most permanent features of the deposits are the total content and relative abundances of the titanium minerals. Rutile and anatase tend to be enriched in those parts of the deposits which underwent a diagenetic or epigene deferrification. This is an apparent, relative enrichment only, a result of the removal of the iron minerals.

5.2.5. RELATIONSHIP BETWEEN MINERAL COMPOSITION AND TEXTURE

The mineral composition of a bauxite tends to undergo certain typical changes at the boundaries of and within the textural elements. In the following, we shall always assess these changes by comparing the mineral composition of the textural element considered with that of the surrounding matrix. The justification for doing so is that the mineralogical interactions that we are concerned with tend to take place between the matrix and the textural elements embedded in it.

5.2.5.1. MATRIX

Even though it is made up of well-defined mineral grains, a pelitomorphic and microgranular matrix may, in view of its very fine particle size, be regarded as a statistically homogeneous mineralogical system. Any changes in its mineral composition tend to be slow and gradual. The matrix consists in its bulk of the rock-forming minerals of bauxite. Some of these (for example the various clay minerals) attain their highest concentrations precisely in the matrix. Others (for example diaspore, corundum, siderite) tend to concentrate in other textural elements.

5.2.5.2. AUTHIGENIC MINERAL GRAINS

These are diagenetic or epigene mineral grains coarser than the matrix minerals. Early diagenetic anatase and rutile grains of 2 to 100 μm in size are the most widespread. They had formed during the slow crystallisation of the matrix, but their particle size is several times that of the surrounding minerals, thanks to their superior speed of crystallisation. Haematite and goethite grains of similar size are rarer; they are late diagenetic as a rule.

In the Mas Rouge deposit of the French Alpilles, Caillère and Pobeguin (1967) observed pseudohexagonal gibbsite crystals of several hundred micrometres in size (fig. 136), which they considered to be of diagenetic origin. Gibbsite crystals of a comparable pseudohexagonal form and 20 to 100 μm in size were described by Bárdossy (1961c) from the bauxite deposits of Sümeg, Hungary. In a few cases, the writer also observed gibbsite twins of diagenetic origin, touching so as to include an angle of 60°.

In the Devonian deposits of the Northern Urals, USSR, acicular or columnar diaspore crystals of 10 to 50 μm in size are frequent. Clay minerals rarely form large authigenic crystals (Gutkin 1978). There are early diagenetic kaolinite crystals of 10 to 1500 μm in size and of vermicular form in the Vrace deposit, Lika Country, Yugoslavia (Šinkovec 1970). Vermicular kaolinite crystals of several hundred micrometres in length were observed by Papiu et al (1971) in a deposit at Haţeg, Rumania.

The epigene mineral grains are larger as a rule than the diagenetic ones. For example, many deposits contain calcite and pyrite crystals of several hundred micrometres to a centimetre in size. They are also frequent in the Hungarian deposits. Near the top of the Gánt deposits, there are late epigene idiomorphic gypsum crystals of 1 to 3 cm in size.

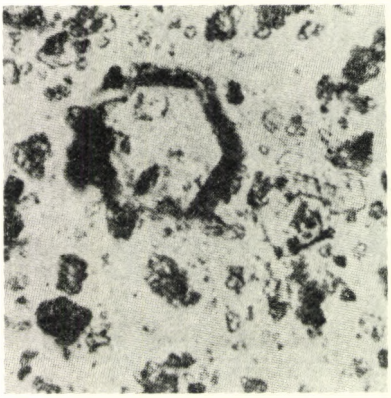


Figure 136. A pseudohexagonal authigenic gibbsite grain in the bauxite matrix. Mas Rouge open pit, Alpilles, France. After Caillère and Pobeguin (1967). Thin section in liquid immersion, 200 ×

Several rare minerals of karst bauxites occur exclusively as authigenic crystals, such as the chalcopyrite grains of the Devonian deposits of the Northern Urals, USSR (fig. 72).

5.2.5.3. PSEUDOMORPHS

They are comparatively rare in karst bauxites. It is the pseudomorphs of goethite and haematite after pyrite that are the most frequent. A case in point, due to the slow epigene oxidation of pyrite, was described by Beneslavsky (1963) from the Devonian deposits of the Northern and Southern Urals, USSR. In one of the small deposits near Cusano Mutri in the Monti Matese, Italy, the writer has found a pseudomorph of boehmite after pyroxene (fig. 73). In addition to the characteristic outline of the crystal, the pseudomorph even copied the cleavage pattern of pyroxene. Beneslavsky (1963) mentions pseudomorphs of gibbsite after quartz occurring in a few places in the deposits of the Western Turgay, USSR.

5.2.5.4. CLASTIC MINERAL GRAINS

They are hard minerals as a rule, resistant to both weathering and wear, found in quantities ranging from a few thousandths to one tenth of a percent in most karst bauxite deposits. They may be concentrated for study by elutriation in the particle size range 0·06 to 0·2 mm. The most widespread minerals in that size range include haematite, quartz, zircon, rutile, tourmaline, corundum, ilmenite, magnetite, disthene, and staurolite. The feldspars, muscovite, epidote, zoisite, spinel, glaucophane, beryl, garnet, and titanite are rarer. Particles of more or less weathered biotite have also been identified in a few deposits.

Šinkovec (1977, oral communication) found particles of detritic gibbsite in the Paleocene–Eocene bauxites of Croatia, which he attributed to the erosion of some older bauxite deposits.

In the course of the electron microprobe investigations which the writer has carried on jointly with György Pantó, they have identified clastic mineral grains far below the

60 μm size limit. These include much the same mineral species as those above 60 μm, but some minerals never before demonstated by microscopic means have also been found. The smallest such particles are of just one or two micrometres in size. The most frequent mineral species are monazite, xenotime and cheralite; galena, sphalerite and chromite are rarer.

The aggregate abundance of clastic mineral grains, the composition of their association, and their particle sizes are most helpful in determining the parent material of bauxite and the pattern of its transport. In the bauxites of the Provence, Southern France, for example Bárdossy and Pantó (1973) demonstrated a clear-cut relationship between distance from the igneous-metamorphic basement outcrops on the one hand and the numbers, the average and largest sizes of monazite, xenotime, zircon, and chromite grains embedded in the bauxite on the other.

5.2.5.5. CLASTIC BAUXITE PARTICLES

In the oxidised deposit parts, only bauxite particles containing oxidised mineral associations have so far been found. The writer attributes this to the fact that no lumps of reduced bauxite could conceivably stand up to surface weathering and reworking. The reduced deposit parts contain, in addition to clastic bauxite particles reduced *in situ* by secondary processes, also bauxite particles of oxidised nature. These could, thanks to their greater compactness, stand up better to the reductive influences than could their surroundings.

Clastic bauxite particles contain more haematite and goethite as a rule than the surrounding matrix, the reason being that bauxite richer in iron is harder and thus more resistant to transport and reworking. The composition of the aluminium-mineral association is independent of matrix, in fact random.

5.2.5.6. BAUXITE PEBBLES AND FRAGMENTS

Our statements concerning the clastic bauxite particles hold also for these. Certain bauxite pebbles are so rich in iron minerals as to constitute a bauxitic iron ore or, indeed, a sedimentary iron ore. They consist largely of haematite, or of haematite plus goethite. Identifying such elements in the French Alpilles and in the Spanish Pyrenees, the writer has derived them from the ferricrete crusts ('cuirasses') of eroded laterite profiles. Their hardness and toughness makes for a preferential enrichment in the course of transport. One of the arguments in favour of this assumption is that the collomorphic textures of these pebbles much resemble the textures of known lateritic ferricretes.

Bauxite pebbles often have a dark brown shiny goethite film a few tenth of a millimetre thick. It had presumably formed immediately after the deposition of the hard and compact pebbles, precipitating on their surfaces from a ground water seeping through a still permeable matrix. According to Nicolas (1970), there are bauxite pebbles traversed by fine veins of calcite of 1 to 3 cm width in the upper parts of the Mazaugues deposit, Brignoles, France. The veins do not pass into the surrounding matrix, thus implying the calcitisation of the pebbles to have taken place prior to their deposition.

Bauxite pebbles, thanks to their greater compactness, were less affected by the epi- and supergene processes than the matrix surrounding them: they could preserve their original composition better. In those places, for example where the upper parts of a deposit came to be deferrified, most of the pebbles retained their original haematite and goethite content. In such cases, the dark red bauxite pebbles are sharply distinct against the background of a faded-out matrix. In Hungary, this is especially conspicuous in the top part of the Kislőd deposit. At Halimba, Nyirád, Szőc and Iszkaszentgyörgy, some bauxite pebbles in the grey pyritic bauxite have preserved their original haematito-goethitic composition and their corresponding red colour.

5.2.5.7. BAUXITE BOULDERS

The above considerarions concerning the pebbles hold also for the boulders. A gradual concentration of haematite and goethite can be observed in the outer zones of some exceptionally large boulders. The writer interprets this phenomenon by the assumption that the bauxite was periodically saturated with water and dried out while the boulders lay about the surface after redeposition. In such boulders, water migrating towards the surface by capillarity could dissolve part of the iron minerals. As the water evaporated from the surface, the dissolved iron precipitated in the form of secondary haematite and goethite.

5.2.5.8. ROUNDGRAINS

The considerations concerning the bauxite pebbles also hold here. Roundgrains tend to be enriched in haematite and goethite above all.

5.2.5.9. PARTICLES OF CONCENTRIC (ORBICULATE) STRUCTURE

In the mineralogical respect, the writer could discern no significant difference between micro-ooids, ooids, pisoids and macropisoids. By comparing their mineralogy with that of the embedding matrix, four groups can be distinguished.

(i) *The particles are enriched in aluminium minerals.* This is a comparatively rare case. In the Carboniferous deposits of Missouri State, US, Allen (1935) found the ooids to be enriched in diaspore and the matrix in kaolinite, anatase and rutile. Some ooids contain 90 to 95% diaspore. The bauxite there is low-iron, containing a mere 1 to 2% of goethite and pyrite. The ooids contain even less iron as a rule. According to Keller at al (1954), the ooids contain no boehmite. The bauxite is pelitomorphic wherever it is boehmitic. The clayey bauxite types are richer in boehmite. The ooidic bauxite is more porous than the pelitomorphic variety. The authors ascribe the formation of diaspore to reducing conditions. They deem it possible for the boehmite to have recrystallised into diaspore in the course of ooid formation.

The pisoids of Ariège bauxite are enriched in diaspore and ferridiaspore according to Caillère and Pobeguin (1965). Some pisoids consist exclusively of highly crystalline diaspore. The matrix in which these pisoids are embedded is diasporo–boehmitic; it contains haematite and less goethite. Caillère et al (1971) made similar observations on

the pisoids and macropisoids of the bauxite deposit near Cadarcet. Some of the pisoids wear a secondary gibbsite film, or are traversed by gibbsite-filled veinlets. Combes (1969) determined the compositions shown in table 28 for the pisoids and the matrix of a reddish-brown bauxite in one of the deposits.

Table 28. Mineral composition of pisoids and matrix in a bauxite from the Cadarcet deposit, France (weight percentages).

Textural elements	Diaspore	Boehmite	Total Al minerals	Haematite	Goethite	Total iron minerals	Kaolinite	Dioctahedral chlorites	Total clay minerals
Matrix	14·5	31·5	46·0	14·7	2·7	17·4	26·6	9·9	36·5
Pisoids	77·0	6·4	83·4	11·4	–	11·4	5·1	–	5·1
Concentration coefficient	5·3	0·2	1·8	0·8	0	0·7	0·2	0	0·1

Diaspore is seen to be enriched more than fivefold in the pisoids, to the detriment of all other mineral species, especially the clay minerals.

In the Turonian bauxite horizon of Central Greece, Nia (1968) found the boehmite in some of the pisoids and ooids to have recrystallised into diaspore as a secondary process. The process entailed the dissolution and removal of some of the kaolinite, haematite and goethite, giving rise to fine micropores in the pisoids and ooids. Recrystallisation started in the nucleus and progressed outward. The orbiculate structure was largely effaced in the process. This phenomenon is most widespread in the diasporic parts of the deposit, where the matrix had also been diasporised.

(ii) *Aluminium and iron minerals both enriched in the particles.* This is the most frequent case in karst bauxites. Aluminium and iron enrichment alternate in successive shells. For an example, consider table 29, presenting the composition of macropisoids in a bauxite from a deposit near Fuentespalda, Maestrazgo deposit group, Northeastern Spain (table 29).

Table 29. Mineral composition of matrix and pisoids from Fuentespalda bauxite deposit, Maestrazgo, Spain (weight percentages).

Textural elements	Diaspore	Boehmite	Gibbsite	Total Al minerals	Haematite	Maghemite	Total iron minerals	Kaolinite	Illite	Anatase + Rutile
Matrix	1	4	48	53	11	–	11	29	2	2
Pisoliths	48	7	10	65	14	8	22	9	–	2
Concentration coefficient	48·0	1·7	0·2	1·2	1·3	∞	2·0	0·3	0	1·0

The macropisoids are seen to be most enriched in iron minerals, somewhat less so in aluminium minerals. Of the aluminium minerals, the concentration of diaspore has increased most, that of boehmite somewhat less, and that of gibbsite has actually decreased. Of the iron minerals, maghemite is found only in the macropisoids. The concentration of these two mineral groups is offset by a certain depletion of the clay minerals. The concertration of titanium minerals is unaffected.

A similar situation has been observed by the writer in the top part of one of the deposits of the Alsópere group in Hungary (table 30). The matrix there has a light greyish brown colour, the pisoids are dark brown.

Table 30. Mineral composition of pisoids and matrix from Alsópere bauxite, Hungary (weight percentages).

Textural elements	Diaspore	Boehmite	Gibbsite	Total Al minerals	Haematite	Goethite	Maghemite	Chamosite	Total iron minerals	Kaolinite	Calcite	Anatase	Rutile
Matrix	1	35	14	50	5	2	1	–	8	15	23	2	1
Pisoids	20	25	9	54	35	–	5	3	43	1	–	1	0·5
Concentration coefficient	20·0	0·7	0·6	1·1	7·0	0	5·0	∞	5·4	0·1	0	0·5	0·5

There, too, it is the iron minerals that are most enriched; the aluminium minerals are less so. The relative enrichment of the aluminium minerals is the same as in the Fuentespalda case. Of the iron minerals, haematite and maghemite are enriched; chamosite is found only in the pisoids. The concentration of the iron and aluminium minerals is offset by a depletion of kaolinite, rutile, anatase, and calcite.

According to Malakhovsky and Lysenko (1964), iron-mineral concentration is accompanied by the appearance of diaspore in pisoids of the Crimean bauxites. The same phenomenon was observed by Caillère and Pobeguin (1965) in a bauxite deposit near Péreille in the Ariège. In both sites, concentration had taken place at the expense of kaolinite. According to Kiskyras (1958), in the pisoids of Eocene bauxite on the Peloponnese Peninsula, Greece, diaspore is enriched together with magnetite. Slatkine and Heller (1961) state that the pisoids of the Maktesh Ramon deposits, Israel, have a higher diaspore/boehmite ratio, the greater their haematite content. They deem diaspore formation to have been promoted by the nucleating influence of haematite. The aluminium and iron minerals' enrichment is also offset here by kaolinite dilution.

Ooids and pisoids may form also under reducing conditions. This was proved by among others the observations of Jurković (1965) on greenish-grey pyritic bauxites in the Suvaja-Šolaja deposits, Yugoslavia. He found the darker shells of the ooids and even most of the nuclei to be enriched in pyrite and organic matter. These ooids agreed in all structural details with the ooids contained in the red bauxites of the deposit group, the only difference being in mineral composition. These ooids formed under a reducing Eh, brought about by decomposing vegetable debris and H_2S. On the other hand the secondary reduction of pre-existing ooids had led to a partial or total effacement of the concentric shelly structure.

Densely ooidic green, greyish-green and dark grey bauxites are frequent also in the Triassic bauxites of Lika County and Slovenia, Yugoslavia. The present writer's field observations have permitted him to infer that the ooids in those deposits had also formed in reducing surroundings.

(iii) *The particles are enriched in iron minerals only.* This case is somewhat less frequent than the foregoing one. For an example, the dark red ooids and pisoids of the Campo Felice bauxite deposit, Italy, are presented (table 31). The haematite content is seen to

Table 31. Mineral composition of matrix and pisoids in bauxite from Campo Felice, Italy (weight percentages).

Textural elements	Diaspore	Boehmite	Total Al minerals	Haematite	Kaolinite	Calcite	Anatase	Rutile
Matrix	–	56·0	56·0	19·4	19·7	1·1	2·1	1·0
Ooids and pisoids	0·6	34·6	35·2	56·3	6·0	–	0·6	0·4
Concentration coefficient	∞	0·6	0·6	2·9	0·3	0	0·3	0·4

have increased almost threefold. All other minerals are depleted except diaspore, which is confined to the ooids and pisoids.

The writer has observed similar mineral distribution in pisoidic bauxite from Vlasenica, Yugoslavia. In the sample studied, the dark brown pisoids are embedded in a pink matrix (table 32). The nature of the changes is much the same as in the foregoing example, the only difference being the presence of small quantities of illite in the Vlasenica pisoids.

Table 32. Mineral composition of matrix and pisoids from Vlasenica bauxite, Yugoslavia (weight percentages).

Textural elements	Diaspore	Boehmite	Total Al minerals	Haematite	Maghemite	Total iron minerals	Kaolinite	Illite	Anatase	Rutile
Matrix	1	71	72	12	2	14	10	–	2·5	0·5
Pisoids	6	53	59	30	2	32	5	2	1·0	0·2
Concentration coefficient	6·0	0·7	0·8	2·5	1·0	2·3	0·5	∞	0·4	0·4

Mineralogical alterations similar to those outlined above are also frequent in the Hungarian bauxites. In Eplény bauxite, the pisoids are enriched in goethite wherever the surrounding matrix is goethitic and in haematite wherever it is haematitic (Bárdossy 1968). Hence, the pisoids were not enriched in just any mineral of iron, but invariably in the one which had been most abundant in the matrix to start with. Despite a reduction in the total aluminium-mineral content, an enrichment of diaspore from a prior range of 1 to 3% to a range of 5 to 6% is observed. Kaolinite is substantially depleted in the pisoids.

One bauxite sample from the József III ore body, Iszkaszentgyörgy, Hungary contained macropisoids of 5 to 15 mm in size. These had a dark brown outer shell 1·0 to 1·5 mm thick. Within it, lighter, russet shells followed. Further inward, there was a vivid red, finely orbiculate part. These shells surrounded a dark brown nucleus of 2 to 3 mm in size. For their mineral composition cf table 33.

In the macropisoids, the total iron mineral content is seen to be three to six times that in the matrix. All other minerals are depleted. Goethite is predominant in the outer shell and in the nucleus; haematite is predominant in all the other shells. The

Table 33. Mineral composition of matrix and pisoids from Iszkaszentgyörgy bauxite, Hungary (weight percentages).

Textural elements	Boehmite	Gibbsite	Total Al minerals	Haematite	Goethite	Total iron minerals	Kaolinite	Anatase	Rutile	Crandallite
Yellow matrix	53·3	26·0	78·3	–	10·3	10·3	6·0	2·0	0·6	1·0
Dark brown outer shell	20·9	16·9	37·8	14·8	39·2	54·0	4·6	0·9	0·4	–
Russet shells	34·2	20·8	55·0	20·2	14·5	34·7	5·9	1·5	0·5	0·3
Vivid red shells	29·4	18·0	47·4	34·9	9·4	44·3	4·4	1·3	0·5	–
Dark brown nucleus	19·8	17·1	36·9	8·1	51·0	59·1	0·5	1·2	0·4	–

boehmite/gibbsite ratio is highest in the parts richest in haematite (1·64), and lowest in those richest in goethite (1·16). The haematite–boehmite and goethite–gibbsite associations, referred to above, are thus seen to prevail also within the pisoids.

(iv) *The particles are enriched in gibbsite and corundum.* This phenomenon has so far been observed exclusively in the Kazakhstan-type deposits of Siberia, Central Kazakhstan and the Turgay Platform, USSR. In the upper parts of the deposits along the Angara River, the pisoids contain 23 to 51% corundum, 9 to 25% gibbsite, 2 to 5% magnetite, 15 to 30% haematite, 1 to 5% goethite and 5 to 8% kaolinite. The matrix contains no corundum or magnetite, but is richer in kaolinite, anatase and rutile (Medvedev 1971). Pisoids in the hard bauxites at Tatarsk deposit group contain up to 16% corundum accompained by gibbsite, magnetite, maghemite, haematite, goethite and very little diaspore (Pel'tek 1971).

The brown-black pisoids and macropisoids in the Arkalyk bauxites contain corundum, gibbsite, haematite, magnetite, maghemite, some kaolinite, boehmite, anatase and rutile (Lisitsyna and Pastukhova 1963). The pisoids in the clayey bauxite types were found to contain 17 to 30% cryptocrystalline corundum; those in the hard blocks of high-grade ore, from 24 to 48%. The authors cited presume that some of the iron ions of the haematite lattice were isomorphously replaced by aluminium ions out of aluminium-rich solutions, and that the mixed crystals so formed continued to grow as corundum crystals. According to Beneslavsky (1963), on the other hand, corundum is a product of alumogel ageing, under conditions excluding hydration. In the deposits of the Turgay Plateau, too, pisoids and macropisoids of such composition are frequent (Venkov 1968, oral communication).

No other changes in mineral composition outside the above four groups have so far been recognised in any concentric particles. The following conclusions of general validity apply:

(i) Concentric particles invariably contain less kaolinite than the surrounding matrix.

(ii) If the surrounding matrix contains some calcite, the concentric particles contain very little or none.

(iii) In most cases, it is the iron minerals that are enriched; the chances are that the mineral most enriched in the concentric particles is the one most abundant in the surrouding matrix. Maghemite is invariably enriched: magnetite is enriched only in the pisoids of corundo–gibbsitic composition.

(iv) It is not often that the aluminium minerals are strongly enriched; they are, however slightly enriched in most cases and may even be depleted if the enrichment of iron minerals is very strong. The greatest enrichment is that of diaspore, whose abundance tends to increase even where the sum of the aluminium minerals decrease. Boehmite contents may either increase or decrease somewhat: gibbsite contents decrease in a majority of cases. Corundum together with gibbsite is enriched in a few Kazakhstan type deposit groups only.

(v) The relative abundances of the titanium minerals remain unaffected. Their total abundance is decreased in most cases in the concentric particles, unchanged in the remaining few ones.

(vi) Concentric particles formed under reducing conditions tend to be enriched in pyrite, organic matter and in aluminium minerals with a concomitant depletion of clay minerals.

5.2.5.10. FILMS AND COATINGS AROUND PRE-EXISTING TEXTURAL ELEMENTS

These usually consist of pure goethite, including for example the films shown in figures 91 and 92. There are some much rarer black manganiferous coatings, for example on bauxite pebbles in a deposit near Hrast in Slovenia, made up of a mixture of lithiophorite and todorokite. Bushinsky (1958) described a diaspore coating 10 to 30 μm thick on authigenic pyrite grains from the Devonian Vagran deposit of the Northern Urals, USSR.

Films and coatings are late diagenetic or epigene in origin. They tend to consist of iron and manganese minerals because, of all the chemical components of bauxite, those two are the most mobile ones.

5.2.5.11. PORE FILLINGS

Mono-mineralic as a rule, these consist most often of gibbsite, diaspore, calcite or siderite. Diagenetic or epigene in origin, they invariably are more coarsely crystalline than the surrounding matrix, as proved among others by the gibbsitic pore fillings of figure 93. In some samples from the bauxite deposits of the Southeastern US, columnar gibbsite crystals of 100 to 500 μm in size projecting inward from the pore walls were observed.

Deposits of secondary resilication tend to include minute white pore fillings of highly crystalline kaolinite. Cases in point have been observed by the writer in the upper parts of some of the Brignoles deposits, France, in the La Llacuna deposit, Ebro Massif, Spain, and in the Paramun deposit near Tran, Bulgaria.

5.2.5.12. IRREGULAR SEGREGATIONS

Mostly mono-mineralic, less seldom made up of two or three phases, they are late diagenetic or epigene. Their most abundant minerals are haematite and goethite.

In the epigenically reoxidised parts of the deposits at Nyirád, Szőc, Halimba and Iszkaszentgyörgy, Hungary, there are frequent dark red or reddish-brown segregations

of haematite and goethite, of a few millimetres or at most one centimetre in size. It is apparent in the thin sections that these do not fill pre-existing pores or cavities: on the contrary, they have impregnated the bauxite by gradually displacing its pre-existing minerals.

The writer has observed an extraordinary abundance of secondary goethite segregations in some bauxite samples from deposits near Bohinj in Slovenia. These segregations are also due presumably to the oxidation of pyrite.

5.2.5.13. NESTS, CONCRETIONS

Most often mono-mineralic, less seldom composed of two or three minerals, they are of late diagenetic and epigene origin.

The oxidation of pyrite may produce nests of pure *alunite*. In Hungary, such nests used to be frequent in the top parts of the Kincses ore body at Iszkaszentgyörgy and in the open pits of Deáki Hill at Nyirád. In the deposit Izamajor III, the writer has found a few bag-shaped nests of 30 to 80 cm in diameter of white alunite; mammillary and reniform as a rule. Nests of *aluminite* are much rarer. In the deposit Nyireskút II at Szőc, the writer has observed a few nests, of 5 to 20 cm in size and of reniform appearance, of snow-white aluminite. Aluminite nest of 2 to 4 cm in size from the deposit Félix II at Szőc was described by Bárdossy and Sajgó (1968).

Vivid red concretions of *haematite* and *goethite* are frequent as replacements in oxidised pyritic bauxite. They are most often irregular, more seldom plate-like. The plates, 2 to 6 cm thick, are embedded horizontally. They are especially frequent in the deposits at Szőc, Malomvölgy and Iszkaszentgyörgy. Bauxite at Nyirád and around Sümeg tends to contain concretions of pure goethite, of spherical, ellipsoidal or cylindrical form and radiated internal structure (fig. 95). Such concretions used to be especially frequent in the open pit at Darvastó VI deposit.

Some deposits contain nests made up of aggregates of *calcite* crystals. The writer has found them to be especially frequent in the Spinazzola deposits of Apulia, Italy. The calcites there are crystal-clear and larger than a centimetre in size.

The above nests are all compact, whereas those made up of pure *gibbsite* are loose and porous as a rule. The products of pyrite oxidation, they are found in sizes of 2 to 5 cm in grey pyritic bauxite or in its reoxidised analog in the deposits at Nyirád, Szőc, Halimba and Iszkaszentgyörgy, Hungary. In Albania, in the Krasta Pengile deposit near Kruja, Hungarian geologists found friable snow-white nests several centimetres in size in multicoloured bauxite underlying a pyritic-marcasitic bauxite. The writer's X-ray analyses have demonstrated these too to consist of pure gibbsite.

The tubular concretions formed along root remains, most often consist of haematite and goethite. In Hungary, they are frequent in the upper parts of the deposits at Nyirád, Szőc and Halimba. The dark violet concretions formed along decayed roots in Eplény bauxite contain in addition to haematite also some *lepidocrocite* (Bárdossy 1968a), only 2 to 3% in the outermost and innermost zones of the concretions, but up to 16% in between (fig. 96).

5.2.5.14. FISSURE FILLINGS

These are also mono- or at most di-mineralic. Veins confined to within individual textural elements consist of gibbsite or diaspore, more seldom of calcite, boehmite or siderite. They had formed partly during the reworking of bauxite, and partly during its diagenesis.

Fissure fillings traversing the matrix also are mostly epigene. In structurally strongly deformed deposits, fissure fillings of calcite are especially frequent, e.g. in the deposits of Central Greece. In the Devonian deposits of the Northern Urals, USSR, fissure fillings are mostly diasporic, less often calcitic, kaolinitic, pyritic, marcasitic, and goethitic. The most frequent di-mineralic filling there consists of diaspore and kaolinite, with the larger, automorphic diaspore crystals cemented together, as it were, by minute crystals of kaolinite. Fissure fillings combining calcite and kaolinite, calcite and diaspore, kaolinite and pyrite, or pyrite and marcasite are also found on occasion. In the Ivdel deposits of this region, the diaspore veins contain automorphic crystals of titanite in places (Beneslavsky 1958).

In the Hungarian deposits, the most frequent fissure fillings are those consisting of calcite. Fillings of kaolinite, crandallite, haematite and haematite–goethite are rarer. There are veins of gypsum wherever a bauxite containing pyrite and marcasite underwent a secondary oxidation. Especially frequent in the deposits at Szőc, and in the uppermost parts of the Gánt deposits, they range in thickness from one tenth of a millimetre to one centimetre.

Nia (1968) found fissure fillings 0·1 to 1 m wide, consisting of coarsely crystalline diaspore, in the Turonian bauxite horizon of Central Greece. These traverse not only the diasporic bauxite but pass also into the adjacent boehmitic bauxite. Their epigene origin is proved by the fact that they intersect the diagenetic textural elements.

Fissure fillings are coarser-grained than their surroundings, regardless of what minerals they are formed from. The lattice structures of their minerals are better-ordered and purer than those of the minerals of the embedding matrix.

To review our findings concerning the textural elements, we may draw the following conclusions of general scope:

(i) The clastic bauxite particles, pebbles and boulders are richer in haematite and goethite than the embedding matrix. The amounts of other minerals that they contain are entirely random as compared with their surroundings.

(ii) The greatest changes in mineral associations are between the matrix and the concentric particles: they are most often due to diagenetic processes.

(iii) The textural elements produced by epi- and supergene processes—nests, concretions, fissure fillings—are most often mono-mineralic, more seldom di- or trimineralic. Some of the minerals making them up are of very low concentration everywhere else in the bauxite.

(iv) Owing to their greater compactness and lower permeability, the clastic textural elements were less affected by secondary processes than the matrix in which they were embedded. On their surfaces, on the other hand, secondary minerals had often segregated to form films and coatings.

5.2.6. MINERALOGICAL CHANGES CONNECTED WITH COLOUR

5.2.6.1. MINERALOGY OF MULTICOLOURED BAUXITE TYPES

The diverse types of multicoloured bauxites were described in the chapter on petrography. The colour patterns of some of these were seen to be entirely independent of texture.

The mineral composition of the near-vertical light veins found in red bauxite is presented in the example of a sample from Nyirád–Izamajor, Hungary (table 34).

Table 34. Mineral composition of the vertical white vein and surrounding bauxite from the Nyirád–Izamajor deposit group, Hungary (weight percentages).

Bauxite types	Boehmite	Haematite	Goethite	Maghemite	Total iron minerals	Kaolinite	Alunite	Anatase	Rutile
Dark rust-red bauxite next to vein	50·6	40·3	3·8	0·8	44·9	3·2	–	1·0	0·3
Pinkish white vein	93·2	0·2	0·4	0·2	0·8	3·3	0·5	1·8	0·4
Concentration coefficient	1·8	0·005	0·1	0·2	0·02	1·0	∞	1·8	1·3

Most of the iron minerals are seen to have been leached out of the vein: haematite was leached out most thoroughly, goethite less so, and maghemite least of all. Boehmite, anatase and rutile underwent some relative enrichment by about equal percentages. Kaolinite content is unchanged. In view of the strong depletion of iron minerals, this must mean that part of the original kaolinite content was also removed. Alunite as a newcomer is obviously a segregation out of the sulphatic solutions of pyrite decomposition. The surrounding brick-red bauxite contains 22 to 26% haematite on average. The haematite content of 40·3% along the vein must thus be secondary in part; it was segregated in the process of leaching, in a thickness of 1 to 2 cm on either side of the vein, impregnating the brick-red bauxite.

Table 35. Mineral composition of the vertical pale yellow vein and the surrounding bauxite from the Szőc–Határvölgy deposit group, Hungary (weight percentages).

Bauxite types	Gibbsite	Boehmite	Total Al minerals	Haematite	Kaolinite	Anatase+rutile
Rust-red bauxite next to vein	55·2	10·0	65·2	26·2	2·7	2·4
Pale yellow vein	70·0	20·0	90·0	2·2	4·2	3·1
Concentration coefficient	1·3	2·0	1·4	0·1	1·5	1·3

The writer established the composition of similar veins found in the Határvölgy mine, Szőc, Hungary, to be as shown in table 35.

The basic pattern of the change is the same as above: iron minerals were dissolved, entailing a relative enrichment of the other minerals. The bauxite here is boehmito–

Table 36. Mineral composition of red bauxite with yellow spots from the Zlabor Hran deposit, Slovenia (weight percentages).

Bauxite types	Diaspore	Boehmite	Gibbsite	Total Al minerals	Haematite	Maghemite	Goethite	Total iron minerals	Kaolinite	Dioctahedral chlorite	Anatase	Rutile
Rust-red bauxite	5	55	2	62	17	2	5	24	7	3	2	0·5
Ochroid spots	6	62	–	68	2	4	12	18	6	3	2·5	1
Concentration coefficient	1·2	1·1	0	1·1	0·1	2·0	2·4	0·8	0·9	1·0	1·3	2·0

Table 37. Mineral composition of multicoloured bauxite from the József III ore body, Iszkaszentgyörgy, Hungary (weight percentages).

Bauxite types	Boehmite	Gibbsite	Total Al minerals	Haematite	Goethite	Total iron minerals	Kaolinite	Anatase	Rutile	Crandallite	Apatite
Brownish-red spots	32·4	23·8	56·2	22·6	6·1	28·7	8·0	2·2	0·5	1·9	–
Ochroid spots	37·9	26·0	63·9	7·6	9·2	16·8	11·4	2·1	0·6	2·7	–
Pink spots	42·0	23·4	65·4	2·0	4·6	6·6	18·7	2·1	0·6	3·7	0·7
Pale violet spots	47·7	24·2	71·9	0·3	1·2	1·5	18·5	2·0	0·6	3·2	–

gibbsitic. Dissolution enriched boehmite more than gibbsite, displacing the gibbsite/boehmite ratio in the vein in favour of boehmite. The writer has made similar findings in his examination of the yellowish veins found in red bauxite at Eplény, Hungary. The bauxite there is haematito–goethitic. The original haematite-to-goethite ratio was 2·1 in the red bauxite: it is decreased to 0·4 in the yellowish veins. In step with the dissolution of iron, the composition changed in favour of goethite. A similar situation has been found to prevail in the Nyirád–Izamajor deposit.

The mineralogical changes in red bauxites with yellow spots are the same by and large as those described above for the veins. On occassion, however, the dissolution of iron minerals is subordinate, as for example in a rust-red bauxite with abundant vivid yellow spots in the top part of a small bauxite deposit near Zlabor Hran in Savinja Valley, Slovenia, whose composition is presented in table 36.

The aggregate iron mineral percentage hardly decreased, but the association underwent a profound change; haematite was reduced by a factor of ten, maghemite increased by a factor of two, goethite by a factor of 2·5. This may be another case of epigene dissolution, but with the iron precipitated almost on the spot, in the form of goethite and maghemite. The aluminium-mineral association was also affected to some extent; gibbsite disappeared, and the contents of boehmite and diaspore increased somewhat.

The spots of the Gánt bauxite shown in colour figure 19 represented a local rearrangement of iron minerals. It is the dark brown parts of goethitic composition that are richest in iron. The ochroid parts are also goethitic, but lower in iron. The pale yellow parts are lowest in iron: they are haematito–goethitic. Finally, the violet spots are pure haematitic and comparatively low in iron.

The mineral composition of a sample of multicoloured bauxite from József III ore body, Iszkaszentgyörgy, Hungary is given in table 37. It is presumably the brownish red spots that represent the original bauxite, the lighter spots being the products of a subsequent leaching of iron. In step with the depletion of iron minerals, the bauxite, haematitic in the original state, became pronouncedly goethitic. This can be quantified in terms of the haematite/goethite ratio, which in the four samples, from the top downwards, has been found to be 3·7, 0·8, 0·4 and 0·25, respectively. At the same time, the aluminium-mineral content increased and the bauxite became more boehmitic, with the boehmitic/gibbsite ratio evolving from 1·36 to 1·46, 1·80 and 1·97, in the above succession. Deferrification affected neither the total percentage nor the relative abundances of the titanium minerals. Kaolinite on the other hand was enriched so much as to imply a local neoformation. The remarkably high crandallite content of the sample further increased in the course of deferrification.

Dark spots on a light background are usually due to textural elements such as bauxite pebbles, roundgrains or pisoids. The mineralogical changes connected with them were described in the foregoing section.

In summary, light veins and spots independent of the textural elements were formed largely by deferrification and the partial reprecipitation of iron. These processes tended to shift the haematite/goethite ratio in favour of goethite. Leaching usually resulted in a relative increase of aggregate aluminium-mineral percentage, accompanied, in those bauxites which contain both minerals, by a shift in favour of boehmite of the

boehmite/gibbsite ratio. The titanium minerals and kaolinite were relatively enriched by deferrification. In several cases, a subsequent addition (neoformation) of kaolinite seems probable.

5.2.6.2. DISCOLOURATION OF REDUCED BAUXITES IN OUTCROPS

The grey pyritic–marcasitic bauxites of Hungary were oxidised whenever they rose into the proximity of the ground surface (even under an overburden 10 to 50 m thick). They cannot therefore be observed in their original state in any surface exposures. The harder, more compact and more resistant Triassic and Paleozoic reduced bauxites, on the other hand, are often also preserved in the outcrops, in which case they wear a russet zone of supergene oxidation on their exposed surfaces. In Slovenia for example, near Podlipa village, a very hard, dark greenish grey bauxite was observed by the writer in outcrops of the Triassic bauxite horizon. It is covered by a vivid red zone of oxidation 1 to 3 mm thick. The interface with the original reduced bauxite is very sharp but of an irregular, zigzagging course. Between the greenish-grey and the red bauxite there is a zone, a tenth of a millimetre wide, of pale yellow bauxite. The mineral composition of the original bauxite and of the red zone is given in table 38.

Table 38. Surface oxidation of grey bauxite in an outcrop of the Podlipa deposit, Slovenia.

Bauxite types	Boehmite	Gibbsite	Haematite	Goethite	Pyrite	Ferro chamosite	Kaolinite	Anatase	Rutile
Dark greenish-grey bauxite	34	–	–	2	2	30	29	1·5	0·5
Vivid red zone	34	3	17	5	–	–	38	1·5	0·5
Concentration coefficient	1·0	∞	∞	2·5	0	0	1·3	1·0	1·0

Surface oxidation decomposed all the pyrite and chamosite. Their iron content thus freed was precipitated as haematite and goethite, and their aluminium and silica content as kaolinite. The percentage of boehmite and of the titanium minerals was unchanged, which implies that the gibbsite detected in the red zone must have formed out of the alumina freed from chamosite.

A similar case of surface oxidation has been observed by the writer in outcrops of another Triassic bauxite, the Vrace deposit of Lika County, Yugoslavia.

Some of the Permian bauxites in Vietnam were eroded and redeposited in the Neogene. The form of their deposit was described in § 3.6 in connection with the Salento-type deposits. Most of the reworked deposits are unburied. The bauxite blocks up to one cubic metre in size are oxidised on their surfaces. The oxidised zone is of varying thickness: a few millimetres in some deposits, more than 10 cm in some others. On cut faces of such bauxite pebbles and boulders, the sharp irregularly zigzagging limit between the original greenish-grey, dark grey or black bauxite and the vivid red or violet oxidised bauxite is clearly visible. The hardness of the bauxite is almost unreduced in the oxidised zone, but its porosity is higher. The writer has investigated the composition of several bauxite pebbles and boulders, and found them to differ somewhat between them depending on their original composition. In a predominantly diasporic sample, the mineralogical changes presented in table 39 could be detected.

Surface oxidation decomposed all the chamosite, leptochlorite and organic matter. Their iron content was changed partly into haematite, partly into goethite, and the liberated silica and aluminium segregated to form kaolinite and metahalloysite. Oxidation depleted boehmite somewhat more than diaspore. Part of the matter dissolved out of these two was precipitated as gibbsite. The aggregate abundance of the titanium minerals remained approximately constant.

On the surface of a dark grey sample consisting of boehmitic bauxite, a russet and orange zone of oxidation 0·5 to 1·5 m thick has been observed. Inward, it runs up with a very sharp limit against a faded-green zone, which in turn passes with no sharp limit into the dark grey original bauxite. The mineral compositions of the three zones are given in table 40.

The mineralogical change starts in the faded green zone, where chamosite, leptochlorite, ilmeno-rutile and siderite are substantially depleted, and organic matter has disappeared altogether. There is some goethite and kaolinite, formed out of the decomposed minerals. This, however, was only a first step in superficial oxidation, and the truly profound changes are those reflected by the russet zone, where all the reduced minerals have disappeared completely, and goethite and kaolinite are further enriched. Gibbsite and haematite appear as newly formed minerals. In contrast to the foregoing sample, the abundance of the aluminium minerals is not decreased but even increased slightly. Even the titanium minerals are somewhat enriched. Sharova and Glazunova (1966) made findings conformable to these on Vietnamese bauxite samples of unspecified origin.

In summary,

(i) The superficial oxidation of reduced bauxites decomposes all the reduced minerals (chamosite, leptochlorite, pyrite, siderite, ilmenorutile and organic matter).

(ii) The iron thus freed precipitates in the form of goethite and haematite. The formation of goethite will begin before that of haematite.

(iii) Liberated silica and aluminium go into the formation of kaolinite, occasionally of kaolinite and metahalloysite.

(iv) Depending on the surface conditions, the abundances of the aluminium minerals are either slightly increased, or unchanged, or reduced. Gibbsite appears as a neoformation in the zone of oxidation.

(v) The aggregate and relative abundances of the titanium minerals hardly change.

5.3. THE MINERALS OF KARST BAUXITES

Let us review the mineral groups in decreasing order of their abundance in bauxites. For the sake of a better overview, the oxide and hydroxide minerals are arranged below according to their principal elements.

In § 5.2, we surveyed in detail the distribution of and the interrelations among the rock-forming Al, Fe, Si and Ti minerals of karst bauxite. In the following section, information concerning these minerals will be confined to such items not already presented in the foregoing section.

Table 39. Mineral composition of an oxidation zone on a greenish-grey bauxite pebble (weight percentages).

(Ba–Xa deposit group, Lang-Son region, Vietnam)

Bauxite types	Dia-spore	Boehm-ite	Gibbs-ite	Haematite	Goethite	Chamosite	Lepto-chlorite	Kaolin-ite	Meta-halloysite	Quartz	Anatase	Rutile	Siderite	Organic matter
Greenish-grey bauxite	39·7	11·7	–	–	12·42	24·1	1·5	6·0	–	0·1	2·6	0·3	0·5	0·2
Brownish-red zone of oxidation	29·9	2·6	5·0	8·6	25·6	–	–	20·0	3·0	0·1	3·0	0·2	–	–
Concentration coefficient	0·8	0·2	∞	∞	2·1	0	0	3·3	∞	1.0	1.2	0.7	0	0

Table 40. Mineral composition of an oxidation zone on a dark grey bauxite (weight percentages).

(Ba–Xa deposit group, Lang-son region, Vietnam)

Serial number	Bauxite types	Diaspore	Boehmite	Gibbsite	Haematite	Goethite	Chamo-site	Lepto-chlorite	Kaolin-ite	Quartz	Anatase	Rutile	Ilmeno-rutile	Siderite	Organic matter
1.	Dark grey bauxite	1·4	53·5	–	–	2·0	30·0	2·3	6·0	0·4	1·9	0·3	0·6	0·2	0·3
2.	Faded green bauxite	1·7	51·6	–	–	11·3	21·7	0·6	8·8	0·4	2·3	0·2	0·2	0·1	–
3.	Russet zone of oxidation	3·4	56·0	1·2	6·2	15·5	–	–	12·8	0·2	2·2	0·4	–	–	–
Concentration coefficients 1 to 2		1·2	0·9	–	–	5·6	0·7	0·3	1·5	1·0	1·2	0·7	0·3	0·5	0
Concentration coefficients 1 to 3		2·4	1·1	∞	∞	7·8	0	0	2·2	0·5	1·1	1·3	0	0	0

5.3.1. OXIDES AND HYDROXIDES

5.3.1.1. ALUMINIUM OXIDES AND HYDROXIDES

Gibbsite, γ-$Al(OH)_3$. The mineral is also called *hydrargyllite*, especially in German. The term gibbsite is better justified on account of its priority (cf Palache *et al: Dana's System of Mineralogy* vol. 1, 1944)

This is the third most abundant aluminium mineral of karst bauxites after boehmite and diaspore. It can be demonstrated in all the textural elements. Soon worn in the course of transport, it is fairly rare as an independent clastic mineral. It forms no isomorphous mixed crystals with any other element in bauxite. It may have a wide range of origins, syn- and diagenetic, epigene or supergene.

Bayerite, α-$Al(OH)_3$. Up to the 1950s it was known only as a synthetic phase. Gedeon (1956) deduced from Habicht-device results that part of the aluminium trihydrate content of Hungarian bauxite is in the form of bayerite; his hypothesis has been refuted by large numbers of X-ray analyses performed since.

Tertian (1966) found at the boundary of a white bauxite and its cover in the Vonji Do bauxite deposit of Crna Gora (Montenegro, Yugoslavia) a white nest of 4 to 5 cm size which he proved to contain a little bayerite in addition to gibbsite and nordstrandite. He presumed the bayerite to have crystallised out of an amorphous gel-like substance.

At the bottom of a bauxite deposit near Marmara, Greece, a small quantity of bayerite accompanying some nimesite was identified in a greenish bauxitic clay by Maksimović (1973).

Khorosheva (1968) identified bayerite in the Tikhvin-type bauxite of the Srednyaya Pridneprov'ya in the Ukraine, USSR. Sokolova (1975) found a very small amount of bayerite accompanying gibbsite and nordstrandite in the deposits, likewise of Tikhvin type, of Sukholebyazhinskoe and Nizhnepodsopochnoe along the Yenisei River.

In recent times, bayerite has been identified in a number of non-bauxitic deposits, e.g. by Náray-Szabó and Péter (1967), in the finest fraction of Hungarian Quaternary clays and by Gross and Heller (1963) in a Cenomanian–Eocene sedimentary succession in Israel. It should therefore be looked for also in karst bauxites, especially in the youngest ones.

Nordstrandite, $Al(OH)_3$. Even as a synthetic phase, the mineral has been known since 1958 only. As a natural phase, it was first identified in different terra rossas (Hathaway and Schlanger 1962; Wall *et al* 1962; Marić 1968).

Tertian (1966) demonstrated nordstrandite in the bauxite deposit of Vonji Do, Crna Gora, in the white nests mentioned in the section on bayerite; nordstrandite was more abundant there than bayerite.

According to Davis and Hill (1974), there are several occurrences of nordstrandite accompanying gibbsite on the northern side of Jamaica. A largish number of chemical and X-ray analyses established the following relationship: nordstrandite is absent from, or very scarce in, bauxites containing less than 1% SiO_2. It is enriched in bauxites containing 1 to 5% silica, so much so that it surpasses gibbsite in some samples. Finally, in the samples of 5 to 12% silica content, it is ubiquitous, and its percentages equal by and large those of gibbsite.

The authors cited ascribe this distribution to the fact that the pH, as determined *in situ*, is between 6·0 and 7·5 in the low-silica bauxites and slightly higher than 7·5 in the high-silica ones. They consider the formation of gibbsite to have been promoted by a weakly acidic and neutral pH and the formation of nordstrandite by a weakly alkaline one. In some of their diffractograms, Davis and Hill observed some accessory 100-intensity peaks between the reflections of gibbsite at 4·85 Å and of nordstrandite at 4·79 Å. They drew the conclusion that there may exist in nature some polytypes of aluminium trihydrate other than gibbsite, bayerite and nordstrandite.

Just like gibbsite, nordstrandite is finely disperse, not attached to any textural elements. It is presumably formed simultaneously with gibbsite as a product of syngenetic and early diagenetic processes.

In two bauxite samples from Niué Island, East Tonga Group, Pacific Ocean, the writer has found by X-ray investigations some nordstrandite besides gibbsite. One sample contained 28% nordstrandite and 5% gibbsite; the other contained 6% nordstrandite and 35% gibbsite. Both samples contained also significant quantities of amorphous ferric and aluminium hydroxides.

Kulikova *et al* (1974) identified nordstrandite in the form of pore and fissure fillings in deposits of bauxitic iron ore at Sokolovsk–Sabraysk, USSR. The acicular nordstrandite crystals were in the size ranged in, from one hundredth of a millimetre to 3 mm. Clearly secondary in origin, the mineral was found to be most abundant near the bauxite's contact with the underlying limestone, at a distance of 10 m above it as a maximum, implying control by a pH higher than 7.

Nordstrandite accompanied by gibbsite and bayerite was demonstrated by Sokolova and Kozlov (1976) to be also from the karst bauxite deposit called Listvennoe along the Yenisei River, USSR. It is cryptocrystalline, and so its primary or secondary origin could not be decided. If it were primary, it would be the oldest known occurrence of nordstrandite, in view of the Senonian–Palaeocene age of the Listvennoe deposit.

Sokolova (1975) found some nordstrandite accompanying gibbsite and a very little bayerite in reworked Tikhvin-type bauxite deposits at Sukholebyazhinskoe and Nizhnepodsopochnoe along the Yenisei River, Siberia, USSR.

No nordstrandite has been found so far in any typical lateritic bauxite.

The writer deems the presence of nordstrandite to be likely also in all other Miocene and Quaternary karst bauxites, an assumption that could be verified by systematic X-ray analyses. Its probability of occurrence is likely to decrease towards the older deposits, as a longer time and overburden pressure, acting either jointly or severally, should be sufficient to recrystallise it into gibbsite.

Tučanite, $2Al_2(OH)_6 \cdot H_2O$. Karšulin (1960, 1964) found a nest of about 50 cm size of a chalk-like, white, porous substance, in a white bauxite in a deposit near Carev Most, Crna Gora (Montenegro), Yugoslavia. Detailed X-ray, chemical and thermal analyses of the substance permitted him to identify in the substance a new mineral which he named after Fran Tučan. Tučanite has not been identified in any other bauxite deposit so far, but it has been prepared in the laboratory by seeding an amorphous aluminium hydroxide gel with tučanite grains. Strunz (1970) deems tučanite not to be an independent mineral species but probably identical with scarbroite. The present writer

disagrees with Strunz on the ground that the lattice constants of the two minerals differ, and so do their unit cell dimensions; he accordingly holds tučanite to be an independent although very rare mineral.

Karšulin did not study the natural conditions of formation of the new mineral: he only stated that at 130 °C it loses water to change first into aluminium trihydrate and then into boehmite. According to his electron micrographs, natural tučanite forms rhombohedral grains of 1 to 2 μm size. The fact that it is found in a nest of white bauxite has prompted the present writer to infer its secondary origin.

Scarbroite, $12Al(OH)_3 \cdot Al_2(CO_3)_3$. This mineral of great rarity is only known so far from the bauxite deposit of Pilisvörösvár, Hungary, where it was proved by Nemecz and Varju (1967) to form hard, massive segregations of a few centimetres size at the contact between a bauxitic clay and its dolostone bedrock. This form of occurrence recalls that of the gibbsite segregations found in many deposits between the bauxite and its bedrock. Their identity of position is further enhanced by the fact that segregations of scarbroite tend to pass laterally into segregations of gibbsite.

In the writer's opinion, the mineral is a product of epigene processes similar to those giving rise to the gibbsite segregations, with the difference that under the simultaneous influence of hydrocarbonate solutions some aluminium carbonate also got introduced into the lattice. The fact that the mineral has been detected in this single deposit so far—except a laterite-bauxite deposit at Gippsland, Australia (Grubb 1973)—implies that the physico-chemical conditions of its origin must be highly special.

Boehmite, γ-AlOOH. As distinct from gibbsite and diaspore, it is found almost exclusively in crystals of less than one micrometre in size. According to Gulyanitsky (1969), there is some crystalline boehmite, of 20 μm size at most, accompanying a predominantly cryptocrystalline boehmite phase in the Carboniferous deposits of the Southern Timan Mountains, USSR. These crystals of authigenic origin are of cask or rhombohedron shape.

The mineral is largely syngenetic, less often diagenetic and epigene. It may also be due to supergene processes on some rather rare occasions.

There is in Nature a type of boehmite whose lattice contains some ferric iron in isomorphous substitution, as first demonstrated by Caillère and Pobeguin (1961) in bauxite samples from the Bédarieux deposit group, France. In their monograph on the mineralogy of French bauxites (1965), these authors posited an iron-containing boehmite, which they called 'ferriboehmite', to be present in practically every deposit group, in places with an AlOOH/FeOOH ratio as low as 1:1. Their conjecture of so substantial a substitution was not confirmed by any shift of lines in their Debye–Scherrer X-ray patterns and but a very slight one in their infra-red absorption spectra. All this made many a specialist doubt the correctness of their findings. De Grammont in his laboratory experiments (oral communication, 1971) found that synthetic ferriboehmite can accommodate 10% FoOOH at most. Substitution is accompanied by an appropriate shift in the X-ray reflections. In the present writer's opinion, this is the greatest degree of substitution admissible in bauxite. In none of his X-ray diagrams of boehmitic bauxite could he observe any shifts of reflections indicating a higher degree of substitution.

All this agrees rather well with the experimental findings of Jónás *et al* (1973) who found that the crystallinity of synthetic boehmite is higher if it is prepared in the presence of ferric iron ions, but that its lattice parameters remain unchanged. Iron thus catalyses the crystallization of boehmite but does not enter its lattice as an isomorphous substituent. The authors cited found the crystallinity of boehmite, which can be determined best by measuring the specific extinction (ε^*) of the mineral in the infra-red spectrophotometer, to vary widely in bauxite.

Yershova *et al* (1975), studying pure boehmite fractions separated from bauxite, likewise found that lattice ordering in natural boehmites is highly varied. It is best in comparatively coarse-grained boehmite of secondary origin, found in pores and fissure fillings. Such boehmite contains no isomorphous substituent. Finely disperse boehmite, on the other hand, is more disordered: those authors proved it to contain 0·2 to 2·5 wt % Fe_2O_3 in isomorphous substitution by means of a Mössbauer device.

Diaspore, α-AlOOH. The second most abundant aluminium mineral of karst bauxites after boehmite, its particle size varies from a few tenths of a micrometre to several hundred micrometres. Present in practically all the textural elements, especially in the concentric ones, it may be syn- and diagenetic or epigene.

Gout (1973) identified two varieties of diaspore in the Unjat bauxite deposit, French Pyrenees. One is of fair, the other of poor crystallinity. The latter decomposes at a lower temperature, and its X-ray reflections are flatter. It is undecided whether the poorly crystalline variety is a secondary product derived by structural (tectonic) deformation from the well-crystallised one, or whether it is an original syngenetic phase.

Bauxites contain a type of diaspore which contains more or less ferric iron as an isomorphous substituent. This variety is termed *ferridiaspore*. Goethite, whose lattice structure is the same as that of diaspore, is in its turn capable of the isomorphous accommodation of aluminium. This, then, is an isomorphous series which extends to 30 to 40 mol % substitution on the goethite side and to 10 to 20% on the diaspore side. The degree of substitution can be accurately determined from shifts in the X-ray reflections and in the infra-red absorption spectra.

Caillère and Pobeguin (1962a) were the first to demonstrate ferridiaspore in karst bauxites. They first identified it in several samples in a deposit near Péreille, Ariège, France, and later on in several other French deposit groups (1965). Their investigations revealed most of the diaspore in the matrix to be pure and the one in the pisoids to be of the ferridiaspore variety. Gout (1971) found ferridiaspore containing 3 to 4 mol % FeOOH in the Péreille deposit of the French Pyrenees and in the Froussia and Dekadendra deposits of Greece. X-ray analyses by the present writer have led him to the conclusion that ferridiaspore is present in most deposit groups containing diasporic bauxite.

Alumogel, AlOOH + aq. Essentially an amorphous aluminium hydroxide gel, it used to figure in some earlier publications by the names *sporogelite* and *cliachite*. It used to be a generally accepted view that all bauxites contain significant quantities of amorphous aluminium hydroxide gel, an assumption not borne out by systematic X-ray diffractometry.

As far as we know it today, there are significant percentages of alumogel only in the Pliocene, Quaternary and recent bauxites. According to Ahmad and Jones (1969), the

bauxites of the Cayman Islands in the Caribbean contain 5% amorphous alumogel on average; some bauxites of the Bahamas contain 5 to 11%. Tercinier (1971) found 5 to 10% in bauxite from Lifou Island, Loyalty Group.

The present writer has demonstrated by X-ray methods 10 to 15% alumogel in Lifou Island bauxite, 20 to 25% in Maré Island bauxite (both of the Loyalty group) and 20 to 25% in Niué Island bauxite (East Tonga Group). The overwhelming majority of older bauxites contains no alumogel at all.

Alumogel is greatly enriched in certain secondary nests and concretions. In most instances, it is semi-crystalline, 'quasi-amorphous' rather than entirely X-ray amorphous. It most often accompanies gibbsite and presumably passes itself into that mineral by a slow rearrangement (ageing). In the Szőc–Határvölgy mine, Hungary, e.g. reoxidised bauxite in the brownish yellow upper part of the deposit contained white, massive, mammillary nests of a few centimetres size which in addition to 40% gibbsite were found to contain 60% semi-amorphous alumogel. White, friable nests collected by Vadász from Margot mine at Nyirád, Hungary, have been shown by the present writer to contain some 10% alumogel in addition to gibbsite and alunite. There is some 20% alumogel, accompanied by gibbsite and aluminite, in a white layer a few centimetres thick at the boundary of bauxite and bedrock in the Szőlőhegy open pit, Sümeg, Hungary. Finally, a gibbsitic layer at the bedrock boundary in the bauxitic clay deposit of Cserszegtomaj contains in places 10 to 50% semi-amorphous alumogel (Bárdossy 1961b).

All these nests and concentrations are partly of epigene and partly of supergene origin. Alumogel, then, is much younger than the bauxite in which it is found. Its formation is frequently connected with the secondary oxidation of pyrite.

Corundum, α-Al_2O_3. Karst bauxites contain three types of corundum. The most frequent one is clastic, found in grains of 0·05 to 0·5 mm in size. It nowhere exceeds 0·1% of the bauxite, and most often amounts to a hundredth or a thousandth of a percent only. This type of corundum is present in all the Hungarian deposit groups (Kiss 1952, 1955; Kiss and Vörös 1965; Vendel *et al* 1971; Vörös 1958, 1971). Presumably the erosion product of a metamorphic basement facies, it has a grain size in the 0·1 to 1 μm range.

The second form of occurrence is in so-called *microcorundum*. The term denotes micro- and cryptocrystalline particles of corundum formed *in situ* under Earth-surface conditions. Its rather poor crystallinity has been revealed by X-ray analysis. Its grain size is generally 0·1–1·0 μm.

Microcorundum was first demonstrated by Terentyeva (1949) from certain Kazakhstan bauxites; it was later identified in the bauxites along the Yenisei River by Beneslavsky (1963) who stated it to be enriched in the interiors of dark brown and brownish black pisoids in the hard types of bauxite above all. According to Medvedev (1971), the pisoids in the Porozhninsk deposit of the deposit group along the Angara River contain 23 to 51% microcorundum, whereas the matrix contains none. The pisoids are found near the deposit tops above all. According to Lavrenchuk *et al* (1974), some hard bauxite pebbles of 1 to 4 cm size contain 20 to 25% microcorundum on average in the Arkalyk deposits.

Microcorundum has so far been identified in eight deposit groups, all of Kazakhstan type. In each of these, it is accompanied by the same mineral association: gibbsite, magnetite, maghemite, haematite, goethite, kaolinite, anatase and rutile. These deformation of significant strength, and the overburden covering them is less than 100 m as a rule. The corundum must therefore have formed of necessity under normal m as a rule. The corundum must therefore have formed of necessity under normal Earth-surface temperature and pressure, presumably as a product of diagenetic processes.

The third type of corundum is found in deposits which have undergone contact and regional metamorphism. It forms well-crystallised grains of 0·1 to 2 mm in size. There is some corundum produced by contact metamorphism, e.g. in the bauxite deposits near Remeţi in the Bihar Mountains, Rumania (Szádeczky 1905). In the USSR, corundum is contact metamorphic in both the Obukhovsk deposit group and in Uzbekistan, in the South Nuratinsk Mountains deposits (Beneslavsky 1963).

Emery is a fully metamorphic bauxite with corundum as its principal rock-forming mineral. Its classical site is Naxos Island in the Aegean, whose mineralogy has been investigated in detail (Lapparent 1946; Papastamatiou 1951). Corundum is found in rock-forming abundance also in the emery deposits of the Western Taurus and the Menderes Massif (de Lapparent 1946; Önay 1949). De Lapparent (1937) described by the name *taosite* an iron- and titanium-bearing corundum mineral from a less strongly metamorphic bauxite found on Samos Island. Strunz (1970) does not admit it as an independent mineral; he identifies it with *högbomite*. The emery deposit at Irtyashskoe in the Urals, USSR, is another instance of metamorphic bauxite with corundum (Buhinsky 1971, 1975).

δ-, γ- and Θ-Al_2O_3. Until quite recently, these have been regarded as exclusively synthetic. Caillère and Pobeguin (1970) demonstrated some 3% theta-Al_2O_3 by natrium hydroxide leaching and X-ray analysis in bauxitic claystones from the Ollières and Comebecave deposits, Southern France. These grains of cryptocrystalline size are attributed by the present writer to authigenic processes.

In a sample from the contact of bauxite with a basalt dyke in the Bédarieux deposit group, France, Bárdossy *et al* (1970) succeeded in identifying each of these three varieties of Al_2O_3. Their aggregate quantity in a contact zone of 5 to 20 cm width was 10 to 30% as a rule, 37% as a maximum. Very poorly crystalline, giving flat, diffuse reflections in the diffractograms, they could not be individually measured quantitatively, only their relative abundances could be established as decreasing from delta to gamma to theta. Accompanied by 1 to 3% corundum, they are associated with the minerals halloysite, hercynite, maghemite, haematite, anatase and rutile, all formed out of a boehmito-haematito–kaolinitic bauxite under contact metamorphism due to the basalt dyke.

Spinel, Al_2MgO_4. It was identified by Vörös and Gecse (1976) in a bauxite of the Iszkaszentgyörgy–Bitó deposit, Hungary, in the form of scarce clastic particles.

5.3.1.2. IRON OXIDES AND HYDROXIDES

Haematite, α-Fe_2O_3. Most frequent among the iron minerals of karst bauxites, it has been identified in almost every one of their textural elements. Its bulk is syn- and diagenetic: epi- and supergene haematite is less abundant.

Bauxite contains, in places, a type of haematite which contains some aluminium as an isomorphous substituent. This variety of the mineral, first demonstrated by Beneslavsky (1963), is called *alumohaematite* following his proposal. Solymár (1970) succeeded in producing alumohaematite by the heating of synthetic alumogoethite in the laboratory. The X-ray reflections of alumohaematite are shifted to an extent depending on the degree of aluminium substitution. Solymár deems the greatest possible Al_2O_3 substitution to be 14 mol%. The heating of aluminiumless goethite produces a deep violet haematite, whereas the ignition of alumogoethite gives rise to a reddish alumohaematite. The colour of haematitic bauxite, which is red here and violet elsewhere, may thus be a function of the degree of aluminium substitution.

Janot and Gibert (1970) developed a high-accuracy method for assaying aluminium substitution by means of a Mössbauer device. Having detected degrees of substitution ranging from 5 to 12 mol % by this method, they state alumohaematite to be typically extremely fine-grained, in the 150 to 400 Å range. According to the present writer's X-ray investigations, the haematite found in the Hungarian bauxites contains 1 to 3 mol % Al_2O_3 as a rule.

The formation of isomorphous mixed crystals is also possible between haematite and ilmenite. The mixed crystal containing 0 to 33 mol % $FeTiO_3$ is called *titanohaematite* (Strunz 1970). Kiss (1955) identified a few grains of titanohaematite by ore microscopy in a bauxite from Gánt. Vörös (1971) found the plus-0·06 mm elutriated fractions of 51 samples from Gánt, Hungary to consist predominantly of titanohaematite. According to X-ray fluorescence analysis they contain 5 to 10% titanium. He held the bulk of the mineral to be authigenic, admitting a clastic origin for the largest particles only. The latter are slightly worn, whereas the former are unaltered and automorphic, although a few grains may be corroded and porous. In Jamaican bauxite, Hartman (1955) found a few percent of titanohaematite concentrated in the heavy minerals fraction.

The present writer follows Strunz's (1970) lead in not regarding as separate minerals *hydrohaematite* and *turgite*, described as such by some authors. These are in fact just finely disperse haematite types containing some adsorbed H_2O.

Maghemite, γ-Fe_2O_3. Identified in 38 deposit groups so far, its usual perecentage range is from 1 to 4%. It is most abundant in the Quaternary bauxites of the Pacific islands. The writer's X-ray investigations have demonstrated 3 to 5% in Lifou bauxite (Loyalty Group) and 3 to 4% in Niué Island bauxite (East Tonga Group).

Particle size is very fine, not exceeding a few tenths of a micrometre. The mineral tends to form a finely disperse phase in the bauxite matrix, and may be substantially enriched in some pisoids. According to Beneslavsky (1963), in the Devonian deposits of the Northern Urals, USSR, it is usually found together with magnetite as an oxidation product of that mineral. It accompanies magnetite also in the corundo–gibbsitic pisoids and macropisoids of the Kazakhstan-type deposits.

In an average sample representing a larger number of bauxite pebbles from Arkalyk, received from Lavrenchuk, the present writer has observed 5% maghemite, accompanied by 14% haematite and 1% goethite. In pebbles of a hard bauxitic iron ore from the Sofievsk deposit of the Tselinograd group, the writer has found 10% maghemite associated with 20% haematite and 6% goethite.

In the Hungarian deposits, there was maghemite in one-fourth at most, in abundances of 1 to 4%, of the several hundred samples studied by the writer. It seems to be confined more or less to grey pyritic bauxite and to the types of secondary bauxite produced by its oxidation.

Maghemite is also enriched in the contact metamorphic zones of bauxites. The Bédarieux deposits e.g. contains 1 to 3% maghemite as a rule, whereas in the contact zone of a basalt dyke traversing the bauxite, Bárdossy *et al* (1970) found it to be enriched to 4 to 8%. The writer has assayed 6 to 12% maghemite accompanied by minor amounts of magnetite and haematite in samples of a contact metamorphic bauxite collected at Jádremete by the late Rozlozsnik.

Beneslavsky (1963) was the first to state that the maghemite lattice may also contain substitutions of aluminium: he gave the name *alumomaghemite* to the mixed crystal thus formed. By heating alumogoethite to 460 °C, Solymár (1970) suceeded in producing a synthetic alumomaghemite, which passed into alumohaematite on further heating. Part of the maghemite in bauxite presumably is alumomaghemite in fact. Its low abundance has so far precluded a quantitative determination.

Magnetite, $(Fe^{3+}Fe^{2+})Fe^{3+}O_4$. It has been identified in 27 deposit groups so far, but in 17 of these its abundance is in the tenth to hundredth of a percent range. Being resistant to wear and alteration, it is usually introduced into bauxite in the form of clastic particles. Such magnetite grains have been demonstrated from the deposits of Jamaica (Strahl 1971), Hispaniola (Goldich and Bergquist 1947, 1948) and the Southeastern US (Overstreet 1964). Clastic magnetite grains are enriched in places in the Boksonsk deposits of the USSR (Beneslavsky 1963). Bushinsky (1971) mentions abundant terrigenous magnetite grains from the Devonian deposits of the Northern Urals, USSR. In Hungary, a few worn magnetite grains were found by Gecse in the Óbarok bauxite and by Vörös and Gecse (1976) in the Nagyegyháza and Iszkaszentgyörgy–Bitó deposits. Some of the grains were turned into maghemite by secondary decomposition.

More seldom, magnetite may form authigenic, automorphic particles (octahedral and hexahedral), as demonstrated e.g. by Papiu *et al* (1970) in the deposits Pădurea Craiului, Rumania. Most of the magnetite there was transformed secondarily into haematite. It is sometimes only the pseudomorphs of haematite that can be observed today. In certain bauxites of Campania and the Abruzzi in Italy, Sinno and Franco (1965) found some octahedral and dodecahedral magnetite grains. In bauxite samples from Payas and Islahiye, Turkey, the present writer has demonstrated up to 3% magnetite by X-ray methods. These bauxites are very high in iron; their principal iron mineral, haematite, attains abundances of 40 to 60%.

In some deposits, magnetite concentrates in the pisoids and macropisoids. The pisoids of the small deposits near Pylos in Greece, e.g. contain according to Kiskyras (1958) 2 to 5% magnetite on average, more than the matrix. In the pisoids, magnetite is

enriched together with diaspore. In the bauxites at Chadobetsk, along the Yenisei River, at Arkalyk and on the Turgay Plateau, USSR, magnetite is found in corundo–gibbsitic pisoids and macropisoids, but cannot as a rule be demonstrated in the surrounding matrix. The iron minerals accompanying magnetite are haematite and maghemite. Most of the original magnetite has presumably changed into maghemite. According to Luković (1962) magnetite is enriched together with maghemite in the ooids and pisoids of the Triassic bauxite at Gornje Polje, Crna Gora (Montenegro), Yugoslavia, especially in the iron-rich nuclei of the latter. According to Beneslavsky (1963), magnetite is also found as a pore and fissure filling in some deposit groups of the USSR.

In the top part of some rust-red, yellow-veined bauxite in the Cseres mine at Halimba, Hungary, a martitized magnetite block of 20–30 cm diameter was found. About one metre above it, there is some grey pyritic bauxite. The block contains 13% magnetite and 6% goethite in addition to 80% haematite (Bárdossy 1961a). The concretion of magnetite may have formed during the slow oxidation of pyritic bauxite. From downward-filtrating solutions of ferrous and ferric sulphate, amorphous ferro–ferric oxide precipitated, and slowly crystallized into magnetite. Subsequently, most of it was martitised (converted into haematite).

Bauxite in the Eleusis–Mandra deposit group of Greece contains according to Kiskyras (1960b) 2 to 5% magnetite on average, up to 10% in places, accompanying haematite. Kiskyras posits the magnetite to be a secondary formation due to strong structural stresses. Kiskyras's hypothesis is born out by the present writer's finding 3 to 8% magnetite besides 10 to 14% haematite in the structurally most deformed deposit (the one near Mazin) of the Lika County bauxite deposit group, Yugoslavia. No magnetite has so far been identified in any other Lika County bauxite.

Bauxite metamorphism resulted in a gradual recrystallisation of haematite and goethite into magnetite. In the contact metamorphic bauxite near Remeţi in Rumania, e.g. magnetite is the only iron mineral (Szádeczky 1905). De Lapparent (1937) observed magnetite grains closely intergrown with haematite in weakly metamorphic bauxite on Samos Island, Greece. Its abundance was found to attain 11% in some of the samples. Weakly metamorphic bauxite in the Menderes Mountains Turkey, contains 1 to 2% magnetite on average; strongly metamorphic bauxite ('emery') contains 10 to 20% (Önay 1949). Some of the magnetite there forms independent grains; the rest forms minute octahedral inclusions in corundum. The classical emery deposit on Naxos Island, Greece, contains 20 to 25% magnetite on average (Papastamatiou 1951). A new generation of magnetite crystals was formed here during a phase of hydrothermal activity succeeding metamorphism.

Beneslavsky (1963) states the magnetite in bauxite to contain an Al_2O_3 substitution attaining up to 15% in places.

Ulvite, $(Fe^{2+}Ti^{4+})Fe^{2+}O_4$. The mixed crystals of ulvite and magnetite, described in earlier literature by the name *titanomagnetite*, have also been grouped here.

Known mostly in the form of clastic grains, the mineral is frequent among others in the Devonian bauxites of the Urals, USSR, where it forms grains of 0·01 to 0·2 mm in size accompanied by magnetite (Bushinsky 1971). It is present in the heavy-mineral fraction of the Jamaican bauxites also (Hartman 1955). It has been identified in a number of other, minor deposit groups. Its abundance is in the thousandth to hundredth of a percent range.

In Northern Urals bauxite, it also appears in the form of authigenic octahedral crystals; indeed, some well-developed crystals of ulvite have been observed on the walls of some smaller cavities and fissures (Beneslavsky 1963).

Chromite, Cr_2FeO_4. A rare mineral in bauxite, it is known to exist predominantly in the form of clastic grains. In Hungary, it is enriched in the Nézsa deposit group above all. Kiss (1952) found it to be the fourth most frequency heavy mineral of that bauxite; its abundance attained a tenth of a percent in places. Kiss (1955) also identified it in the Gánt bauxite; in addition to the clastic, worn grains, he also found there a few entirely undamaged, automorphic (octahedral) chromite crystals, which he ascribed to epigene processes. A few chromite grains of clastic origin were found by Vörös and Gecse (1976) in bauxite from Nagyegyháza and Iszkaszentgyörgy–Bitó.

Deposit Mandra II of the Eleusis–Mandra group in Greece contains 0·2 to 0·6% chromite (Papastamatiou and Maksimović 1970). The mineral is somewhat more abundant in the lower part of the deposit (0·4 to 0·6%) than in its upper part (0·2 to 0·4%). Minute quantities of clastic chromite grains are also ubiquitous in Jamaican bauxite (Strahl 1971).

In the course of their electron microprobe studies, Bárdossy and Pantó (1973) demonstrated clastic chromite grains of 3 to 25 μm in size in samples from the deposit called Pas de Recoux of the Brignoles group, France. The grains were embedded in iron-rich clastic bauxite particles, which is indicative of their redeposition. No chromite was detected by similar studies on other deposits of the Brignoles group. It is the Pas de Recoux deposit that is closest to the crystalline basement cropping out in the nearby Maures massif: this is where the chromite had presumably come from. The fact that chromite could only reach the deposit closest to the basement outcrops implies that the mineral cannot stand up to any longer transport.

Hercynite, Al_2FeO_4. The only place where this mineral has been identified so far is the contact metamorphic halo of a basalt dyke traversing the bauxite at Bédarieux in Southern France; its abundance ranges from 4 to 25% there. X-ray diffractometry by Bárdossy *et al* (1970) proved it to be very poorly crystalline, with a particle size less than one micrometre. Its formation is attributed to the thermal action of the basalt.

Goethite, α-FeOOH. This is the second most frequent iron mineral of karst bauxites after haematite. Its bulk is syn- and diagenetic; less abundant epi- and supergene goethite, is usually due to pyrite oxidation.

Bauxite includes a type of goethite in which part of the iron is replaced by aluminium. This variety is called *alumogoethite*, a term proposed by Beneslavsky (1958). Thiel (1963) succeeded in synthetising an alumogoethite in which the AlOOH substitution attained 33 mol %. He found the X-ray reflections of the product to be displaced in proportion to the degree of substitution. Solymár (1970) and Solymár and Jónás (1971) proved on synthetic samples of alumogoethite that both the peaks of the infrared absorption spectra and the endothermic peaks of the derivatograms are also displaced in proportion to the degree of substitution. Making use of these findings, they assayed alumogoethite contents and degrees of substitution in 50 Hungarian bauxite samples, and found the degree of substitution to be the higher, the higher the average goethite content of the bauxite: 7 to 30 mol % in the goethite of Iszkaszentgyörgy bauxite, 10 to

20 mol % in Gánt bauxite, 8 to 15 mol % in Szőc bauxite, and 5 to 15 mol % in Nyirád bauxite.

Janot and Gibert (1970) extended their investigations with the Mössbauer device, mentioned in connection with haematite, to goethite also. Checking a broad range of bauxite samples, they found the maximum of AlOOH substitution to be 23 mol %. They further found the goethite crystallites to be the smaller, the higher the degree of substitution. The finest crystallites are in the 100 to 200 Å size range.

The present writer in the course of his systematic mineralogic investigations on bauxite samples of different regions has found alumogoethite to be much more widespread in bauxites than chemically pure goethite. The latter tends to be confined to epigene textural elements (nests, concretions, incrustations and fissure fillings) above all. The degree of substitution in alumogoethite is most often in the 5 to 20% range. The highest degree observed by the writer, 32 mol %, was in the Quaternary and subrecent bauxites of the Pacific islands.

Taking the lead of Strunz (1970), the present writer does not regard *hydrogoethite*, mentioned by certain authors, as an independent mineral, but rather as a finely disperse goethite containing a great deal of adsorbed water.

Lepidocrocite, γ-FeOOH. A very rare, secondary mineral of bauxite; no incontrovertibly primary lepidocrocite has so far been identified in any karst bauxite sample. Lepidocrocite accompanied by haematite, magnetite and goethite, found in pisoids of the bauxite deposits of Porozhninsk on the Angara River, USSR, seems to be diagenetic (Medvedev 1971).

The mineral is epigene as a rule. In Hungary, the only deposit in which it has been identified is Eplény II. In the upper part of that deposit, there are dark violet haematitic concretions formed along the traces of some roots. The outermost and innermost parts of the concretions contain only 2 to 3% lepidocrocite. In between, however, it is enriched to 16%. It is one of the concretions that is shown in figure 98. Their textures imply these traces of roots to have been pyritic originally. In the writer's opinion (Bárdossy 1968a), the lepidocrocite found in them is the product of a very slow, tranquil oxidation of pyrite, as is also the haematite accompanying it.

In the Suvaja–Šolaja deposits of the Grmeć Mountains, Yugoslavia, Jurković (1965) identified pseudomorphs of lepidocrocite after pyrite. The mineral also forms very fine veins in multicoloured, reoxidised bauxite. There, too, it presumably is an epi- and supergene product of oxidation. In the Bijele Poljane deposit group of Crna Gora (Montenegro), Yugoslavia, Dragović (1966) found a few percent of lepidocrocite in a violet bauxite clay. In those deposits, grey pyritic bauxite is fairly widespread, and the violet bauxitic clay presumably had been pyritic originally. In the Pădurea Craiului deposit group, Rumania, a small quantity of lepidocrocite was observed in the vicinity of some pyrite grains by Papiu et al (1970), who attributed it to pyrite oxidation.

Combes (1969) found some secondary gibbsito–goethitic nests in an ochroid clay in the lower part of a bauxite deposit near Durban in the Ariège, France, to contain 2 to 3% lepidocrocite. He observed similar quantities also in a dark brown ferruginous crust enriched in goethite. Not far above these formations, the vestiges of a grey pyritic bauxite are found. Also in this instance, it seems plausible for the lepidocrocite to have formed by the slow oxidation of pyrite.

Siderogel, FeOOH + aq. This is the term used here after Strunz (1970) for amorphous ferrihydroxide gels. It is not to be confused with *limonite*, which is a mixture of submicroscopic crystallites of goethite and lepidocrocite. Limonite is a comprehensive term then which is to be used in field descriptions prior to a detailed mineralogical analysis.

Earlier, many authors used to be of the opinion that a significant portion of iron in bauxites is in the form of a ferrihydroxide gel. X-ray analyses in recent years have shown that only the youngest bauxites contain siderogel in significant quantities. In the bauxites of Lifou and Maré islands, Loyalty Group, and Niué Island, East Tonga Group, Pacific Ocean, e.g. the writer has found 2 to 10% siderogel. In the overwhelming majority of pre-Miocene bauxites, siderogel is absent or present in traces only.

Siderogel seems to be unstable, recrystallising into goethite or haematite as time passes, a process promoted by overburden pressure.

5.3.1.3. TITANIUM OXIDES AND HYDROXIDES

Anatase, TiO_2. This is the most widespread and most abundant titanium mineral of karst bauxites. Some of the Iranian deposit groups are strikingly rich in anatase (4 to 7% at Kerman, 4 to 6% at Yazd, 3 to 5% at both Bukan and Jajarm).

Most of anatase in bauxite is syngenetic, finely dispersed in the bauxite matrix, having a particle size of 0·1 to 1·0 μm. Using the electron microprobe, Bárdossy and Pantó (1970) proved authigenic anatase grains of 1 to 10 μm in size to be very widespread in various bauxites. These also are syngenetic or early diagenetic. Anatase is found in clastic bauxite particles, roundgrains, bauxite pebbles and boulders also. Independent grains of clastic anatase, on the other hand, are rare.

In the concentric particles of diagenetic origin, the abundance of anatase is most often somewhat less than, more seldom as great as, in the surrounding matrix. Most of the epigene textural elements contain either no anatase at all or just small quantities in films, coatings, nests, concretions and fissure fillings. One exception is the Pădurea Craiului deposit group in Rumania in which Bárdossy and Pantó (1970) using an electron microprobe observed veins of anatase and rutile 30 to 100 μm wide, of epigene origin. In the images of figure 137, the veins are clearly seen to avoid the prior texture elements.

Rutile, TiO_2. This is the second most frequent titanium mineral after anatase. Some of it is syngenetic, finely dispersed in the bauxite matrix, in particles of a few micrometres in size. Acicular authigenic rutile crystals of a few micrometres in size are also frequent, as observed e.g. by Beneslavsky (1963) in the Devonian bauxites of the Northern Urals, USSR, and by Kozlowsky (1956) in the Austrian bauxites.

The mineral is very widespread in the form of clastic grains of 10 to 200 μm in size. These are present in all the Hungarian deposit groups, as proved by the micromineralogical analyses of Kiss (1952, 1955), Vörös (1958, 1971), and Vendel *at al* (1971). They are frequent also in the bauxites of Jamaica (Strahl 1971), France (Demangeon 1965, 1970) and the Urals (Beneslavsky 1963) and also in a number of minor deposit groups. There are also small quantities in the clastic bauxite particles, roundgrains, bauxite pebbles and boulders of clastic origin.

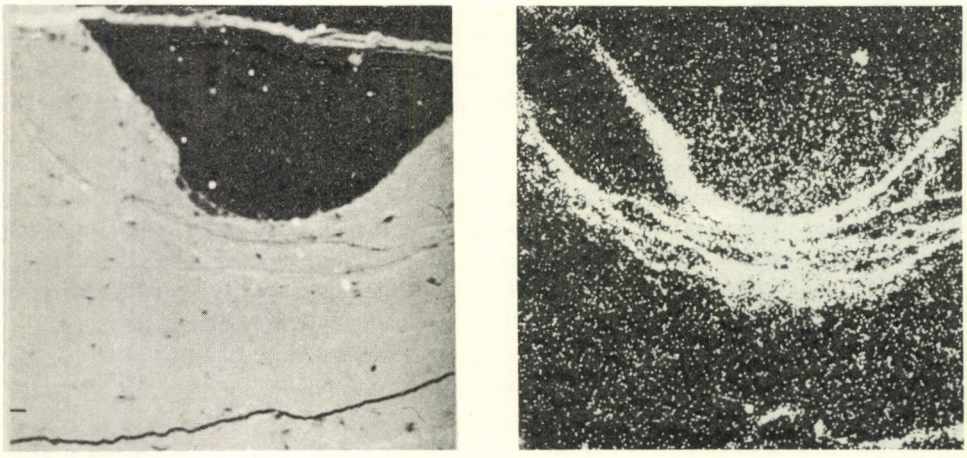


Figure 137. Anatase–rutile veins of epigene origin in bauxite. Pădurea Craiului, Rumania. Electron microprobe images, 268 ×
(A) electron image; (B) X-ray image of titanium distribution

Very rare as a secondary mineral, it is a decomposition product of ilmenite. In Hungary pseudomorphs of rutile after ilmenite were observed in Nézsa bauxite by Kiss (1952), and in bauxite from Nagyegyháza and Iszkaszentgyörgy–Bitó by Vörös and Gecse (1976). Vörös (1971) found pseudomorphs of rutile after titanohaematite in bauxite from Gánt. Beneslavsky (1963) observed sagenite grids (grids consisting of rutile) in decomposed ilmenite grains is the Devonian bauxites of the Northern Urals, USSR.

Mindszenty (1970) discussed issues of lattice structure energy connected with ilmenite decomposition. According to her, ferrous iron contained in ilmenite was oxidised and hydrated by oxygen-bearing ground water seeping through the bauxite, and was subsequently precipitated in the vicinity. The titanium left behind assumed a rutile structure while preserving the outer form of the ileminite crystal.

Rutile is not known from any nests, concretions or incrustations in bauxite. The Pădurea Craiului deposits repeatedly referred to above are the only group where it forms secondary veins together with anatase.

Brookite, TiO_2. Much rarer than anatase and rutile, it is known at present only in abundances of a few thousandths or hundredths of a percent, in the form of clastic grains. According to Kiss (1955), Gánt bauxite, Hungary, contains among others some brookite grains. Strahl (1971) states it to be present in Jamaican bauxite. Demangeon (1965) identified a few grains of clastic brookite in the Pierrerue deposit of the Languedoc, France. According to Beneslavsky (1963), the only deposits in the USSR in which it has been demonstrated so far are those of Tatarsk, where it forms minute platy crystals.

In a bauxite sample containing 7·05% TiO_2 from the upper part of the Arkalyk deposit group, received by courtesy of Lavrenchuk, the writer's X-ray examinations have found the titanium content to be distributed in the form of 84% anatase, 10% rutile

and 6% brookite. This implies a content of 0·4% brookite in the entire sample. The sample is a low-iron, clayey bauxite of pisoidic texture. This remarkable enrichment of brookite is presumably authigenic rather than clastic. This assumption is confirmed also by the extremely fine grain of the bauxite. Kessmann's laboratory experiments (1966) proved that brookite rather than anatase will form even at low temperatures in presence of alkalies (sodium in particular). The present writer deems it conceivable that, also in the sample discussed, a local concentration of alkalies in the ground water had promoted the formation of a little brookite besides anatase and rutile.

Pseudobrookite, $Fe^{3+}TiO_5$. Its grains of clastic origin were identified by Vörös and Gecse (1976) in Nagyegyháza bauxite, Hungary.

Ilmenite, $FeTiO_3$. This is the third most frequent titanium mineral after anatase and rutile, identified so far in 30 deposit groups altogether, most often in the form of clastic grains. It has been demonstrated in all the Hungarian deposit groups in quantities of a few tenths or hundredths of a percent. According to Kiss (oral communication) its abundance exceeds 1% in the bauxites of Óbarok, Újbarok, Vázsonypuszta and Nagyegyháza. Hartman (1955) states the heavy mineral fractions separated from Jamaican bauxites to contain 8 to 33% ilmenite. The particles are partly leucoxenised. In the Pădurea Craiului deposit group, Rumania ilmenite is strongly weathered as a rule; often, only pseudomorphs converted into rutile can be identified. In places, the nuclei of ooids and pisoids consist of ilmenite (Papiu *et al* 1970).

Clastic ilmenite grains have been identified in several deposit groups of Yugoslavia, e.g. in the Triassic bauxites of Lika Country and of Montenegro (Crna Gora) (Šinkovec 1970; Burić 1966), and in the deposits of the Grebnićka Planina (Misirlić and Mitrović 1969). In the USSR, ilmenite has been identified in the bauxites of the Northern Urals, of Tatarsk and of the Turgay Plateau. According to Beneslavsky (1963), the tabular crystals of ilmenite are more or less worn; locally they concentrate in a layer-like form. The ilmenite grains are often decomposed, leucoxenised.

Epigene ilmenite is considerably rarer. Komlóssy (1966) stated 70 to 80% of the elutriation residue of multicoloured bauxite underlying grey bauxite in the Kincses ore body at Iszkaszentgyörgy, Hungary, to consist of epigene ilmenite. He assumed it to have formed out of rutile, under the influence of solutions enriched in sulphuric acid and ferrous sulphate by pyrite oxidation. Mindszenty (1970) deems it conceivable on thermodynamic grounds for ilmenite to grow on under the influence of reducing epigene solutions of low pH seeping downward in bauxite. This assumption, however, has not been borne out so far by any microscopic observations.

Peyronnet (1970) identified substantial quantities of ilmenite in deposits to the north of Alanya in Turkey: 3 to 10% on average and even 13% in one sample in the bauxites of the southern, more strongly metamorphosed part of the region. In the northern, less metamorphic part of the region, on the other hand, he found no ilmenite in a majority of samples, and only 1 to 5% in the rest. This geographical distribution seems to imply the ilmenite to have been produced during the regional metamorphism of bauxite, together with corundum, disthene, chloritoid and pyrophyllite.

This assumption is confirmed by the present writer's X-ray examinations on bauxites in the vicinity of Remeţi, Rumania. These bauxites were metamorphosed by rhyolites and dacites erupted in the Neogene. In the original bauxite, anatase is the principal

titanium mineral, accompanied by less rutile and even less ilmenite, whereas, in the contact metamorphic bauxite, ilmenite is the most abundant of the three, and anatase and rutile are considerably scarcer.

Ferri-ilmenite. Following Strunz (1970), the present writer uses this term for isomorphous mixed crystals in which 1 to 33% of $FeTiO_3$ is replaced by Fe_2O_3. According to Mindszenty (1970), 70 to 90% of the 0·06 to 0·20 mm elutriation residue of Hungarian bauxites consists of ferri-ilmenite. The grains of ferri-ilmenite are worn in their majority, just as those of ilmenite, which implies their clastic origin. A minor percentage is unworn, automorphic: these are regarded as epigene. Vörös (1968) performed chemical and X-ray analyses on ferri-ilmenite grains from the Iszkaszentgyörgy deposit group, Hungary. He found them to contain 1·7% MgO and 0·6% MnO in addition to 63·4% Fe_2O_3 and 35·7% TiO_2; they thus represent a transition towards *geikielite.* Their X-ray reflections are significantly displaced against those of pure ilmenite.

A substantial percentage of the ferri-ilmenite grains is weathered, and transformed into rutile by weathering.

Perovskite, $CaTiO_3$. Beneslavsky (1963, 2nd ed. 1974) mentions it without any detailed description from the bauxites of the Northern Urals, USSR. It is known from no other deposit group at present.

Adopting Strunz's (1970) standpoint, the writer does not consider *leucoxene* an independent mineral. The term is a comprehensive one for any aggregates of different titanium minerals. In literature on bauxite, it is used most often to denote the decomposition products of ilmenite.

Lacroix (1913a) surmised the presence of an amorphous titanium hydroxide gel in the French bauxites; he named the phase *doelterite.* The present writer has not so far succeeded in identifying this mineral in any of the samples examined by him from either Hungary or abroad. In agreement with Strunz (1970), he deems it questionable whether this mineral exists in nature at all. Chemical investigations at any rate have proved that freshly precipitated amorphous titania gel rearranges very soon into anatase.

Geikielite, $MgTiO_3$. A few geikielite grains of clastic origin were identified in bauxite from Iszkaszentgyörgy, Hungary by Vörös and Gecse (1976).

5.3.1.4. OXIDES AND HYDROXIDES OF MANGENESE

Pyrolusite, β-MnO_2. Strahl (1971) states it to be a secondary mineral present in minute quantities in Jamaican bauxite.

Cryptomelane, $K_{<2}Mn_8O_{16}$. Known from a very few deposits so far. At Eplény, Hungary, it is of clastic origin, found in pebbles and lumps of manganese ore at the base of the bauxite deposits, as erosion product of nearby manganese ore deposits.

The writer has identified epigene cryptomelane in two deposit groups. He observed a sooty black, pulverulent porous segregation a few millimetres thick at the bedrock boundary in the Nazda deposit of the Vlasenica group, Yugoslavia. It consisted of

poorly crystalline cryptomelane accompanied by less abundant groutite. In the deposit named Krasta Pengile near Kruja in Albania, multicoloured bauxite contains gibbsite nests of fist to head size. The white, loose, porous gibbsite is traversed by jet-black veins a few millimetres wide. These consist of poorly crystalline cryptomelane accompanied by a little psilomelane. In both cases the remarkably poor crystallinity of cryptomelane has been found to be worthy of notice.

Hausmannite, Mn_2MnO_4. A frequent secondary mineral in Jamaican bauxite (Strahl 1971) it nowhere exceeds an average abundance of a few tenths of a percent.

Groutite, α-MnOOH. Strahl (1971) states it to be a secondary mineral present in minute quantities in Jamaican bauxite. In the Vlasenica deposit group, Yugoslavia, it accompanies epigene cryptomelane, which itself occurs in minute abundance.

Manganite, γ-MnOOH. Identified in Jamaican bauxite in abundances of a few tenths of a percent by Strahl (1971), who holds it to be of secondary origin.

Lithiophorite, $Li_2Al_8Mn_2^{2+}Mn_{10}^{4+}O_{35} \cdot 14H_2O$. The most frequent mangenese mineral of karst bauxites, it tends to concentrate in textural elements of secondary origin. In the La Boissière deposit group of Southern France, bauxite contains 1 to 3% lithophorite on average, together with smaller amounts of todorokite. The writer has found in this bauxite lithiophorite in segregations of a few millimetre in size and irregular form in the matrix, on the one hand, and in coatings covering the ooids, on the other. The deposits to the north of Cotignac in the Haut-Var deposit group of Provence, France, invariably contain 1 to 3% lithiophorite, partly disseminated in the matrix and partly as films a few micrometres thick on the ooids. The writer holds the lithiophorite of these two deposit groups to be partly of syn- and partly of diagenetic origin.

The writer found bluish-black pebbles of 2 to 10 mm in size in a bauxite of conglomerate texture near Hrast in Slovenia, Yugoslavia: the pulverised and homogenized sample of a larger number of pebbles gave 43% lithiophorite, 28% gibbsite, 13% haematite, 5% kaolinite, 4% goethite, 4% maghemite, 1% anatase and $^1/_2$% rutile. Lithiophorite in this deposit was incontrovertibly carried in from the outside in its present mineral form. It had presumably formed in some short-lived bauxite deposit, in the first phase of its primary reworking. Many of the bauxite pebbles in the deposit wear a film of black lithiophorite a tenth of a millimetre thick: those films are epigene, however.

Lithiophorite is ubiquitous in the reworked deposits around Lang Son in the north of Vietnam. The statistical evaluation of nearly 100 mineralogican analyses has given the following relationship between the sizes and lithiophorite contents of the bauxite pebbles and fragments:

+20 mm fraction	0·0 to 0·8%;	0·1% on average
10 to 20 mm fraction	0·0 to 1·5%;	0·9% on average
5 to 10 mm fraction	0·3 to 1·6%;	0·9% on average
2 to 5 mm fraction	0·4 to 3·1%;	1·2% on average.

Lithiophorite, then, is most abundant in the smallest particles. The primary, Palaeozoic bauxites of this region contain but very seldom a few tenths of a percent of

lithiophorite. The above-outlined concentration, then, is a phenomenon connected with the reworking of the bauxite.

In rust-red bauxite in the upper part of the Eplény deposits, Hungary, the writer has found in several places black nests of 1 to 2 cm in size, in which lithiophorite is enriched to 20 to 40%. It is accompanied by smaller quantities of todorokite. In the open working of Malomvölgy X lens at Halimba, Hungary, the writer has observed nests of 5 to 10 cm diameter similar to the above ones and also veins of irregular course and 1 to 2 cm width in the upper part of the brick-red bauxite. Their greyish- or bluish-black colour is in sharp contrast to the surrounding bauxite. These textural elements consisting of almost pure lithiophorite are also held to be epigene by the writer.

Lithiophorite is very frequent in the epi- and supergene crusts at the bedrock boundary. In Hungary, the best-developed lithiophorite-bearing crusts are those under the bauxitic clay deposits of Cserszegtomaj, which also contain a little todorokite. The crust may be up to 5 cm thick locally. In places, segregations of a few millimetres size of lithiophorite alternate with snow-white segregations of gibbsite. The writer found a similar lithiophorite-bearing crust under the Eplény deposits in Hungary (Bárdossy 1968a) and, in the course of his systematic examinations in recent years, also in the Nyirád, Szőc, Kislőd and Iszkaszentgyörgy deposit groups.

In Greece, in the Jurassic and Cretaceous bauxite deposits of the Parnassos and Kiona ranges, lithiophorite segregations similar in all particulars to those described above were identified by Bárdossy and Mack (1967). The present writer found similar ones also in the Helikon Mountains and in the Eleusis–Mandra deposit group (Bárdossy 1967). Under the latter deposits, the bedrock limestone is impregnated and dyed black by lithiophorite to a depth of 10 to 15 cm.

In the Combecave open pit, Brignoles deposit group, France, a black manganiferous crust at the bedrock boundary was found to consist of 44% lithiophorite and 56% kaolinite by Caillère and Pobeguin (1969). On the strength of this observation, they presumed the manganiferous crust observed in the other French deposit groups to also contain lithiophorite. The present writer's examinations of samples collected from the deposit groups of Villeveyrac, Loupian, the Alpilles, Brignoles, and the Haut-Var have confirmed this assumption. In several places, lithiophorite is accompanied, more seldom replaced, by todorokite.

Strahl (1971) in a letter to the writer stated lithiophorite together with todorokite to be the most frequent manganese mineral of Jamaican bauxites, concentrating most often in the proximity of the bedrock. This is confirmed by Sabiston (1975) who found Jamaican lithiophorites to contain less than 0·25% Li_2O but also quantities of a tenth of a percent of nickel, copper and cobalt. In the Lydford pit on St Ann Plateau, the present writer has observed black segregations a few centimetre thick at the bedrock boundary, which he has found to contain almost 40% lithiophorite, accompanied by gibbsite, goethite, boehmite, and crandallite.

In these crusts on the bedrock surface, the crystallinity of lithiophorite is highly varied. Poorly crystalline in most cases, it is fairly crystalline in some others. Its typical particle size is a few micrometres.

Todorokite, $(Mn^{2+}, Ba, Ca)_2Mn_6^{3+}Mn_{11}^{4+}O_{33} \cdot 8H_2O$. Occurring most often together with lithiophorite but in smaller quantities, it has been indentified in bauxites

from the La Boissière deposit group, France, in Jamaican bauxite, and in the epigene manganiferous nests of Eplény bauxite, Hungary.

It is found most often in the manganiferous crusts at the bedrock boundary, likewise in the company of lithiophorite. For example, Valeton (1970) found a black manganiferous incrustation 10 cm thick at the bedrock boundary in a deposit called La Rouge near Brignoles, France; it consisted of todorokite and less abundant kaolinite. In a deposit called L'Issart-Rouge of the Bédarieux deposit group, France, the writer has observed a black manganese-rich incrustation 3 to 5 cm thick, the only manganese mineral of which is todorokite, at the boundary of the underlying dolostone. It is accompanied by 40 to 60% calcite.

Todorokite is formed in all probability simultaneously with lithiophorite as a product of epigene processes. The reason why lithiophorite is the preferred product in some cases and todorokite in some others is not well understood.

Woodruffite, $(ZnMn^{2+})_2Mn_5^{4+}O_{12} \cdot 4H_2O$. A secondary mineral in Jamaican bauxite (Strahl 1971), it is of rather a resticted distribution. It is known from no other locality.

Psilomelane, $BaMn^{2+}Mn_8^{4+}O_{16}(OH)_4$. Epigene, it accompanies cryptomelane in an abundance of a few percents in gibbsite nests in the Krasta Pengile deposit near Kruja in Albania.

Rancieite, $(Ca_{0.55}Mn_{0.22}^{2+}Mg_{0.04})\ 0{\cdot}4MnO_2 \cdot 3{\cdot}66H_2O$. It has been identified by Bárdossy and Brindley (1978) at the bottom of a bauxite deposit near Itea, Greece, where it forms a black crust of 10 to 20 cm thickness at the surface of the footwall limestone. It is presumed to be of epigene origin by precipitation from Mn and Ca containing low pH solutions by the neutralising effect of the limestone.

5.3.1.5. OTHER OXIDES AND HYDROXIDES

Takovite, $Ni_5Al_4O_2(OH)_{18} \cdot 6H_2O$. Red bauxite in the bottom part of the deposit in the Codouls open pit near Brignoles, France, has yielded some concretions of a few centimetres in size. Off-white in the middle, they gradually assume a greenish tint further outward. X-ray diffractometry by the writer has revealed a concentration of boehmite at the centre and of takovite further outward. The writer holds them to be epigene. In a letter to the writer, Gsell (1968) states takovite to have been identified also in the nearby Recoux deposit and also in the Mazaugues deposit to the south of Brignoles. Dietrich *et al* (1976) identified takovite in the form of minute greenish nests accompanied by boehmite, kaolinite and erythrine in the nearby Blanquette deposit.

In Central Greece, in the Turonian bauxite horizon of the Kiona Range, on the boundary between the bauxite and its bedrock, a hard greenish-blue incrustation a few millimetres thick was found by Bárdossy and Mack (1967). It consisted of pure takovite of fair crystallinity. The present writer holds it to be epigene, assuming that a ground water of acidic pH, seeping through the bauxite, dissolved all its finely disperse nickel content and then precipitated it at the bedrock contact, presumably as a result of an abrupt rise of pH, triggered by the influence of the limestone bedrock.

Nickel hydroxide, $4Ni(OH)_2 \cdot NiOOH$. This mineral, first described from a Canadian serpentinite by Jambor and Boyle who gave it no name, figures without a name also in Strunz's 'Mineralogische Tabellen'. It was identified by Durozoy *et al* (1966) in the nickeliferous concretions of the above mentioned open pit of Codouls. Nicolas and Kotschoubey (1972), re-examining some concretions from Codouls, confirmed the mineralogical determinations of Durozoy and his co-workers. The present writer, examining the concretions, has found some of them to contain some nickel hydroxide, the rest being made up of unmistakable, takovite. The two related minerals presumably occur in association here. Both are epigene. Nicolas and Kotschoubey surmise the influence of a reducing ground water of acidic pH, whose reducing character they ascribe to organic matter or pyrite.

Baddeleyite, ZrO_2. Bárdossy *et al* (1977), using the electron microprobe, found a few rounded baddeleyite grains of 2 to 15 μm in size and a clastic appearance embedded in iron-rich bauxite particles in the Monte Turchio and Monte del Morrone deposits of Italy. The mineral has not been found in any other deposit group so far.

5.3.2. SILICATES

5.3.2.1. NESOSILICATES

Garnets. Mineral grains of exclusively clastic origin in bauxite, their abundance is a few thousandths or hundredths of a percent. They are determined by micromineralogical (microscopic) means. In Gánt bauxite, Hungary, a few pale pink or colourless garnet grains were identified by Kiss (1955), and a few grains of *spessartite* by Vörös (1971). According to Vendel *et al* (1971), it is a rare clastic mineral in the bauxites of Nyirád, Szőc, Halimba, Kislőd, Eplény, and Iszkaszentgyörgy. They observed only slight enrichment of garnet grains in a few samples from Malomvölgy XI/A lens, Halimba. Antal (1973) identified the mineral in several bauxite samples from the Rákhegy ore body, Iszkaszentgyörgy, Hungary.

Garnet was identified in Jamaican bauxite by Hartman (1955), in Lika County bauxite, Yugoslavia, by Šinkovec (1970), and in the deposits along the Angara River in Siberia, USSR by Medvedev (1971).

Zircon, $ZrSiO_4$. Found in the bauxite largely in the form of clastic grains, it is very widespread and even enriched upto a few tenths of a percent in some deposit groups, thanks to its resistance to wear and weathering. Micromineralogical investigations have shown it to be of 50 to 200 μm particle size typically, 1000 μm as a maximum. Pantó and the writer, using the electron microprobe, demonstrated that zircon grains of 1 to 50 μm in size were also frequent in bauxite (fig. 74). The nuclei of ooids and pisoids often are clastic zircon grains of a few micrometres in size.

Zircon is frequent in every Hungarian deposit group. It is most enriched at Nézsa, where Kiss (1952) states it to be the most frequent heavy mineral. Its grains are of 200 to 1000 μm in size there; its slightly worn, columnar crystals often contain inclusions (fig. 138). The Rákhegy ore body at Iszkaszentgyörgy, Hungary, contains according to Antal (1973) in addition to the worn zircon grains also a few acicular, automorphic zircon grains, which holds to be epigene.



Figure 138. Zircon grains, columnar with inclusions, elutriated out of a bauxite, with a pyrite grain at bottom centre, 50×. After Kiss (1952)

Zircon is especially abundant in the Triassic deposits of Lika County, Yugoslavia (Šinkovec 1970). It is known also from the deposits of the Northern Urals, USSR (Beneslavsky 1961, 2nd ed. 1974).

Bárdossy and Pantó (1973) using the electron microprobe proved that the zircon grains of some bauxite samples from Brignoles, France, and Campo Felice, Abruzzi, Italy, contained small amounts of yttrium, scandium and phosphorus as isomorphous substitutions, implying a passage towards xenotime. More recently, Bárdossy and Pantó have identified such grains of zircon in bauxite from Nézsa and Nagyharsány, Hungary. They are of 1 to 5 μm in size as a rule.

Andalusite, $Al_2(SiO_5)$. Clastic grains in very small abundance in the deposit called Les Cannonettes of the French Alpilles (Demangeon 1965) and in the Bagolyhegy deposit at Gánt, Hungary (Vendel *et al* 1971).

Disthene (kyanite), $Al_2(SiO_5)$. It is found in the form of clastic grains. Vendel *et al* (1971) demonstrated very small quantities at Nyirád, Halimba, Kislőd, Eplény and Gánt, Hungary. According to Demangeon (1965), it is present in most French deposit groups, being most abundant in the Alpilles and around Saint Chinian. Demangeon derives it from the erosion of the Massif Central crystalline.

Regional metamorphism is known to have produced authigenic disthene in bauxite. Automorphic disthene crystals of 70 to 170 μm in size were observed by Peyronnet (1970) in the weakly metamorphic bauxite deposits to the north of Alanya in Turkey, in an abundance of a few tenths of a percent. Sharova and Glazunova (1966) found some 8% disthene in a sample of black diasporo–chamositic bauxite from Vietnam, likewise the product of a weak metamorphism. Emery on Naxos Island, Greece contains some 5 to 10% disthene (Papastamatiou 1951).

Staurolite, $2Al_2(SiO_5) \cdot Fe(OH)_2$. Found in clastic grains, it is less widespread than disthene. It occurs in Malomvölgy X lens, Halimba, Hungary (Vendel *et al* 1971).

Grains of staurolite were found by Antal (1973) in Iszkaszentgyörgy bauxite (Rákhegy ore body) and by Vörös and Gecse (1976) in Nagyegyháza bauxite.

Staurolite is formed also by the regional metamorphism of bauxite. In Naxos Island emery, for example its abundance exceeds 1% (Papastamatiou 1951).

Topaz, $Al_2(OH, F)_2(SiO_4)$. Identified by Önay (1949) in the strongly metamorphic bauxites and emeries of the Menderes Range, Turkey, it forms minute automorphic (prismatic) grains accompanied by diaspore crystals. Minute aggregates of topaz are found in places in the fissures traversing the rock.

Titanite, $CaTi(SiO_5)$. Comparatively rare as a clastic component, it has been identified so far only in the form of a few grains (Nyirád–Darvastó and Halimba–Malomvölgy among the Hungarian deposits; Vendel *et al* 1971). It was identified also in the Grebnička Planina deposit group of Yugoslavia (Misirlić and Mitrović 1969) and in Jamaican bauxite (Strahl 1971).

Authigenic titanite grains 50 to 200 μm in size were found by Šinkovec (1970) in the Lika County bauxites of Yugoslavia. Authigenic titanite forms fine veins in places. Its grains are frequent in the Devonian bauxites of the Northern Urals, USSR, where they form comparatively large, envelope-shaped crystals. Titanite is a frequent decomposition product of ilmenite, in which case it is found together with rutile and calcite. In the Ivdel deposits of that region, diaspore-filled veins locally contain large idiomorphic crystals of titanite (Beneslavsky 1958, 1963).

Chloritoid, $Fe_2^{2+}AlAl_3[(OH)_4O_2(SiO_4)_2]$. One of the products of the regional metamorphism of bauxite. The partly metamorphic bauxite in the Vanoise mountain group, France, contains e.g. 50 to 70% chloritoid, predominantly microcrystalline and evenly disseminated in the rock. It also occurs in the form of larger greenish crystals in places (Ellenberger 1955). It is accompanied by pyrophyllite, ferrocarpholite, diaspore, haematite and muscovite according to Goffé and Saliot (1977). In places, chloritoid is displaced by paragonite.

In the Aegean, there is 2 to 13% chloritoid in the partly metamorphised bauxite on Samos Island (de Lapparent 1937), but none in Naxos Island emery. In the emeries of the Menderes Range, numerous analyses by Önay (1949) demonstrated an average chloritoid content of 12%. There are some chloritoid-rich emeries in which it is enriched to 30 to 70%. Önay assayed 10 to 40% chloritoid in the less strongly metamorphic, diasporic bauxites. He observed under the microscope that the chloritoid formed tabular automorphic crystals. There are, in places, dark green chloritoid crystals 5 to 6 cm in size in the fissures of the rock.

In the Western Taurus, Turkey, north of Alanya, there are bauxites of weak and middling metamorphism. These contain according to Peyronnet (1970) 50% chloritoid on average and even up to 80% in some samples. Partly disseminated in the matrix, and partly found in the interiors of ooids, they also form frequent chloritoid-filled fissures. Sharova and Glazunova (1966) found 1 to 25% chloritoid in some weakly metamorphic bauxites from Vietnam.

In the Obukhovsk deposit group of the Salair region, USSR, there are bauxites which were exposed to contact metamorphism in the proximity of some igneous intrusions. These also contain several percent chloritoid accompanied by corundum.

diaspore and margarite (Beneslavsky 1963). The mineral was identified in small quantities also in the weakly metamorphic bauxites of the Pamir Ranges, USSR (Konnov 1972).

5.3.2.2. SOROSILICATES

Epidote, $Ca_2(Fe^{3+}, Al)Al_2[(O, OH)SiO_4 \cdot Si_2O_7]$. It is found in the form of clastic grains according to Vendel *et al* (1971) in the bauxites of Nyirád–Darvastó, Nyirád–Dűlt-Nyires and Halimba–Malomvölgy, and rare in the bauxites of Szőc, Halimba, Kislőd, Eplény, Iszkaszentgyörgy and Gánt, Hungary. Vörös and Gecse (1976) identified it in bauxite from Nagyegyháza, Hungary.

A few grains of epidote were found by Karanov (1965) in the bauxite at Tran in Bulgaria. It was also identified in the bauxites along the Angara River in Siberia, USSR (Medvedev 1971). It is a rare clastic mineral in the Southern Timan deposit group (Demina 1971).

Clinozoisite, $Ca_2Al_3[(O, OH)SiO_4 \cdot Si_2O_7]$. A very rare clastic mineral, it is found occasionally in the bauxites of Nyirád–Dűlt-Nyires, Szőc and Eplény, Hungary (Vendel *et al* 1971).

Orthite, allanite, $(Ca, Ce)_2Fe^{2+}, Fe^{3+}Al_2[(O, OH)SiO_4 \cdot Si_2O_7]$. A very rare clastic mineral, it is sporadic in the bauxites of Nyirád–Dűlt-Nyires and Nyirád–Darvastó, Halimba–Malomvölgy, Kislőd, and Gánt–Harasztos, Hungary (Vendel *et al* 1971).

Zoisite, $Ca_2Al_3[(O, OH)SiO_4 \cdot Si_2O_7]$. A very rare clastic mineral. According to Vendel *et al* (1971), it is found occasionally in the deposits of Halimba Malomvölgy, and very seldom in Nyirád and Szőc bauxite.

In the Southern Timan deposit group, it is known as a rare clastic mineral (Gulyanitsky 1969).

Vesuvian, $Ca_{10}(Mg, Fe)_2Al_4[(OH)_4(SiO_4)_5(Si_2O_7)_2]$. A very rare clastic mineral, it is found on occasion in the bauxites of Nyirád–Darvastó, Szőc, Kislőd, Eplény and Gánt–Bagolyhegy; it is more frequent in the lenses of Halimba–Malomvölgy (Vendel *et al* 1971).

5.3.2.3. CYCLOSILICATES

Beryl, $Al_2Be_3(Si_6O_{18})$. A rare clastic mineral in bauxite, it was demonstrated in an abundance of a few hundredths of a percent by Kiss (1952, 1955) in Nézsa and Gánt bauxite, Hungary.

Tourmaline, $Na(Mg, Fe^{2+}, Mn, Li, Al)_3Al_6[(OH)_4(BO_3)_3 \cdot (Si_6O_{18})]$. A widespread clastic mineral, known from all the Hungarian bauxite deposit groups, it is comparatively scarce in the Halimba III ore body, and most abundant in Eplény bauxite (Vendel *et al* 1971). It is one of the most frequent clastic minerals of French bauxites. In the deposit groups of the Languedoc, its grains may be up to 500 and even 1000 μm in long, but they are of only 150 μm in size in the Alpilles (Demangeon 1965). According to Nicolas (1970), tourmaline grains are especially enriched in the

Mazaugues deposit near Brignoles, in an ooidic bauxite layer about 25 cm thick in the middle of the deposit. The grains are remarkably long (up to 1·0 mm).

Tourmaline has been identified also in most Yugoslav and Jamaican bauxites (Strahl 1971). According to Schüller (1957), it is also frequent in the bauxites of the Gun District, China.

Authigenic acicular tourmaline grains forming spherolite-like aggregates were found by Orlov (1937) in a small bauxite deposit near Mojtin in Slovakia. Papiu *et al* (1971) state the bauxite around Haţeg in Rumania to contain besides clastic tourmaline some authigenic grains of 40 to 60 μm in size.

Tourmaline could also be formed from the metamorphism of bauxite. In the Aegean, de Lapparent (1937) demonstrated its presence in bauxite of medium metamorphism on Samos Island; it forms fissure fillings in Naxos Island emery (Papastamatiou 1951).

5.3.2.4. INOSILICATES

Pyroxenes. Rare and exclusively clastic in bauxite, their abundance attains some hundredths of a percent only. Owing to the small size and usual decomposed state of the grains, it is often impossible to assign them with any reliability to this or that member of the pyroxene group.

Of the Hungarian deposit groups, it has so far been demonstrated in very small amounts in the bauxites of Nyirád–Darvastó, Halimba–Malomvölgy, Kislőd, Eplény and Gánt (Vendel *et al* 1971). In a small bauxite deposit near Cusano Mutri in the Monti Matese in Italy, the writer has found a pseudomorph of boehmite after pyroxene (fig. 73). In the figure, the characteristic cleavage lines of pyroxene, preserved by the pseudomorph, are clearly visible.

Amphiboles. Also rare in bauxite, they are exclusively clastic. Owing to their fine grain size, the members of the amphibole group cannot be distinguished. Amphiboles were identified by Vendel *et al* (1971) in bauxites from Nyirád–Dűlt-Nyires, Szőc, Halimba, Kislőd, Iszkaszentgyörgy and Gánt. They are most abundant in the deposits of Halimba–Malomvölgy. They were also found by Vörös any Gecse (1976) in Nagyegyháza bauxite.

Tkachenko and Bol'shun (1971) found a few grains of amphibolite in the Bogoslovsk deposit of the Urals, USSR. Duboul–Razavet and Périnet (1960) identified as *glaucophane* the few grains of amphibole found in the bauxite of the French Alpilles.

Magnesiokarpholite, $(Fe^{2+}_{0\cdot3}Mg_{0\cdot7})(Al_{0\cdot95}Fe^{3+}_{0\cdot05})Si_2O_6(OH)_4$. This mineral was identified by Goffé *et al* (1973) in bauxite from the Vanoise, France. The mineral is higher in magnesium than the ferrokarpholite described by Roever in 1951 from Sulawesi (Celebes). The bauxite of the Vanoise contains it in the following mineral association: chloritoid > pyrophyllite > paragonite > muscovite > magnesiokarpholite > diaspore > haematite > chlorite. Found in prisms of 100 to 200 μm in length, it is the product of the low-temperature, high-pressure metamorphism of bauxite, at about 360 °C and 4 to 6 kbars, corresponding by and large to the lawsonite–pumpellyite metamorphic facies.

5.3.2.5. PHYLLOSILICATES

(A) Kaolinite group

Kaolinite, $Al_4(OH)_8(Si_4O_{10})$. The most widespread clay mineral of karst bauxites, it is found in almost all textural elements. The only type of texture not encountered anywhere so far is the pure kaolinite concretion; in fact, most concretions contain very little kaolinite, if any.

Kaolinite contained in bauxite may exhibit a variety of degrees of lattice ordering, from the highly crystalline, ideally ordered type to types that are more or less disordered along the *b* axis. According to the writer's X-ray diffractometric and derivatographic measurements, most of the kaolinite in Hungarian bauxites is of a degree of medium ordering in its bulk. Bauxitic clay from Cserszegtòmaj is especially disordered in the *b*-axis direction. It is the epigene segregations of kaolinite that are best-ordered.

According to Périnet (1964), in the Peygros deposit near Brignoles, France, kaolinite is highly crystalline in the bauxitic clay at the bedrock boundary, and of medium crystallinity in the bauxite proper. He states the kaolinite lattice to be the more ordered, the less the organic content of the bauxite, without giving any explanation of this relation.

Epigene segregations of kaolinite prepared from several samples of Yugoslav, French and Spanish bauxites have been found by the writer to be highly or very highly ordered. The kaolinite contained in the surrounding bauxite matrix, on the other hand, was of mediocre ordering.

Prokaolin, described by Karšulin (1953) from Yugoslav bauxite, is according to recent investigations not an independent mineral, but a kaolinite of very poor crystallinity.

Dickite, $Al_4(OH)_8(Si_4O_{10})$. In bauxite, it is found rarely and in small quantities. Beneslavsky (1963) identified a little dickite in the Boksonsk, Tatarsk and Krasnooktyabrsk deposits of the Northern Urals, USSR, and also in the Gun-syan deposit of China. He connected its formation with the sideritisation and chloritisation of bauxite (i.e. with its reduction to use a more general term). Papiu *et al* (1971) found 10 to 20% dickite accompanied by kaolinite in the more clayey parts of the Haţeg deposit group, Rumania. In one sample, dickite was enriched to 89%. It is held to be the product of muscovite decomposition there. Karapetov (1968) demonstrated 25 to 30% dickite in some of the deposits of the Pamir Ranges, USSR.

All the occurrences of dickite in Hungarian bauxites are epigene. Barnabás (1957) e.g. found minute white nests in yellowish-brown clayey bauxite in the upper part of the Cseres deposit, Halimba, which were proved by X-ray and DT methods to consist of a mixture of dickite and kaolinite. The present writer found a white porous friable layer at the bedrock boundary of the bauxite in the open pit of Sümeg–Szőlőhegy, which he proved to contain some 40% dickite accompanied by 50% kaolinite (Bárdossy 1961c). The causes and conditions of epigene dickite formation have not been elucidated as yet.

Nacrite, $Al_4(OH)_8(Si_4O_{10})$. Even rarer in bauxite than dickite, it was identified by Beneslavsky (1961, 2nd ed. 1974, 1963) in the Northern Urals, USSR. He also found thin films and coatings of nacrite, accompanied in places by even less abundant dickite,

on parting planes in bauxite in the Boksonsk deposits of the USSR and in the Sao Shi and Fu Shun deposits of China. These segregations are clearly epigene.

Metahalloysite, $Al_4(OH)_8(Si_4O_{10})$. A rare mineral in bauxite, it has been identified in eight deposit groups so far. In Portugal, in a deposit of bauxitic clay near Andorinha, predominant kaolinite is accompanied by some 10 to 20% metahalloysite (Figueredo Gomes 1965). The present writer has found 4 to 10% metahalloysite in association with 10 to 20% kaolinite in the bauxite matrix in a small deposit near Hrast, Slovenia, Yugoslavia. In the middle part of the deposit, there is an interbedding, 30 cm thick, of well-bedded bauxite-like clay, which grades upward and downward into bauxite (colour figure 13). The writer has found 30 to 40% metahalloysite and 10% kaolinite in the zone of transition and 80 to 96% metahalloysite in the well-bedded bauxite-like clay which contained no kaolinite at all. In some of the beds, 4 to 5% halloysite also appeared.

Ščavničar *et al* (1968) found a few percent of metahalloysite accompanying the principal clay mineral kaolinite in four of 47 bauxite samples from around Mostar, Yugoslavia. Luković (1964) observed small amounts of metahalloysite in bauxites exposed to secondary resilification in the Nikšićka Župa deposit group, Crna Gora (Montenegro), Yugoslavia. In an ochroid ooidic bauxite from the Tran deposit group of Bulgaria, the writer has identified 5% metahalloysite in association with 15% kaolinite.

The Quaternary bauxites of the Pacific islands contain 0·5 to 2% metahalloysite accompanying kaolinite and halloysite. The writer has demonstrated metahalloysite as a supergene formation in an abundance of 3 to 6% formed out of the chlorite and chamosite contained in the original Paleozoic bauxite, in several bauxite samples from the reworked bauxite deposits of Vietnam. Metahalloysite is confined to the oxidised outer parts of the grey and green bauxite pebbles. The composition of the weathering products of such a bauxite pebble is shown in table 39.

Halloysite, $Al_4(OH)_8(Si_4O_{10})\cdot(H_2O)_4$. Bauxites with halloysite as their only clay mineral make up a mere 0·1% of total world karst bauxite tonnage. The tonnage of the bauxite which contains some halloysite besides kaolinite has been estimated at about 10%. These are the youngest karst bauxites to remain unburied ever since their origin (cf table 22).

Halloysite is the most widespread silicate mineral in the Tulsk–Ryazan deposit group. In certain samples, it is the product of slow crystallisation of amorphous allophane. The writer holds this very friable, porous bauxite to be of Pliocene or at most Quaternary age on the strength of its form of deposit and petrographic features. The above-mentioned age rule, then, also holds here.

The writer has observed some epigene nests of halloysite in yellow bauxite in the upper parts of the Halimba–Malomvölgy deposits, Hungary. The nests are of a few centimetres diameter, snow-white and friable. Halloysite is accompanied by somewhat smaller percentages of alunite (Bárdossy 1966).

Halloysite is also found in the epigene incrustations at the bedrock boundary. For example, the compact, snow-white nests of soapy touch and conchoidal fracture, of mammilliary and reniform shape and 1 to 5 cm size, found in places in a gibbsite incrustation at Cserszegtomaj, Hungary, consist of pure halloysite (Bárdossy 1961b).

Pure halloysite nests identical in position and appearance to those at Cserszegtomaj were fond by Bárdossy and Mack (1967) in the Upper Cretaceous bauxite deposits of the Kiona Range, Greece. Electron micrographs revealed the halloysite crystals to form tubules 0·3 to 1·0 μm long and 0·05 to 0·1 μm in diameter, in good agreement with the halloysite forms described in literature.

Clarke (1966) described fibrous segregations of gibbsite and halloysite from the top of the bedrock limestone in the Eufaula deposit, Southeastern Coastal Plains, US. The writer holds it probable that more detailed mineralogical investigations will identify halloysite in many more deposit groups, in the form of epigene segregations and nests at the bedrock contact.

(B) Serpentine group

According to Beneslavsky (1961, 2nd ed. 1974), serpentine forms epigene fissure fillings in Boksonsk bauxite, USSR. Metamorphism has given rise to a fissure grid filled with serpentine and mica in the Salair bauxites, USSR. Beneslavsky did not identify any closer the members of the serpentine group represented in those deposits.

(C) Septechlorite group

Chamosite, $(Fe^{2+}, Fe^{3+}, Mg)_6\,(OH)_8(Si_2Al_2O_{10})$. Identified in 33 deposit groups so far, this mineral has a typical abundance in the 1 to 5% range, and exceeds 20% only in two deposit groups, in the Triassic and Permian deposits near Jajarm in Iran.

In certain deposits it may be substantially enriched, in the grey, green and dark brown bauxite types above all, attaining 40 to 60% on occasion. Even in those bauxite types, however, chamosite is but seldom the only clay mineral: it is most often accompanied by kaolinite, more seldom by leptochlorites, as was revealed by the tables in Section 5.2.2.3.

In the Hungarian bauxites, chamosite is very rare. Its only significant enrichment is known from the Újfeltárás open pit at Gánt. A very hard greenish iron-rich bauxite is found there in a thickness of 0·5 to 2 m in the top part of the deposit: it grades downward into a yellowish-brownish bauxite. This greenish bauxite type of 10 to 15% silica content has been proved by the writer's X-ray investigations to contain 30 to 50% chamosite. Chamosite largely concentrates in the bauxite matrix, where it may attain 50 to 80%. On the other hand, the bauxite pebbles in this bauxite of conglomerate texture, contain but 5 to 20% chamosite. The bauxite is boehmitic with a little gibbsite. The greater the abundance of chamosite, the less gibbsite there is; the boehmite content, on the other hand, is hardly reduced. It is the writer's general experience that the bauxite types containing chamosite are mainly diasporic or diasporo–boehmitic. No gibbsito–chamositic bauxites have been identified, so far ar least.

Within the Mediterranean bauxite belt, chamosite–rich types of bauxite are most widespread in Yugoslavia, Greece and Rumania. At the bottom of the Middle Triassic bauxite deposits near Ljubljana in Yugoslavia, e.g. there are a few metres of grey and greenish-grey clayey bauxite, revealed by the writer's investigations to contain 30 to 40% ferrochamosite accompanied by 10 to 30% kaolinite. In a similar stratigraphic position, greenish-grey clayey bauxite and bauxitic clay are also found at the bottom of the Lika County bauxites, Croatia, Yugoslavia. In samples collected there, the writer

has found 20 to 50% chamosite. At both sites, the mineral concentrates in the matrix above all.

In Greece, in the Turonian bauxite horizon of the Parnassos and Kiona ranges, bauxite types of greenish colour are found in places between a grey pyritic and a red haematitic bauxite. The writer demonstrated 30 to 40% chamosite in the greenish bauxite. In the adjacent bauxite types it is entirely missing, its place being taken by kaolinite (Bárdossy and Mack 1967). In some greyish-brownish bauxite samples from the Lower Malm bauxite horizon around Distomon, the writer has also found 30 to 40% chamosite. In the top part of the deposit called Kephanedes in the Helicon Range, the writer has observed a vivid green bauxite which has turned out to contain 10 to 40% chamosite, accompanied by a few percent of orthochlorite and kaolinite (Bárdossy 1967).

In some deposits of the Pădurea Craiului deposit group, Rumania, there is at the cover boundary a greenish, greenish-grey bauxite 10 to 30 cm thick, which has been shown by the writer to contain 50 to 80% ferrochamosite, accompanied by a little kaolinite, some diaspore and boehmite, and a little goethite. Remarkably enough, chamosite is enriched there in the abundant ooids as well as in the matrix.

There are high concentrations of chamosite in the green and grey reduced bauxite types of the Devonian deposits in the Urals, USSR, where chamosite is accompanied by kaolinite on the one hand and leptochlorites on the other. It concentrates in addition to the matrix also in the pisoids and ooids of diagenetic origin (Sharova and Rudnova 1964). In the greenish, greyish and black bauxite types of the Permian deposits of Vietnam, the writer identified 40 to 60% chamosite, accompanied most often by a few percent of kaolinite. The chamosite is enriched in the matrix and in the ooids as well.

In most cases, then, chamosite is finely disperse in the matrix, as a syngenetic mineral. It is somewhat less frequent as a diagenetic mineral in the ooids and pisoids, and rare as an epigene formation. Epigene chamositisation was described from Chusovsk District on the western slope of the Urals by Gorbachev (1966a). The drill samples studied by him were from the 1700 to 2200 m depth interval. Chamosite was accompanied in them by kaolinite, diaspore and boehmite. Gorbachev inferred the chamosite to be the product of a reaction, under reducing conditions, of iron-rich bauxite with the silica content of ground water.

No chamosite is formed by supergene processes. On the contrary, chamositic bauxite types tend to decompose gradually in the outcrops, and to oxidise to kaolinite, haematite and goethite (tables 38 to 40).

Strigovite, $2(Fe^{2+}, Mg)O \cdot (AlFe^{3+})_2O_3 \cdot 2SiO_2 \cdot 2H_2O$. According to Serdyuchenko (1953), this mineral is transitional between chamosite and delessite. Sharova and Rudnova (1964) found it to be one of the principal chlorite minerals of the Southern Urals bauxites, USSR. Sapozhnikov (1973) states strigovite to be present in the 'diasporo-chamositic rocks' (essentially clayey bauxites and bauxitic clays) into which the high-grade bauxite deposits grade laterally. Found both in the matrix and in the ooids and pisoids, it is hard to distinguish by X-ray means from ubiquitous chamosite. Its abundance ranges from 30 to 50%. Sapozhnikov holds it to be partly of syngenetic and partly of diagenetic origin. These bauxitic clays had deposited under reducing conditions in the sea, rather close inshore, whereas the pure bauxite deposits,

formed along the same coastline, had deposited while still on dry land. The composition and pH of sea water and the reducing environment presumably were favourable to strigovite formation. This septechlorite mineral is known from no other deposit group so far.

Nimesite, $(Al, Fe^{2+}, Fe^{3+}, Ni, Mg, Ti)_{2,6}\ (Si, Al)_2O_{4,5}\ (OH)_{4,5}$. De Weisse (1967) found a nickel-rich green bauxitic clay at the bedrock boundary in a bauxite deposit near Marmara, Greece. From it, Maksimović (1972, 1973) described a new nickel-rich septechlorite which he called nimesite to indicate the proximity of its lattice to that of amesite. The mineral is characterised by a vivid green colour and an extremely fine grain size.

A lateritic weathering product of nearby ultrabasic rocks, the bauxite had a much higher-than-average nickel content to start with. During bauxitisation in the karst dolines, part of the Al, Fe, Si and Ni was leached out and carried downward in solution. At the bedrock boundary, the transition into an alkaline pH resulted in the precipitation of nimesite. The mineral is not known to occur in any other bauxite.

(D) Chlorite group

Dioctahedral subgroup:

Donbassite or dioctahedral aluminium chlorite (sudoite)

$Al_2[(OH)_2AlSi_3O_{10}] \cdot Al_2(OH)_6$. A total number of 11 occurrences of this mineral have so far been identified in karst bauxites. It is also represented very probably in other deposit groups, but it is fairly hard to identify especially in the presence of trioctahedral chlorites. It is most widespread in clayey bauxites and bauxitic clays, in which it accompanies kaolinite. Finely dispersed in the bauxite matrix, it probably is of syngenetic or early diagenetic origin. No segregations of secondary donbassite are known from any bauxite deposit.

It was first described by Caillère and Pobeguin (1962b) from the small bauxite deposits of Saint Paul de Fenouillet in the French Pyrenees, by the name dioctahedral aluminium chlorite. The bauxitic clays there contain 40 to 60% donbassite: this is the highest content found so far in any bauxite. According to Ahmad and Jones (1969), in the Caribbean there is 33 to 55% dioctahedral chlorite in the clayey bauxites and bauxitic clays of the Bahamas and 40 to 50% in the bauxitic clays of the Cayman Islands.

The writer has found a few percent of dioctahedral chlorite in some bauxite samples from Spinazzola, Italy, from the Yugoslav deposits along the Savinja river, from the Senonian bauxites in the Grmeć Mountains, and from the Vrsar bauxite deposits on the Istrian Peninsula, Yugoslavia. Combes (1969) identified the mineral in the bauxites of the Ariège in abundances of 2 to 16%. Keller (1964) found some of it in the diasporic clayey bauxite of Missouri State, US.

Donbassite is ubiquitous in the Eplény deposits, Hungary. Bárdossy (1968a) found it to be hardly less abundant there than kaolinite: it even attains 20% in the upper, clayey part of the deposits.

Cookeite, $Al_2(OH)_2AlSi_3O_{10} \cdot LiAl_2(OH)_6$. It was identified by Vrublevskaya *et al* (1975) in the Jalair deposit of the Turkestanian Ranges, where it accompanies pyrophyllite. According to detailed X-ray investigations by those authors, it is a single-

layer triclinic polytype of remarkably high crystallinity. They ascribe it to a reaction between lithian solutions and pyrophyllite during the metamorphism of the bauxites there. Goffé (1977) found some cookeite in the weakly metamorphic bauxite of the Vanoise, France, where some of it fills out fine fissures, and the rest forms microscopic crystals in the bauxite matrix, together with chloritoid, diaspore, pyrophyllite and haematite. It probably is a product of bauxite metamorphism, of processes similar to those at Jalair.

Trioctahedral subgroup:

Orthochlorites. Orthochlorites are fairly rare in bauxites. Gedeon (1956) assumed most of the magnesium present in Hungarian bauxites to be in the form of chlorite, mainly of *clinochlore*. Chlorite in quantities so minute being below the threshold of X-ray and thermal identification, it has been impossible so far to prove Gedeon's conjecture. Only in some bauxite types of the Nagyharsány deposit group could the writer identify 1 to 4% orthochlorite, and even 8% in a sample of pale greenish-brown clayey bauxite. Its small quantity precluded a more detailed identification, but its X-ray parameters stood closest to those of *prochlorite* (Bárdossy 1961a).

The writer has identified 10% *diabantite* in grey bauxite from Unterlaussa, Austria. The bauxite is boehmitic; its principal clay mineral is kaolinite. In some greenish-grey clayey bauxite and bauxitic clay samples from the Vrace deposit, Lika County, Croatia, Yugoslavia, the writer has found 10 to 20% *clinochlore*, accompanied by 20 to 40% chamosite and some kaolinite in some of the samples. The orthoclorite minerals, finely dispersed in the matrix of this boehmito–diasporic bauxite, are presumably of syngenetic or early diagenetic origin.

Greenish macropisoids in diasporic bauxite samples from the Kandahar deposit group, Afghanistan, contain 30 to 50% *prochlorite* of diagenetic origin. In red diasporic bauxite on Amorgos Island, Greece, dark green veins ranging in width from one millimetre to one centimetre could be observed, which the writer has proved to consist of *clinochlore*. Its epigene origin is beyond a doubt there.

Bauxite, then, contains both syn- and diagenetic orthochlorites. Epigene orthochlorite, however, is very rare. All the orthochlorites known so far have been found in diasporic and boehmitic bauxites; none have been found in gibbsitic ones.

In addition to the orthochlorites formed in the bauxite, clastic orthochlorite grains have also been found in several deposit groups. Their abundance, however, is only in the thousandth- to tenth-of-a-percent range. In Hungary, a few chlorite particles identified as *pennine* were found by Kiss (1955) at Gánt; dark green chlorite grains of clastic origin, of 150 to 300 μm in size were found by Vörös (1958) in the József ore body at Iszkaszentgyörgy, and derived from metamorphic rocks. Vendel *et al* (1971) found a few very scarce chlorite grains in bauxite samples from Nyirád–Dűlt-Nyires and Kislőd.

Leptochlorites

Usually much more abundant than orthochlorites in bauxite, they have nevertheless been found almost exclusively in reduced, green, grey and dark grey bauxites, in which they often attain abundances of 20 to 40%, occasionally even up to 60%. In the USSR, they are especially abundant in the Devonian deposits of the Northern and Central Urals, in the Devonian bauxite of Berdsk and in the Cambrian bauxite of Boksonsk.

According to Beneslavsky (1961, 2nd ed. 1974, 1963), they are finely dispersed in the matrix, but are also present in the ooids and pisoids. He observed some instances pseudomorphs of leptochlorite after pyrite and siderite. In the Northern Urals deposits, kaolinite is gradually replaced by chlorite down the dip. At any depth greater than 500 m, chlorite is the only silicate mineral. The reason for this chloritification has not as yet been elucidated. Sharova and Rudnova (1964) state *thuringite* to be the principal leptochlorite in the deposits of the Northern Urals.

Thuringite could be identified also in some iron-rich bauxites from Afghanistan: indeed, it constitutes the bulk of some of the samples. The writer has found 5 to 10% thuringite accompanied by 20 to 30% kaolinite in some samples from the Barremian bauxite horizon in the Grmeć Mountains, Yugoslavia. In the Lescalé open pit, Ariège, France, in an olive-green clayey bauxite at the top of the deposit, the writer has identified 30% thuringite besides a smaller percentage of kaolinite. Greenish-greyish clayey bauxite from the Velika Planina deposit group in Slovenia, Yugoslavia, contains 40 to 50% *daphnite* according to the writer's examination. It is accompanied by a little kaolinite in some of the samples. The bauxite is diasporic. Daphnite and bavalite were also identified in the Devonian bauxites of the Northern Urals, USSR (Beneslavsky 1961, 2nd ed. 1974, 1963).

According to Papiu *et al* (1970b), the abundance of leptochlorite in the Pădurea Craiului bauxites, Rumania, is greatest (3 to 17%) in green and black bauxite, less (3 to 11%) in partly reduced brownish bauxite and least (0 to 3%) in oxidised, red bauxite. A few percent of leptochlorite was found also in the Haţeg bauxites, Rumania (Papiu *et al* 1971). Its abundance was found to be 10 to 20% in reduced grey bauxite from the Sohodol-Cîmpeni deposit group, Rumania (Papiu *et al* 1970a).

In all these bauxites, the leptochlorites are of syn- and diagenetic origin. Similarly to the chamosites and orthochlorites, they are confined to the diasporic and boehmitic bauxite types and are absent from the gibbsitic ones. No leptochlorite grains of clastic origin have so far been found in any bauxite.

Expanding chlorites

The group contains those chlorite minerals whose basal reflection at 14 Å expands to 16 to 18 Å on treatment with glycerine. Caillère and Pobeguin (1962b) were the first to identify this mineral by the name 'pseudochlorite', in quantities of 3 to 5% in red bauxites from Bédarieux, 4% in a pink one from Cazouls-les-Béziers and 6% in a pale yellow one from Saint Chinian, France. All the samples were boehmitic; in them, expanding chlorite is accompanied by more or less kaolinite. The mineralogical identifications are confirmed by a MgO content of 0·3 to 1·3% in the samples. Remarkably enough, all the bauxites in question were of the oxidised type.

(E) Group of micas and hydromicas

Dioctahedral subgroup:

Muscovite, $KAl_2(OH)_2AlSi_3O_{10}$. It is found in bauxite in the form of clastic grains. In the Hungarian deposit groups, it was identified in quantities from a few thousandths to a hundredth of a percent by Kiss (1955) at Gánt, by Vörös (1958) at Iszkaszentgyörgy, by Vörös and Gecse (1976) at Nagyegyháza, and by Bárdossy

(1968a) at Eplény. According to Vendel *et al* (1971), it is present also in the bauxites of Nyirád, Szőc, Halimba and Kislőd.

Sohodol bauxite, Rumania, contains more than one percent muscovite according to Papiu *et al* (1970a). The muscovite grains are 200 to 400 μm in length; most of them are bent (fig. 139). The bauxitic clays of the Haţeg deposit, Rumania, contain according to

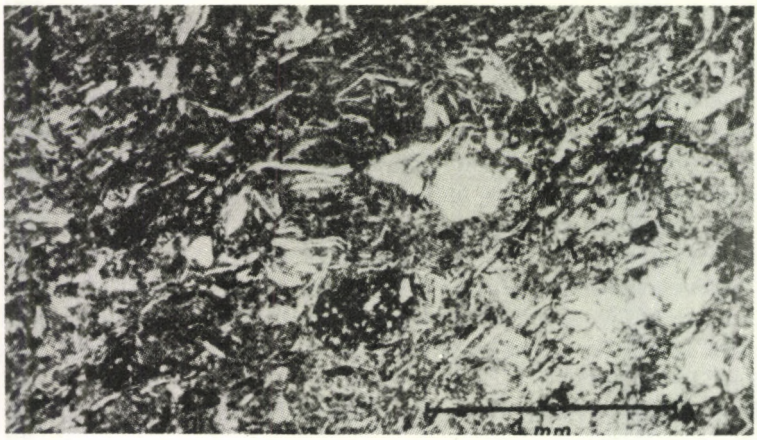


Figure 139. Muscovite grains of clastic origin in a bauxite. Sohodol-Cîmpeni, Rumania. After Papiu *et al* (1970). Thin section, parallel nicols

Papiu *et al* (1971) 2 to 4% muscovite as a rule, and 9% as a maximum in a few samples. The sheaf-like muscovite grains of 200 to 400 μm length are usually kaolinised at both ends.

In most of the bauxites in the Mediterranean zone, muscovite contents are less than a tenth of a percent. The finding of Roch and Deicha (1966) that clayey bauxite and bauxitic clay at the bottom of the Brauch open pit near Brignoles, France, contains 1 to 10% muscovite, is quite exceptional. The mineral is derived by those authors from a nearby crystalline basement.

Jamaican bauxite very seldom contains even a few muscovite grains (Hill 1955), and so do the other Caribbean bauxites. It is the Chinese bauxites that are richest in muscovite. According to Schüller (1957), its abundance is 5 to 20% in the Gun District bauxites, where it is found in the bauxite pebbles, ooids and pisoids as well as in the matrix. Some of the grains are kaolinised. There is as much muscovite in the other Chinese deposits (Beneslavsky 1963). The Cambrian bauxite of Boksonsk, USSR, also contains several percent muscovite (Orlova 1958).

In several bauxite samples from Amorgos Island in the Aegean, the writer has detected by X-ray measurements about 10% muscovite accompanied by about as much paragonite. On nearby Samos Island, regional metamorphism of the bauxite gave rise to 15 to 20% muscovite associated with somewhat less chloritoid. From the original bauxite minerals, diaspore, kaolinite and haematite, however, substantial percentages remain (de Lapparent 1937). The situation is much the same in the partly metamorphosed bauxite deposits and emeries in Western Asia Minor, which according to Önay (1949) often contain up to 30% muscovite, usually in association with

margarite. Önay observed some pseudomorphs of muscovite after diaspore in some of the fissures in the bauxite. In the weakly metamorphic bauxite of the Vanoise, France, Goffé and Saliot (1977) proved muscovite to be present largely in the form of fine fissure fillings, usually in the company of paragonite. Those authors ascribe it to the influence of alkali-rich solutions.

The famous Naxos Island emery, the most metamorphic of all, also contains muscovite. It is most enriched in the hydrothermal veins, in the company of corundum, margarite, magnetite and tourmaline (Papastamatiou 1951). A muscovite produced by a hydrothermal influence was mentioned by Orlova (1958) from the Boksonsk deposits, USSR. The thin hydrothermal veins there are filled out with muscovite accompanied by chlorite, pyrite and calcite. Orlova observed traces of metasomatism along the veins.

Paragonite, $NaAl_2(OH)_2AlSi_3O_{10}$. It is known from two deposit groups so far. In Amorgos Island bauxite, mentioned in connection with muscovite, the writer has detected 5 to 20% paragonite by X-ray analysis of several samples. Goffé and Saliot (1977) demonstrated it in the bauxite of the Vanoise, France, where, in association with muscovite it tends to fill some fine fissures traversing the bauxite. In those authors' opinion, both minerals are due to the influence of alkali-rich solutions acting in the course of the bauxite's metamorphism.

Hydromuscovite, illite, $(K, H_2O)Al_2[(H_2O, OH)_2AlSi_3O_{10}]$. Identified so far in 32 deposit groups, it is most often confined to small parts of the deposits and to just a few bauxite types.

In the Hungarian deposit groups, the writer has found in the cores of Borehole P-1, Alsópere, 25% illite in the half-metre of bauxitic clay immediately above the bedrock, and 3% in they clayey bauxite above it. None of the samples of the bauxite ore body overlying this section contain any illite. In the deposits of Halimba–Malomvölgy, 4 to 10% illite has been found by the writer, exclusively in the 5 to 50 cm thick bauxitic clay overlying the bedrock, and identical amounts in the deposits at Kislőd, in bauxitic clay at the bottom of the deposits.

In their investigation of a lignitic bauxite profile at Gánt–Bagolyhegy, Kiss and Vörös (1965) found a few percent of illite in three of their samples. The present writer's X-ray examinations gave 5 to 10% illite for clayey bauxite samples from near the bedrock of the Újfeltárás deposit, Gánt. Vendel *et al* (1971) demonstrated small amounts of illite in bauxitic clay from the top of the deposit in their mineralogical investigation of a profile of Lens Malomvölgy X, Halimba.

Illite, then, is confined to the clayey parts of the deposits, most often to those at the bottom, more seldom than to those at the top, whereas in the high grade bauxite constituting the central orebody there are just a few scattered illite grains.

The small deposits at Dréveneuse in Switzerland are remarkable for a high, 0·3 to 2·2% K_2O content (Badoux and de Weisse 1959). In a sample from this deposit, received courtesy of de Weisse, the present writer has found 20% illite, and has inferred therefrom the probable presence of substantial amounts of illite also in the other high-potassium samples.

In all other Mediterranean deposit groups, illite has been found exclusively in the lowermost, clayey portions of the deposits, e.g. by Périnet (1964) in the Peygros deposit

near Brignoles, by Caillère and Lahodny-Šarc (1966) in the nearby Mine de Maron, by Cocco and Pecorini (1959) in the Nurra deposits of Sardinia and by Combes (1969) in the deposits near Fuentespalda in Spain, and in claystones underlying the Ariège deposits. In the latter, its abundance was found to be 20 to 40% as a rule and up to 66% exceptionally. Šinkovec (1970) found some illite at the bottom of the Lika deposits, Yugoslavia, whereas the writer in the course of his mineralogical studies has observed it in a similar position at the bottom of the Spinazzola deposits in Italy, in abundances of 12 to 14%.

The writer has also found illite in the central and upper parts of some Yugoslav bauxite deposits, notably 2 to 10% in the Malm deposits of the Hrušica Mountains in Slovenia, in green bauxite of the deposits near Kozina in Istria, and in the small deposits of the Savinja valley in Slovenia. He has further observed 1 to 8% illite in the central part of the Campo Felice deposit, Italy, and 7 to 13% in its top part; in fact, in a calcareous–clayey bauxite next to the cover boundary, the mineral was enriched to 22%. In Greece, in samples of green clayey bauxite from the top of the Malm deposits near Itea, Bárdossy and Mack (1967) assayed 22% illite.

Illite is present in most Iranian bauxite deposits in abundances of 1 to 20%, but it is confined to certain parts of the deposits. It has been demonstrated by the writer to exist in some of the Permian bauxites of Vietnam. According to Sharova and Glazunova (1966), some bauxitic clays from Vietnam contain 1 to 17% hydromuscovite accompanied by pyrophyllite. They attribute the origin of both minerals to an exposure to weak metamorphism. Illite is widespread in abundances of 5 to 10% in all the Chinese deposits, in which it is associated with muscovite and sericite. According to Beneslavsky (1963), it is present in the USSR in the bottom part of the Krasnooktyabrsk deposit. The deposits of the Turgay Plateau and of Central Kazakhstan are embedded, as a rule, in clays in which hydromicas grow increasingly abundant with increasing distance from the ore bodies. The situation was found to be the same by Gulyanitsky (1969) in the south of Timan Mountains.

In the Caribbean deposit groups, it was only in some Jamaican deposits, in clayey bauxite next to the bedrock, that Strahl (1971) could identify a little illite.

This mineral, enriched next to the bedrock, is held by several authors to be an alteration residue of the carbonate bedrock. However true that may be as a part-explanation, the bulk of illite more probably derives, together with muscovite and sericite, from the alumosilicate parent rocks of the bauxite, as weathering products of the potash feldspars and micas originally contained in those. This is especially obvious in the Chinese bauxites. The writer holds the bulk of illite in the central and upper parts of the bauxite deposits to be also allothigenic; the small flaky illite grains were readily entrained by the surface waters. Subsequent illitisation was presumably confined to those places where part of the bauxite was deposited in the sea, or a marine transgression took place immediately after its deposition. It was in both cases under the influence of the alkalies dissolved in seawater. The Itea, Dréveneuse and Hrušica bauxites referred to above may be cases in point.

Glauconite, $(K, Na, Ca)_{<1}(Al, Fe^{2+}, Fe^{3+}, Mg)_2[(OH)_2Al_{0.35}Si_{3.65}O_{10}]$. Antal (1973) found a very few glauconite grains of clastic origin in the Iszkaszentgyörgy–Rákhegy ore body, Hungary. According to Beneslavsky (1963), an olive green

glauconitic clay is found in places in the lower parts of the Krasnooktyabrsk deposit group, USSR.

A green clay 1 to 10 cm thick was found by Bárdossy *et al* (1977) over an area of a few square metres at the bedrock boundary in one of the bauxite lenses near Campo Felice, Abruzzi, Italy. Upward, it graded into a red bauxitic clay. The green clay contains 60 to 80% glauconite, accompanied by 15 to 20% kaolinite and 5 to 10% montmorillonite. The clay is impregnated with calcite in places. Its mineral composition and texture, imply the green clay to be a strongly weathered derivative of a tuff fallen onto the limestone surface. It was probably the bauxitisation of this tuff that produced the parent material for some of the deposits of this group.

Trioctahedral subgroup:

Biotite, $K(Mg, Fe)_3[(OH)_2AlSi_3O_{10}]$. A rare clastic component in bauxite, its abundance is kept below a hundredth of a percent by its proneness to weathering. Among the Hungarian deposit groups, a few partly faded biotite grains were identified by Kiss (1955) at Gánt, by Vörös (1958) and Antal (1973) at Iszkaszentgyörgy, and by Vörös and Gecse (1976) at Nagyegyháza. Demangeon (1965) states it to be very rare in the French bauxites; it was only at Saint Chinian and Bédarieux that a few grains could be identified. Papiu *et al* (1971) found some very scarce, strongly weathered, partly kaolinitised or chloritised biotite grains in Haţeg bauxite, Rumania. Karanov (1965) identified the mineral in the deposits near Tran in Bulgaria.

De Lapparent (1937) found a little biotite in the partly metamorphosed bauxite of Samos Island on the Aegean. It presumably is of metamorphic origin there. True, no biotite has so far been found in the much more strongly metamorphic emeries of Asia Minor and of Naxos Island, Aegean, but this may be due to difference in composition, temperature, and pressure.

Pyrophyllite-talc subgroup:

Pyrophyllite, $Al_2[(OH)_2Si_4O_{10}]$. A comparatively rare mineral in karst bauxites, it was first identified by Tučan (1934) in a Triassic bauxite near Mazin in Croatia, Yugoslavia. According to Šinkovec's (1970) microscopic observations, it forms platy grains of 5 to 200 μm size there. The present writer has found pyrophyllite to be the principal silicate mineral of Mazin bauxite, attaining abundances of 20 to 50% in the bauxite, proper and 50 to 90% in the bauxitic clay. The red bauxite is traversed in many places by snow-white veins of several millimetres width, filled out by pyrophyllite and a few percent of kaolinite. The bauxite itself is diasporic. The Mazin deposit group is situated in a zone of strong structural deformation, where several overthrusts of substantial size took place during the Neogene. Pyrophyllite was presumably produced by structural stress out of the original kaolinite content of the bauxite. For this to be the case, some silica must have been added, most likely in ground water. In any case, the writer has observed no trace of thermal metamorphism either in the bauxite or in the rocks surrounding it.

Pyrophyllite was identified by Beneslavksy (1963) in the bauxites of Boksonsk, USSR and Gun-syan, China. In the latter, the mineral is enriched especially in the matrix, where it is accompanied by diaspore. Beneslavsky in places found some aggregates of its bent flakes. He discovered pyrophyllite more seldom also in the

pisoids, and in a few places even as pore and fissure fillings, or as coatings on surfaces of parting. On the evidence of all this, Beneslavsky assumed a multiphase pyrophyllitisation, without giving any details of how it could have taken place.

Robert (1971) identified pyrophyllite in several samples from the small deposits near Kymi on Euboea Island, Greece where it is the only silicate mineral in parts of the bauxite; it is accompanied by kaolinite elsewhere. In one sample of red diasporic bauxite, he found almost 40% pyrophyllite. In two bauxite samples from nearby Skopelos Island, the present writer also has found some pyrophyllite. The bauxite is diasporic and also contains some chloritoid.It had presumably been exposed to a weak metamorphism, of which pyrophyllite and chloritoid are the products. Goffé and Saliot (1977) identified pyrophyllite in the weakly metamorphic bauxites of the Vanoise, France.

The bauxite near Djulfa in the Little Caucasus, USSR, also contains some pyrophyllite (Mikhailov *et al* 1973). More than 40% pyrophyllite accompanied by 10% kaolinite has been observed by the present writer in a red bauxite sample from the Jalair deposit in the Turkestanian Mountains, USSR. The bauxite is diasporic and also contains a few percent of corundum, the product of weak metamorphism. Vrublevskaya *et al* (1975) state the pyrophyllite to form mainly macro- and microcrack fillings in the bauxite of that deposit. It has crystallised in the lattice variety 2M. A few pyrophyllite flakes attain up to 4 mm length. Some pyrophyllite has been found in some bauxitic clays of the Jajarm deposit group, Iran. Sharova and Glazunova (1966) identified 1 to 27% pyrophyllite filling hair cracks inside pisoids in Northern Vietnamese bauxites.

In the majority of the deposit groups enumerated, the geological environment as well as the great hardness and the mineralogical composition of the bauxite imply an incipient metamorphism, and an especially strong structural stress. This was probably the prime cause of pyrophyllite of formation. It should not be forgotten, however, that pyrophyllite may also form under normal surface temperature and pressure, as proved by its presence in a laterite profile at Kassongoni, Guinea (Balkay and Bárdossy 1967). The writer deems it probable for multi-generation pyrophyllite in the Chinese deposits and pyrophyllite on Euboea Island to have formed under surface temperature and pressure out of the kaolinite in the bauxite, as a result of silica intake from ground water.

Subgroup of brittle micas:

Margarite, $CaAl_2[(OH)_2Al_2Si_2O_{10}]$. A product of bauxite metamorphism, it was identified by de Lapparent in some samples of Samos Island bauxite, Greece (1937). It is widespread in the emeries of Western Turkey, where its abundance, 2 to 5% on average, locally attains up to 30 and 40% (Önay 1949). It is present in Naxos Island emery also, especially in cracks of the rock (Papastamatiou 1951).

In the USSR, Beneslavsky (1963) stated the metamorphic bauxites of Salair to contain a magnesian variety of margarite. Karapetov (1968) found some margarite accompanied by corundum and mica in some strongly metamorphic bauxites in the Pamir Ranges, USSR. The present writer assayed 10 to 20% margarite accompanied by diaspore and kaolinite in samples of a red, partly metamorphic Carboniferous bauxite from the Aktau Mountains, Central Asia, USSR.

(F) The smectite group

Montmorillonite, $(Ca, Na)_{0.7}(Al_{3.3}MgO_{0.7})Si_8O_{20}(OH)_4 \cdot nH_2O$. Extremely rare in karst bauxites, it was identified in Hungary by Nemecz (1953) in one bauxite sample from Iszkaszentgyörgy–Kincses. Kiss and Vörös (1965) found 1 to 6% montmorillonite in a lignitic bauxite profile in the Bagolyhegy deposit of Gánt, where it was most abundant near the bedrock. Vendel *et al* (1971) observed some montmorillonite in one profile of the Harasztos open pit, Gánt, especially in the multicoloured and grey bauxitic clays in the proximity of the cover. It has not been identified so far in any other Hungarian deposit group.

Ahmad *et al* (1966) examining two bauxite profiles on Jamaica found in one of them about one percent montmorillonite immediately above the bedrock limestone. The demonstration of so small a quantity of montmorillonite was made possible by an alkali-leaching process which concentrated the clay minerals in the bauxite. In the opinion of those authors, montmorillonite formation was promoted by local concentrations of alkalies made possible by the poor permability of the underlying limestone bedrock. This is probably a highly exceptional case even in a Jamaican relation, as no other author has mentioned any montmorillonite from any other bauxite of the island.

In sample from a light grey nest of soapy touch, found in a bauxite in the lower part of the Krasnaya Shapochka deposit, Northern Urals, USSR, and collected by Bushinsky, the present writer has found some 70% montmorillonite, which seems epigene on the evidence of its mode of occurrence. A glauconitic green clay at the base of the Campo Felice deposit, Italy, also contains 4 to 5% montmorillonite.

The processes of bauxitisation generally result in alkali removal. This is why montmorillonite must be so rare—even exceptional—in bauxite. It may be an alteration residue of the parent rock, or a product of subsequent processes of resilification involving local rises in alkali concentration.

(G) The allophane group

This is the group encompassing the X-ray amorphous hydrosilicates of aluminium. In earlier literature, its presence used to be assumed in many deposit groups. Systematic X-ray analyses, however, revealed its being confined to the youngest karst bauxites, where it is found in abundances of a few percent, in the company of likewise amorphous hydroxides of aluminium and of iron, as established for the Caribbean by Ahmad and Jones (1969) studying samples of clayey bauxite from the Cayman Islands and the Bahamas, and by the present writer studying samples from the Loyalty Group and from Niué Island in the East Tonga Group.

The only place where allophane is found in significant quantities is in the small bauxite deposits of Tulsk–Ryazansk, USSR (Trubina 1958), where it is finely dispersed in the bauxite matrix, together with halloysite in places and halloysite and kaolinite elsewhere. It occasionally forms hard concretions in a bauxite which is highly friable otherwise. According to Trubina, allophane was the first-formed mineral; part of it was subsequently converted into halloysite and then kaolinite.

In the bauxite deposit Vonji Do of Crna Gora (Montenegro), Yugoslavia, a nest of 5 to 6 cm size was found at the boundary of a white bauxite and its cover, which in

addition to white gibbsite and nordstrandite contained some translucent vitreous amorphous material, identified by Tertian (1966) as allophane. This identification was confirmed by laboratory tests of Lahodny-Šarc, Mihaličić-Gostiša and Karšulin (1969), who stated the formula of allophane in this deposit to be $Al_2O_3 \cdot SiO_2 \cdot 7H_2O$. Using infrared absorption methods, those authors proved the lattice ordering of the mineral to tend towards that of kaolinite. On the strength of some nearby segregations of basaluminite and gypsum, they presumed aluminium and silicon to have been leached out of the bauxite by sulphuric acid resulting from pyrite oxidation, and the allophane to be precipitated out of the solution thus formed (that is, to be the product of an epigene process).

5.3.2.6. TECTOSILICATES

Quartz, SiO_2. Found in the form of clastic grains in bauxite, its abundance in the Hungarian deposit groups is most often just a few hundredths of a percent. The bauxites of Nézsa, however, contain according to Kiss (1952) 0·3 to 1·2% quartz, 0·5% on average. It was found also in the bauxites of Gánt (Kiss 1955), Iszkaszentgyörgy (Vörös 1958), Eplény (Bárdossy 1968a), Halimba and Nyirád (Vendel *et al* 1971). Its grain size ranges from 10 to a few hundred micrometres. In the reworked deposits, e.g. at Nagyvázsony, Zalahaláp, Isztimér, Óbarok and Újbarok, quartz is enriched to 1 to 10%.

In the Mediterranean-type deposits, the abundance of quartz grains is a few hundredths of a percent as a rule, seldom more than a tenth of a percent. The only exception is Puerto Rican bauxite, which according to Hildebrand (1960) contains 20 to 30% quartz on average, in worn grains smaller than 0·5 mm. They were probably introduced into the bauxite from the volcanics, sandstones and granodiorites lying in the southern foreland of the bauxite region. In view of their high quartz content, the present writer feels those bauxites to be possibly reworked ones, rather strongly contaminated with quartz in the process.

The Kazakhstan-type deposits contain somewhat more quartz than those of Mediterranean type. Average contents of 0·1 to 1·0% have been found in most of the deposits of Central Kazakhstan and of the Turgay Plateau, USSR, and in the small deposits of the Coastal Plain and of the Valley and Ridge Province, US. There is 1 to 3% quartz in the bauxites at Chadobetsk and along the Yenisei River (USSR). Slukin (1971) states the quartz grains to be often strongly coroded and partly replaced by gibbsite. This replacement reaction took place in the deposit already, in the course of bauxitisation. There are quartz grains within some of the pisoids as well as in the matrix.

Quartz is most abundant in the secondarily reworked—Salento type—deposits. There is, e.g. 3 to 7% quartz on average according to Trubelja and Vasiljević (1971) in the secondary deposits at Baraći near Jajce, Yugoslavia. In the other, non-redeposited, Senonian-covered deposits at Jajce, which are only a few kilometres away, there is just a few hundredths of a percent of quartz.

There is also some very rare epigene quartz in bauxite: Beneslavsky (1961, 2nd ed. 1974) found cryptocrystalline segregations of quartz in some of the bauxites along the Yenisei River, USSR. According to Jurković (1965), the Suvaja–Šolaja deposits,

Yugoslavia, contain some very sporadic epigene quartz grains and quartz-filled microcracks. Caillère and Pobeguin (1965) found inward-projecting bipyramidal crystal-clear or milky crystals of quartz accompanied by calcite on the inner walls of small cavities in bauxite from the Peygros mine near Brignoles, France. Those authors found no quartz grains of clastic origin in this bauxite. They attribute both the quartz and the calcite to epigene processes, to a process of segregation out of ground water seeping through the bauxite.

Cristobalite, SiO_2. An extremely rare clastic mineral in bauxite, it was identified by Ščavnićar *et al* (1968) in abundances of a hundredth of a percent in some bauxite samples from the Čitluk and Nevesinja deposits near Mostar, Yugoslavia.

Volcanic glass. Hartman (1955) in the course of his micromineragolical investigation of Jamaican bauxite found a few particles of volcanic glass. He holds them to be alteration residues from the underlying limestone, in which he has found similar grains of volcanic glass.

Feldspars. Exclusively clastic, they are found in abundances from a few hundredths to a tenth of a percent. It was only in Puerto Rican bauxite that Hildebrand (1960) found *oligoclase* and *sanidine* grains in quantities of a percent or so.

Of the Hungarian deposit groups, a few strongly weathered grains of feldspar were found by Antal (1973) in the Rákhegy ore body and a few scattered grains of *microcline* by Vörös and Gecse (1976) in the Bitó deposit, both at Iszkaszentgyörgy.

A few grains of *bytownite*, *labrador* and *anorthite* and one grain of *microcline* were identified by Bárdossy (1961a) in the 0·1 to 2·0 mm grain size fractions of samples from the Dreistätten bauxite deposit, Austria. Karanov (1965) mentions a few grains of feldspar without any closer identification from the deposits near Tran in Bulgaria.

Hartman (1955) and Strahl (1971) found a few grains of basic plagioclase, partly altered to sericite, in Jamaican bauxite. There were a few scarce grains of *albite* and *microcline* also. In the USSR, there are a few feldspar grains, partly altered into hydromicas, in the Devonian bauxite of the Northern Urals according to Beneslavsky (1961, 2nd ed. 1974) and also in the Ordovician bauxite of Tselinograd according to Dolgopolov *et al* (1971). Weathered feldspar grains were described by Krivtsov (1968, 1969) from the Western Ubagansk, and by Beneslavsky (1963) from the nearby Krasnooktyabrsk deposits. The latter author observed also some pseudomorphs of gibbsite after feldspar at Krasnooktyabrsk.

Zeolites. Very rare in karst bauxite. In the USSR, *chabasite*, *phillipsite* and *ferrierite* in the company of calcite were identified by Beneslavsky (1961, 2nd ed. 1974, 1963) in the bauxites of the Northern Urals. In the Sosvinsk deposits of that region, the abundance of chabasite locally attains 4·5%. In the bauxites of the Central Urals, Beneslavsky *(loc cit)* found a little *heulandite* and *harmotome* forming pore and fissure fillings and thin coatings. They are of secondary origin.

Ščavnićar *et al* (1968) found about one percent *natrolite* in one bauxite sample from the Listica deposit near Mostar, Yugoslavia. According to Misirlić and Mitrović, (1969), bauxite in the deposit group of the Grebnićka Planina everywhere contains 0·5 to 1·5% zeolite.

5.3.3. CARBONATES

Calcite, $CaCO_3$. It is found in quantities ranging from a tenth of a percent to several percent in most deposit groups. Calcite syngenetic with bauxite deposition is found only where the deposit grades with continuous sedimentation into an overlying limestone, as e.g. in the Devonian deposit groups of the Urals and of the Altay–Sayan ranges, USSR. The abundance of calcite gradually increases in the zone of transition which ends in a pure marine limestone. Such calcite is very fine-grained, and its forms of appearance agree with those of the calcite making up the limestone of the immediate cover.

Most of the calcite in bauxite is of diagenetic and epigene origin, formed out of ground water seeping down from a carbonate cover. There is a remarkably high concentration of calcite in the top parts of the Alsópere deposits of Hungary, as revealed e.g. by the profile of Borehole P-1 (table 27). Calcite contents in the top one metre of the deposit attain 16%. Microscopic studies revealed that the calcite had precipitated in the bauxite matrix, avoiding the more compact roundgrains and ooids.

In the middle and lower parts of the deposits, calcite is sporadic, most often taking the form of nests and cavity-fillings. In the deposits of Halimba and Szőc, the writer has in places observed cavity fillings of several centimetres in size. The cavities are not as a rule filled out completely with calcite; it is only their walls that are studded with automorphic crystals of a few millimetres in size.

A peculiar form of calcitisation has been observed by the writer in a freshly exposed bauxite profile in an open pit working at the Kislőd deposit. In the top part of the deposit, a hard bauxite of conglomerate texture was traversed by numerous fissures and cracks, in a thickness of 0·1 to 1·5 cm. They are filled with snow-white, friable calcite of spongy appearance. There are also numerous bauxite pebbles wearing a porous snow-white calcite coating a few millimetres thick. The thickness of the calcite segregations rapidly decreases downward, those disappear altogether at a depth of two metres below the cover. These friable segregations of calcite are obviously of late epigene or supergene origin, having precipitated out of ground water seeping down from the overlying limestone.

Calcite may be substantially enriched in the vicinity of tensional faults traversing the deposits. These facilitate ground water seepage and permit calcite to segregate in the cracks along the faults. Epigene calcite crystals are also frequent at the boundary of bauxite and bedrock. They attain sizes of a few centimetres in places, as observed by the writer in some of the deposits of Kislőd and Halimba, Hungary.

Calcite content is highest in the small deposits near La Boissière in France (10 to 20% average). It has impregnated the matrix only, avoiding almost entirely the pebbles of bauxite and the ooids contained in it. The calcite content of the Alpilles deposit group average 5 – 14% in the form of pore fillings.

Elsewhere, calcitic impregnations are restricted to the top section of the deposits, as observed by the writer e.g. in the San Giovanni Rotondo deposit group of Italy. The impregnation is also confined to the matrix in this case. Its average abundance nevertheless ranges from 5 to 25%. In the upper parts of the deposits called Präfing and Schwarza of the Unterlaussa deposit group, Austria, there is according to Haberfelner (1951) a calcite-impregnated bauxite containing 3 to 30% calcite in a thickness of one to

two metres. The impregnation there is also restricted to the matrix. Haberfelner found less than 0·1% calcite in the ooids picked out of the bauxite.

In a greenish hard bauxite from one of the deposits of the Pădurea Craiului group, Rumania, the writer has observed some translucent single crystals of calcite of 1 to 6 mm size. Large shiny crystal faces are conspicuous on the fracture surfaces of the rock. These too presumably are epigene pore fillings. According to Combes (1969), each of the secondarily leached-out pisoids in the bauxite of the Lescalé deposit in the Ariège, France, is replaced by a large single crystal of calcite.

Calcite-filled fissures are especially frequent in deposits of strong structural deformation as observed by the writer e.g. in the upper parts of the deposits in the Eleusis-Mandra group, Greece, in which the fissures attain a width one centimetre in places. According to Beneslavsky (1963), calcite-filled fissures are also very frequent in the Devonian deposits of the Northern Urals, USSR; they are 1 to 10 cm wide there. He also observed in thin sections some fine acicular crystals of calcite in the bauxite matrix. In the Krasnooktyabrsk deposit, he found calcite-filled cavities in which the calcite formed concentric shells. Some of the large calcite crystals there wore thin films of pyrite. There are some very rare pseudomorphs of calcite after siderite in the bauxites of the Urals (Beneslavsky 1963).

Dolomite, $CaMg(CO_3)_2$. Dolostone pebbles and debris are known from numerous bauxite deposits; the mineral dolomite, however, is very rare. It was demonstrated e.g. by Beneslavsky (1963) in very small amounts from the Devonian bauxites of the Northern Urals, USSR.

In recent years, a concentration of dolomite that may well be called unique has been recognised in the large bauxite deposit filling out the Halimba Basin in Hungary. That deposit contains abundant scattered dolostone pebbles and layers of dolostone gravel (cf figs 113 and 114). In addition to those, the writer has also observed finely disperse authigenic dolomite grains of 2 to 80 μm size in the top part of the Halimba III ore body, whose abundance monotonically increases upward as a rule; more seldom, it fluctuates rhythmically. This dolomitic bauxite layer is 1 to 3 m thick. The writer, performing quantitative mineralogical analyses on more than 150 dolomitic bauxite samples, found their dolomite contents to vary from 1 to 90%, the rest having the composition of a bauxite, clayey bauxite or bauxitic clay.

Owing to their fine grain and to a finely dispersed ferruginous pigment, the dolomite grains are rather indistinctly visible in thin sections. The writer and Pantó have therefore also examined some of the samples by means of the electron microprobe. The grains turned out to be of extremely irregular, intricate outline, with a modal grain size in the 5 to 20 μm range. Grain morphology was especially conspicuous in the back scattered-electron images (fig. 140.) The X-ray recordings of elemental distributions revealed the dolomite grains to contain inclusions of minerals of aluminium, iron and titanium in the 0·1 μm size range, which they could acquire only by forming in place and incorporating during their growth the finely disperse mineral grains of the bauxite mud. On the strength of these observations, the writer holds the bulk of these dolomite particles to be of a syngenetic or early diagenetic origin.

A slow subsidence of the relief presumably gave rise to a flat-bottomed freshwater lake in the Halimba Basin, just before the end of bauxite accumulation. In the water of

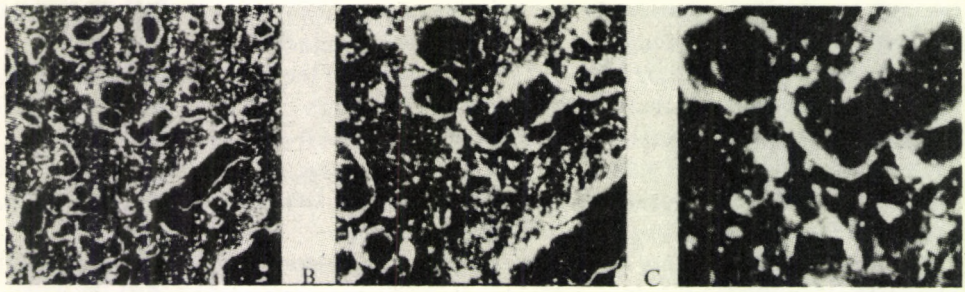


Figure 140. Authigenic dolomite grains in a bauxite matrix. Halimba III mine, Hungary. Electron microprobe, backscattered electron pictures
(A) 300 × ; (B) 600 × ; (C)1200 ×

this lake, dissolved calcium, magnesium and hydrocarbonate were gradually enriched, because there were dolostone outcrops all around on the surface, from which the rains could dissolve those components and carry them into the lake. After they had attained a certain concentration, the precipitation of dolomite could begin in the unconsolidated muddy bauxitic sediment of the lake bottom. The extremely great and very active specific surfaces of the bauxite particles presumably also helped to initiate the process. Dolomite partly grew onto minute clastic dolostone particles, and partly formed independent precipitate grains. In the course of their growth, these grains, as pointed out above, engulfed some minute particles from the bauxitic mud surrounding them.

This is a peculiar, freshwater form of dolomite formation. Investigations with the electron microprobe have not borne out the assumption of a secondary, metasomatic dolomitisation of calcite crystals of primary precipitation. According to Krauskopf (1967), the crystallisation of dolomite at Earth-surface temperature is so slow as to defy adequate reproduction in the laboratory. In the Halimba Basin, there was time enough for this process to take its course.

In a set of samples collected by Robert on Skopelos Island in the Aegean, the writer has indentified a few samples of violet-brown dolomitic bauxite and bauxitic dolostone. Their finely disperse dolomite content has an authigenic look about it. The circumstances of formation of these samples of peculiar composition are not well understood as yet, but so much seems probable that they also are the products of a syngenetic precipitation of dolomite.

Ankerite, $Ca(Mg, Fe)(CO_3)_2$. Very rare in bauxites, it was identified in small quantities by Beneslavsky (1963) in the Devonian bauxites of the Urals, USSR. In a dark brown bauxite sample, likewise from the Urals, the writer has determined 7% ankerite associated with 16% siderite and 1% calcite. The writer has further found 2% ankerite in one sample from the Tavernes deposit of the Haut-Var, France. The grey Permian bauxites of Vietnam often contain a few tenths of a percent of ankerite; indeed, 5 to 10% has been found in some samples by the writer's team. It presumably is of diagenetic or epigene origin.

Kiss (1955) identified few ankerite in a ferruginous crust at the bedrock boundary in the Gánt–Hosszúharasztos deposit, Hungary. It is epigene, as are all the other minerals of that crust.

Siderite, $FeCO_3$. Demonstrated in abundances exceeding 0·1% from 30 deposit groups so far, it is confined in most instances to reduced, green, grey and black bauxite types. It is most often epigene and diagenetic, less often syngenetic.

In Hungary, it was Vadász (1946) who first demonstrated it in a reddish-brown bauxite of the Nagyegyháza deposit. His examination indicated a substantial quantity of siderite, a finding that has been confirmed by the bauxite prospection of recent years: the bauxite drilled up in the Nagyegyháza Basin often contains 1 to 10% siderite, even up to 25% in some samples. Microscopy has shown it to be largely spherulitic, and to form fillings in pores or root remains; it may also fill fine fissure networks in places. It is accompanied by pyrite in grey bauxite (Tóth 1976). Most abundant in the upper parts of the deposits, getting gradually scarcer downward, it is associated with the coal seams above the deposits. Partly diagenetic, partly epigene, it presumably was precipitated in the bauxite out of downward-seeping strongly reducing solutions rich in carbon dioxide.

Green chamositic bauxite in the upper part of Gánt–Újfeltárás deposit is traversed by veins 1 to 5 mm wide filled with pure siderite. The writer's X-ray analyses have proved the siderite to contain a little manganese, which makes it transitional towards *oligonite*. Grey pyritic bauxite at Nyirád and Halimba also contains a few percent siderite in places, and the writer could even identify it in a few sample of the underlying red bauxite, in abundances of 0·1 to 0·5%.

In the USSR, Beneslavsky (1963) found some syngenetic siderite in the Devonian bauxites of the Northern Urals. It is predominantly dispersed in the matrix in very fine crystals, but may also form smallish spherulites. He also found in the same deposits some epigene siderite crystals of 1 to 10 mm, exceptionally even of 10 to 15 cm size. The smaller ones tend to form fine fissure fillings or coatings on the parting or cleavage planes of the bauxite.

Grey bauxites in the Triassic deposits at Volchansk and Bogoslovsk in the Northern Urals contain 15 to 25% siderite on average. Some of it is finely disperse in the matrix; the rest constitutes spherulites. Spherosiderite nodes of 6 to 7 cm size have also been found in places. Some of the pisoids are surrounded by a film of siderite (Tkachenko and Bol'shun 1971).

Several deposit groups of the Turgay Plateau exhibit a strong epigene sideritisation. In certain parts of the Krasnooktyabrsk deposits, almost all the iron minerals are converted into siderite, the abundance of which is 1 to 12% as a rule but up to 21 and 25% in places. Siderite spherulites and sideritic fissure fillings are frequent also there (Kirpal' and Tenyakov 1974).

There is also a little siderite in the Timan Mountains bauxite, mainly in the form of minute spherulites in the lower parts of the deposits (Gulyanitsky 1969). There are also some rather rare automorphic rhombohedral siderite crystals there.

Of all the deposit groups of the Mediterranean bauxite belt, it is in the greenish-grey reduced bauxite types of the Sohodol-Cîmpeni deposit group of Rumania that siderite content is highest (Papiu *et al* 1970a). The mineral tends to form spherulites of a few

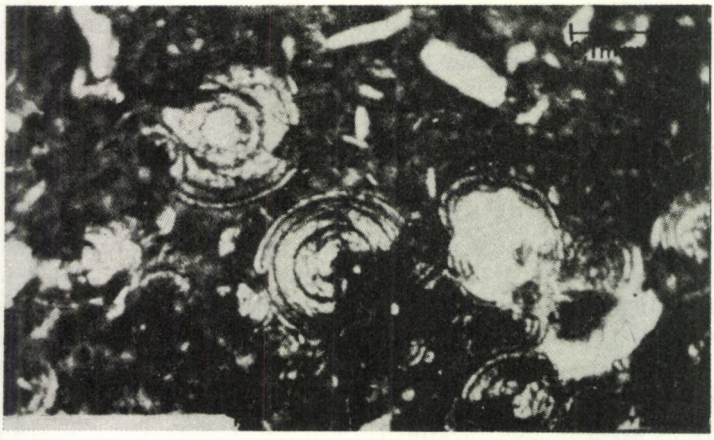


Figure 141. Siderite spherulites in a bauxite matrix. Sohodol-Cîmpeni, Rumania. After Papiu *et al* (1970)
Thin section, parallel nicols

tenths of a millimetre size in the bauxite matrix (fig. 141). In France, Caillère and Pobeguin (1968) assayed 6 to 28% siderite in some samples from the Nalzen deposit in the Ariège. In this grey, reduced bauxite, siderite mainly forms minute glinting plates and fine veins. In the Unjat deposit of this same group, Combes (1959) observed an epigene ferruginous crust composed predominantly of siderite and goethite.

In Yugoslavia, the Triasssic bauxite of the Gornje Polje deposit group in Crna Gora (Montenegro) also contains a little siderite in the form of minute spherulites according to Luković (1962). The present writer has determined 1 to 3% siderite in some samples of grey bauxite from the Triassic bauxite horizon of Slovenia, 1 to 2% in the grey and green Oligocene bauxites of the Velika Planina, and likewise 1 to 2% in samples from the small bauxite deposit of Untersberg near Salzburg in Austria. In the lower part of the Campo Felice deposits, Italy, he has found 0·1 to 1·2% siderite, usually in the places where green clays grade into a red bauxite.

Rhodochrosite, $MnCO_3$. A very rare mineral in bauxite; the writer has identified it in abundances of 1 to 2% in samples he had collected in the La Boissière deposit group of Southern France. Its presence is explained by the remarkably high manganese content of that bauxite, which is largely in the form of lithiophorite and todorokite. The bauxite is impregnated by 15 to 25% of epigene calcite. Hydrocarbonate-rich ground waters seeping through the bauxite had presumably dissolved a little manganese which, together with the calcite, came to be precipitated as rhodochrosite. The mineral has been identified also in some samples from the Campo Felice deposits, Italy, in abundances of 0·4 to 0·5%.

Aragonite, $CaCO_3$. Kiss (1955) found a cave of several metres height and about half a metre width along a post-Eocene fault plane in the northeastern part of the Gánt–Hosszúharasztos deposit, Hungary, the walls of which were covered by a cauliflower-like or jagged aragonite crust 5 to 20 cm thick. Most of the aragonite had changed into calcite secondarily. Kiss attributed the aragonite to hot-water solutions rising along the fault.

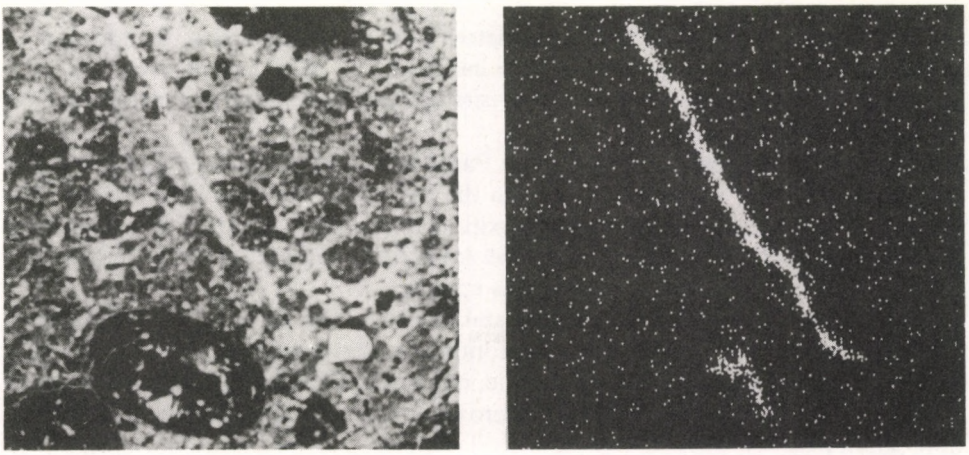


Figure 142. A microcrack filled with bastnaesite in a bauxite. San Giovanni Rotondo, Italy. Electron microprobe images, 600 ×
(A) electron image; (B) X-ray image of cerium distribution

Alumohydrocalcite, $CaAl_2(OH)_4(CO_3)_2 \cdot 3H_2O$. A very rare mineral in bauxite, it was identified by Beneslavsky (1963) in the form of segregations of several millimetres size in the Devonian deposits of the Northern Urals, USSR. It fills out cavities and pores in those bauxites in association with calcite and diaspore.

It has also been discovered in the small bauxite deposits of the Tulsk–Ryazansk region, USSR, where it forms smallish reniform segregations. Under the microscope, it is seen to form minute acicular crystals. It is accompanied by gibbsite and halloysite (Trubina 1958).

Malachite, $Cu_2(OH)_2CO_3$. De Weisse (1975) found a greenish incrustation a few millimetres thick at the bedrock boundary of a bauxite in a small open pit on Orsello Hill in the Abruzzi, Italy; it covered the bedrock limestone in an area of several square decimetres. X-ray and chemical analyses proved the crust to consist of malachite. It is of supergene origin according to de Weisse. Rainwater seeping through the bauxite in the exposure could dissolve its copper content. At the bedrock boundary, the precipitation of malachite was triggered by the neutralisation of the formerly acidic solutions.

In the deposit called Nazda in the Vlasenica deposit group of Yugoslavia, there is in places a vivid green porous layer 1 to 3 mm thick at the boundary of bauxite and bedrock. It is accompanied by a jet-black cryptomelane layer 2 to 3 mm thick on the bedrock side. The green layer has been identified as a highly crystalline malachite by X-ray analyses performed by the writer, who assumes both layers to be of epigene origin. They may have formed by a process similar to the one described by de Weisse for the malachite segregation at Orsello.

Bastnaesite, $CeFCO_3$. Bárdossy and Pantó (1973) could identify this mineral in several deposit groups, the first of which was San Giovanni Rotondo in Italy. It was found at the bottom of one of the deposits there, filling micropores of 10 to 100 μm diameter and forming microcrack fillings. The cracks of irregular course are 2 to 5 μm wide and 200 to 400 μm long (fig. 142). In places, bastnaesite is enriched in nebular

impregnations, of 100 to 200 μm diameter, with no sharp limits. In addition to predominant cerium, it contains a little lanthane and even less promethium. It is accompanied in places by roentgenite, parisite and synchisite, the calcic members of the bastnaesite group.

In the profile examined, Bárdossy and Pantó have found monazite grains of a few micrometres in size in every sample from the bottom to the top of the deposit. The ground waters seeping through the bauxite had presumably partly dissolved these grains: bastnaesite was then precipitated from the solutions at the bottom of the deposit. It is thus justified to regard it as epigene.

Bastnaesite was found by Bárdossy and Pantó in the bauxite of Nagyharsány, Hungary, which in a brick-red matrix contains greenish bauxite pebbles. This is a secondary occurrence, as is the Italian one; it is associated with the calcic members of the bastnaesite group in pore fillings, microveins, rambling segregations and nebular impregnations.

Parisite, $CaCe_2\ F_2(CO_3)_3$; *Roentgenite*, $Ca_2Ce_3\ F_3(CO_3)_2$; *Synchisite*, $Ca\ Ce\ F(CO_3)_2$. These three calcic members of the bastnaesite group can be identified as a subgroup but cannot be distinguished from one another by means of the electron microprobe, containing as they do the same elements in slightly different proportions.

Bárdossy and Pantó (1973) first found in an iron-rich bauxite pebble in a bauxite sample of conglomerate texture from the Allauch deposit near Marseilles, France, a few grains of irregular shape and 10 to 15 μm in size which contained in addition to cerium somewhat less neodymium and lanthane, as well as calcium and fluorine. The grains also contained some carbonate. It could not be decided beyond doubt whether these minute segregations had formed *in situ* or were redeposited in their present form into the bauxite in which they are found, embedded in the iron-rich pebbles. The extreme compactness and the corresponding low permeability of the iron-rich pebbles favours the second possibility.

The Nagyharsány sample mentioned in connection with bastnaesite was found to be remarkably rich in parisite, roentgenite and synchisite. All these are typical secondary segregations, pore fillings of irregular shape and 100 to 200 μm in size, nebular impregnations with no sharp limits in the matrix and in some ooids, rambling segregations next to some bauxite pebbles, and microveins of irregular course. These latter are of 2 to 20 μm width and several hundred micrometres in length: they can be seen to cross in places (fig. 143).

In addition to cerium, they contain smaller amounts of lanthane and neodymium. In a minute segregation, the order of decreasing abundance was Ca > La > Nd > Ce > P > Th > Sm. In this particle, then, the leading role of cerium was taken over by lanthane and neodymium; phosphorus also appeared. In another segregation, a small quantity of vanadium also was identified.

In the above sample, as well as in other bauxite samples from Nagyharsány, many grains of monazite were identified. These also contained smaller quantities of lanthane and neodymium in addition to cerium. The minerals of the bastnaesite group are presumably the products of an epigene leaching of the monazite grains and of precipitation out of the solutions thus formed. Their lanthane and neodymium content also confirms this assumption.

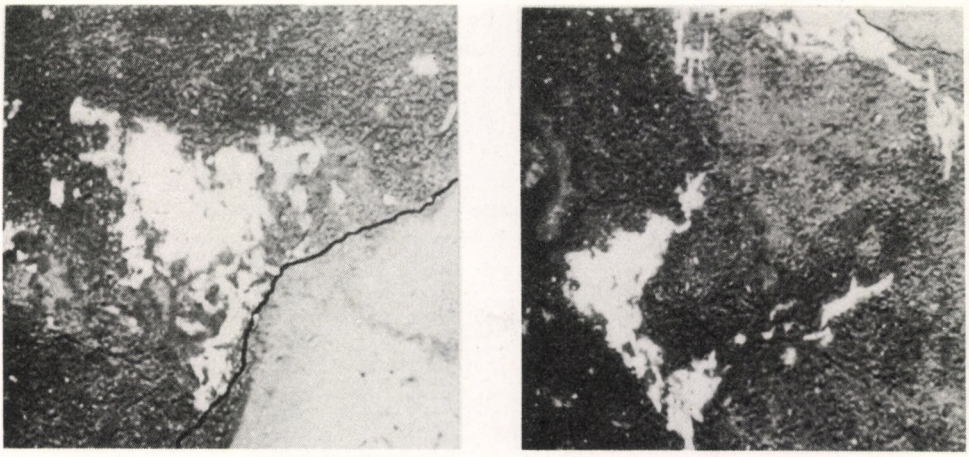


Figure 143. Segregations of irregular shape and veins of parisite, roentgenite and synchisite in bauxite. Nagyharsány, Hungary. Electron microprobe, electron images, 300 ×

5.3.4. SULPHIDES AND ARSENIDES

Pyrite, FeS_2. The most frequent sulphide mineral of karst bauxites, it has so far been identified in 48 deposit groups. Concentrated in the reduced, grey, green and black bauxite types above all, it is found sporadically and in minute quantities also in the oxidised ones. Its average abundance is 5 to 15% in the reduced bauxite types, but may attain even 30 to 46%. Most of the pyrite is diagenetic: syngenetic pyrite is rarer, epigene pyrite is scarce.

In Hungary, the highest abundances of pyrite were found in reduced grey bauxite at Iszkaszentgyörgy and in the Halimba–Cseres deposit. Pyrite is abundant also in the grey bauxites of Nagyegyháza and Bakonyoszlop. It is sporadic in the Halimba Basin and in the deposit groups of Nyirád.

The pyrite grains are varied in morphology and composition, features which were studied by means of the electron microprobe, Bárdossy and Pantó (1972). The following principal types could be distingished.

Automorphic grains, in the size range from 0·005 mm to 8 mm. The most frequent forms are the hexahedral and octahedral; the dodecahedral is less frequent. Such grains are found in fine dissemination in the matrix of bauxite, but more seldom also in its ooids and pisoids. According to Komlóssy (1966), the grain size of automorphic pyrite grains gradually increases from the top downward in the grey bauxite at Iszkaszentgyörgy, Hungary: they are smaller than 0·06 mm at the top, and larger than 0·3 mm at the bottom. The writer has found pyrite grains several millimetres size in Félix I mine at Szőc, Hungary. Some of them were of a typical zoned constitution (fig. 144). Bárdossy and Pantó using the electron microprobe and the X-ray diffractometer found that the light bands consist of pure pyrite, whereas the dark ones contain extremely fine inclusions of gibbsite, boehmite, anatase and rutile. The pyrite in the course of its growth presumably rhythmically engulfed these mineral grains smaller



Figure 144. Zoned pyrite crystals. Szőc, Félix I mine, Hungary
(A) polished section, 163 × ; (B) electron microprobe, electron image, 268 ×

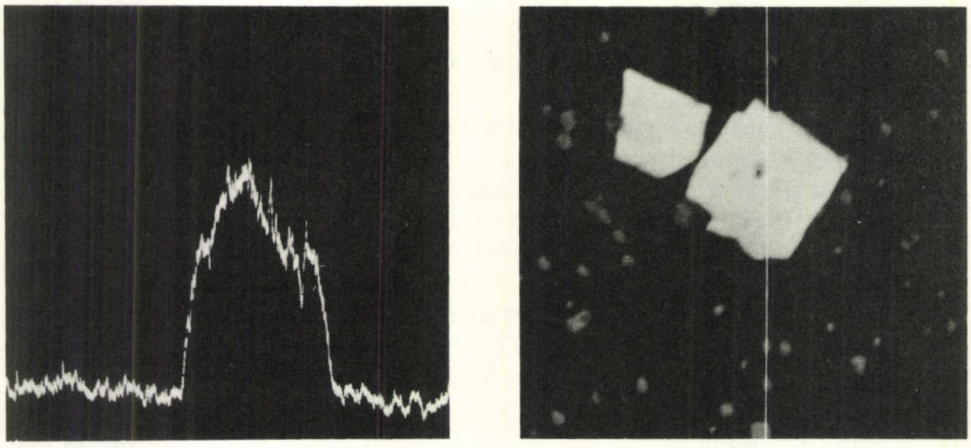


Figure 145. A nickel-containing pyrite grain in a Middle Triassic bauxite. Podlipa, Slovenia, Yugoslavia. Electron microprobe images, 537 ×
(A) electron image; (B) X-ray profile of nickel distribution along white line indicated on (A)

than a micrometre in size. The nuclei of the large pyrite grains usually consist of pure pyrite.

Automorphic pyrite grains, usually of less than 0·5 mm in size, are frequent in the Devonian deposits of the Northern Urals, USSR (Beneslavsky 1963). They are rather evenly spread in the matrix, forming aggregates in a few places.

The automorphic pyrite grains are often trace-element scavengers. In the pyrite grains of a Triassic bauxite from Slovenia, Yugoslavia, Bárdossy and Pantó (1972) have observed a concentration of nickel. The more automorphic the grains, the higher their nickel content. The highest nickel content was 2%. One such grain with the diagram of its nickel concentration is shown as figure 145.

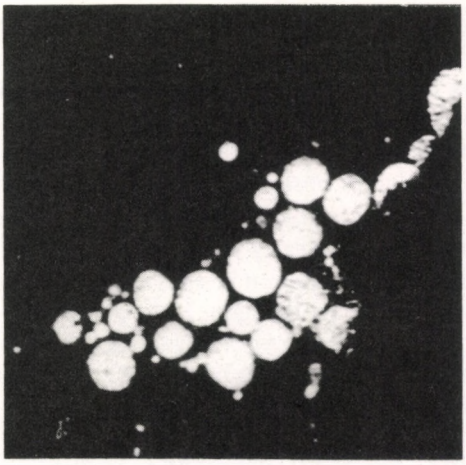


Figure 146. Spherical bacteriopyrite grains in a bauxite. Szőc, Félix I mine, Hungary. Electron microprobe, electron image. 537 ×

The automorphic, well-crystallized pyrite grains are in places aggregated into *nodes*, the diameter of which may attain up to 3 cm. Such nodes were found by Dragović (1966) in light grey bauxite types of the Bijele Poljane deposit group of Crna Gora (Montenegro), Yugoslavia. The matrix between the nodes exhibited finely disperse pyrite crystals of about a tenth-millimetre in size.

Formless (irregular) pyrite grains are somewhat rarer than the automorphic ones. Some are of highly irregular, rambling form, most often of 50 to 500 μm in size. Microprobe examinations by Bárdossy and Pantó have proved them not to contain any trace element concentrations. Caillère and Pobeguin (1963) found in a grey bauxite of the Durban deposit, Ariège, France, some pisoids filled out almost exclusively by such a pyrite.

Spherical pyrite grains of 2 to 50 μm in size are even rarer than the above types. They are very probably the products of bacterial activity; this is why they are usually called *bacteriopyrites*. They most often occur in aggregates of larger or smaller size. Bárdossy and Pantó found large numbers of bacteriopyrite aggregates in the Félix I mine at Szőc, Hungary (fig. 146). Burić (1966) observed bacteriopyrite grains of 10 to 100 μm in size in the Gornje Polje deposit of Crna Gora (Montenegro), Yugoslavia.

There are a few rare pyrite nodes of 1 to 20 mm in size and spherical or ellipsoidal shape in grey bauxite, such as those observed by Komlóssy (1966) at Iszkaszentgyörgy, Hungary and by Dragović (1966) in the Bijele Poljane deposit of Crna Gora (Montenegro), Yugoslavia. Very rarely, pyrite may also form *coatings*. According to Beneslavsky (1963), e.g. some of the pisoids of the Devonian bauxites of the Northern Urals, USSR, are covered with films of pyrite.

Very seldom, pyrite forms *fissure fillings*, partly in grey bauxite, and partly in the red and multicoloured bauxite right beneath it. In Hungary, such fissure filings were found by Komlóssy (1966) in the Iszkaszentgyörgy deposits. The present writer has found it also to be frequent in the Halimba II ore body. The cracks there are 0,5 to 20 mm wide, and of a dark grey colour: in addition to pyrite, they contain organic matter and a little

kaolinite. Fissure fillings consisting of pyrite are mentioned by Beneslavsky (1963) from the deposits of the Northern Urals, USSR, and by Jurković (1965) from the Suvaja-Šolaja deposits of Yugoslavia. All these are of an obviously epigene origin.

The pyrite grains found by Combes (1969) in one of the deposits near Unjat, Ariège, France, in a haematitic *crust* at the bedrock boundary, are likewise epigene. In the nearby Péreille deposit, Combes observed a crust consisting of pyrite at the bedrock boundary. The crust is several centimetres thick, greyish black and very hard. It is accompanied by a little gypsum, clearly the product of pyrite oxidation.

A pyrite of colloidal apperance, present in several bauxite deposits of the USSR, was described by Beneslavsky (1963) by the name *melnikovite–pyrite*. According to Strunz (1970), this is just a cryptocrystalline pyrite of gel-like appearance which is unjustified to regard as an independent mineral. Cryptocrystalline melnikovite–pyrite seems to grade into macrocrystalline pyrite in many places.

Bravoite (nickel-pyrite), $(Fe, Ni)S_2$. In grey and greenish-grey chamositic bauxite in one of the Lower Malm bauxite exposures near Distomon, Greece, Bárdossy *et al* (1973) found some pyrite grains substantially enriched in nickel. The electron microprobe analysis of one of the particles gave the following result: Fe, 39·0%; Ni, 6·0%; Co, 0·2%; As, 0·9%; S, 51·8%, a total of 97·9%. Pyrite of such constitution is called bravoite in the literature. Bravoite grains are of less than 10 μm in size as a rule in this bauxite. Some of them have been oxidised secondarily so that today only pseudomorphs remain. Other grains are surrounded by a zone of oxidation, with only their nuclei preserving the original composition. Figure 147 shows such a bravoite grain.

The chemical composition of the grain was determined along the line shown in white. Sulphur is entirely absent from the outer, oxidised part of the grain. In the oxidised zone, nickel migrated outward, attaining a peak of secondary enrichment near the periphery. Cobalt behaved similarly. The removal of sulphur resulted in a relative concentration of iron. The authors found the oxidised zone to consist of nickel-bearing goethite. They assumed a syngenetic reduction to have acted on this bauxite horizon, on the evidence of the position and distribution of the reduced bauxite types. In their opinion, the bravoite grains are syngenetic, as are all the other reduced minerals here.

Marcasite, FeS_2. Very rare on its own, it almost invariably occurs in the company of pyrite. Its abundance, much less than that of pyrite, seldom attains 5%. In Hungary it forms according to Komlóssy (1966) the nuclei of the pyrite concretions in Iszkaszentgyörgy bauxite. The present writer has demonstrated by X-ray measurements that it also occurs in Halimba and Nyirád bauxite.

According to Beneslavsky (1963), it is found in the Northern Urals bauxites, USSR, in roundish nests and reniform segregations, the cores of which often consist of cryptocrystalline marcasite. Combes (1969) states that there are, in places, nests of pyrite and marcasite in a claystone layer at the base of the Durban deposit of the Ariège, France. The majority of the grains has been secondarily oxidised into goethite.

Hydrotroilite, $FeS \cdot nH_2O$. Papiu *et al* (1970b) assumed some colloidal hydrotroilite accompanied by a great deal of organic matter, to be present in the black bauxite types of the Pădurea Craiului deposits of Rumania.

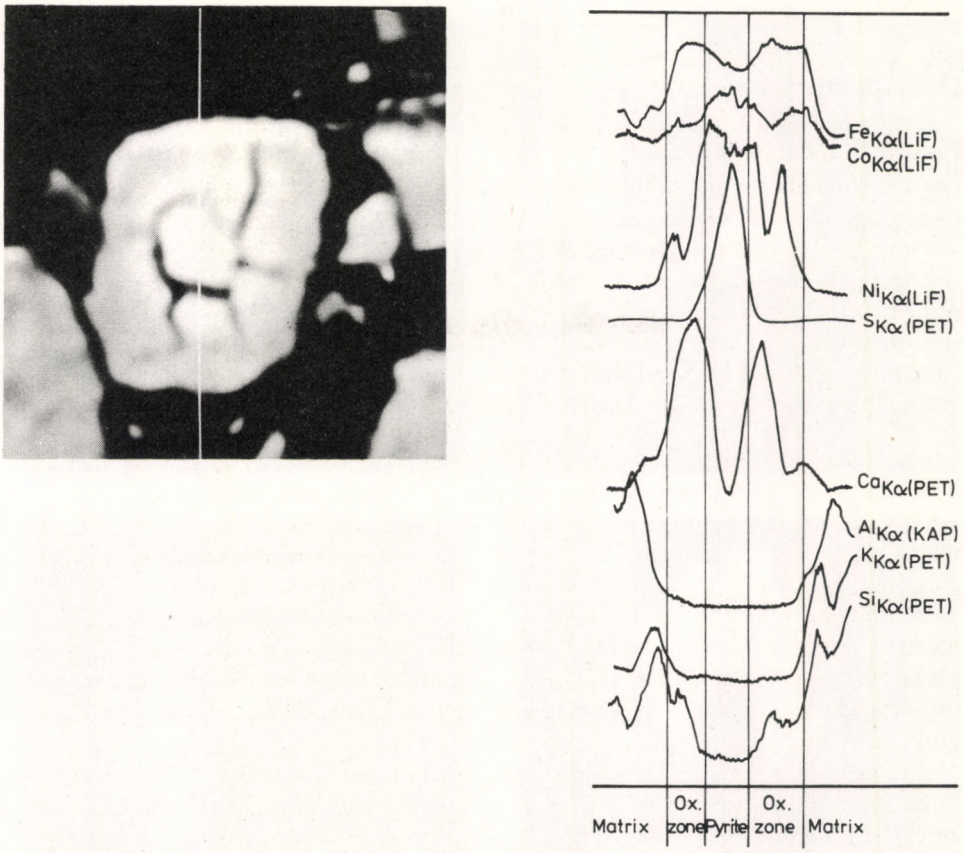


Figure 147. A bravoite grain with an oxidised halo in a Malm bauxite. Distomon, Greece. Electron microprobe

(A) electron image, 2234 × ; (B) distribution of elements along the white line in (A)

Sphalerite, ZnS. A few sphalerite grains of irregular outline and 20 to 30 μm in size (fig. 148) were found by Bárdossy and Pantó using the electron microprobe on a sample from Gánt–Újfeltárás. The form of the grains make their *in situ* formation probable. In the USSR, it is found sporadically in the company of chalcopyrite in the Devonian bauxites of the Northern Urals (Beneslavsky 1963). In Southern Timan bauxite, Gorbachev (1966b) found some tubular formations of 1·5 cm diameter, consisting of pyrite and marcasite. In their inner cavities, he observed inward-projecting sphalerite crystals of up to 4 mm in size. Gorbachev attributed them to secondary processes under reducing conditions, subsequent to the segregation of pyrite and marcasite.

Dietrich *et al* (1976) found sphalerite spherules of one millimetre size accompanied by takovite and erythrine in the Blanquette deposit of France. Their nodular–concretionary appearance implies a secondary origin.

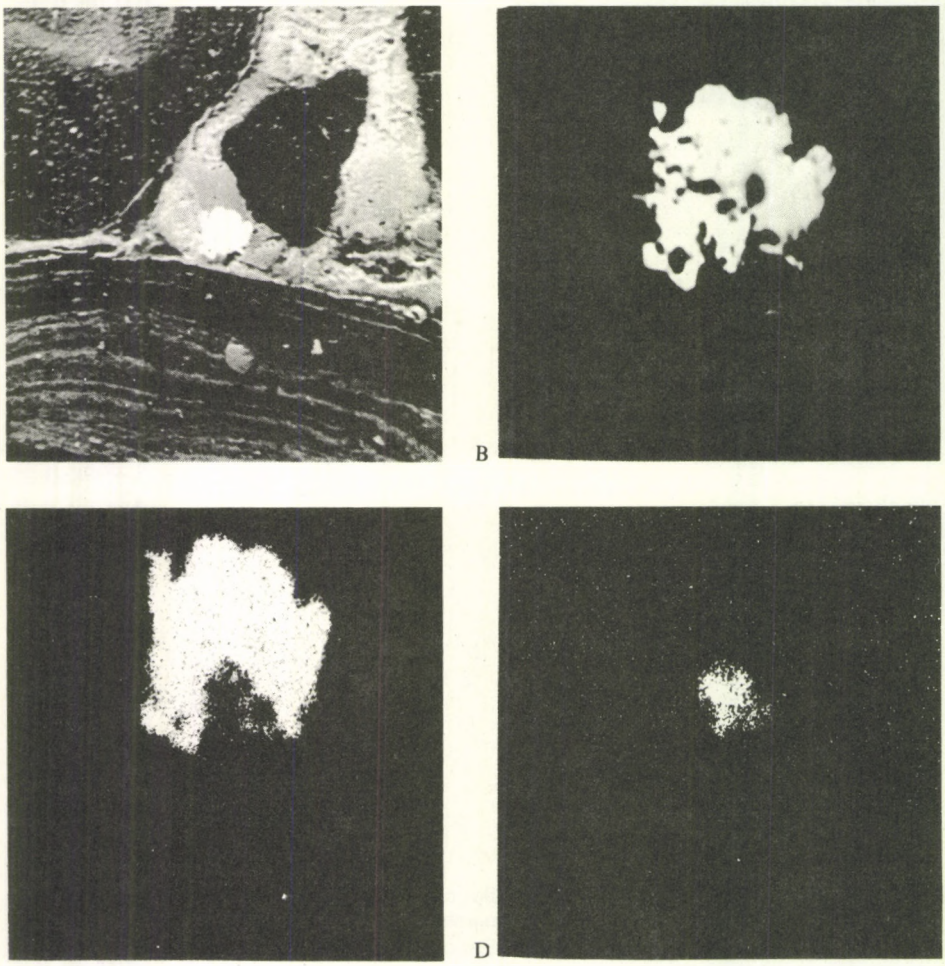


Figure 148. Sphalerite and galena in a yellowish-green bauxite. Gánt–Újfeltárás, Hungary. Electron microprobe images

(A) electron image, 300 × ; (B) *idem*, strong contrast; (C) X-ray image of zinc distribution, 1200 × ; (D) X-ray image of lead distribution, 1200 ×

Galena, PbS. It was identifed in a bauxite of conglomerate texture from Gánt–Újfeltárás, Hungary, by Bárdossy and Pantó's electron microprobe studies. Intergrown with sphalerite, the particle was of 8 by 10 μm in size (fig. 148). It had presumably formed *in situ*, together with the sphalerite, as confirmed by the fact that the bauxite in which both minerals were found is of a reduced character and of chamositic composition.

Chalcopyrite, $CuFeS_2$. An extremely rare mineral in bauxite, it was identified by Beneslavsky (1963) in independent grains and aggregates in the Devonian bauxites of the Northern Urals, USSR. It sometimes forms very fine inclusions in pyrite crystals.

He observed in place the chalcopyrite grains to wear coatings of *bornite*, *chalcocite* and *covellite*. In a bauxite sample from the Northern Urals, Bárdossy and Pantó (1970) could also identify a chalcopyrite grain of 40 by 60 μm of size. The extremely irregular, rambling form of the particle suggests its authigenic origin (cf fig. 72). Vörös and Gecse (1976) found a few grains of chalcopyrite in samples from the Iszkaszentgyörgy–Bitó deposit, Hungary.

Luković (1962) found a few chalcopyrite grains in a Triassic bauxite of the Gornje Polje deposit group, Crna Gora (Montenegro), Yugoslavia. De Weisse (1973) described chalcopyrite from the Eleusis deposit group of Greece, without characterizing it in any detail.

Bornite, Cu_5FeS_4. Known so far only from the Devonian bauxites of the Northern Urals, USSR (Beneslavsky 1963), it forms minute inclusions there in a heaematite produced by pyrite oxidation. It may originally have been an inclusion in the pyrite.

Niccolite, NiAs. Its single occurrence so far is in the Blanquette deposit near Brignoles. France, where it was identified by Dietrich *et al* (1976) in concretions of 2 by 3 cm in size in a pisoidic bauxite. Niccolite forms a ring-shaped segregation of 2 to 5 μm in size and of irregular outline in the nucleus of the concretion. It is presumably of secondary origin.

5.3.5. SULPHATES

Alunite, $(K, Na)Al_3(OH)_6(SO_4)_2$. The most frequent sulphate mineral of karst bauxites, it forms in them nests and concretions ranging in size from a few millimetres to several decimetres. An epigene and supergene mineral, it is the product of pyrite oxidation as a rule.

The first to recognise alunite in a Hungarian bauxite was Vadász (1943), who in the upper part of the Kincses ore body at Iszkaszentgyörgy found some cauldron-shaped formations filled with pink bauxite, containing some white alunite nodes of 1 to 10 cm in size. Submitting this material to an X-ray analysis, Bidló (1955) found the white nests to contain some 85% alunite in addition to some 10% gibbsite. In a white band a few centimetres thick, interrupted here and there, parallel to the cover in the top part of the bauxite in the Sándor mine at Nyirád, Gedeon (1946) assayed 37% alunite, 43% kaolinite and 14% gibbsite. Kiss (1952) observed a few sporadic alunite grains of a few millimetres in size in Nézsa bauxite.

Further to Hungary, in Lenses II and III at Nyirád–Izamajor, the writer has found at several places white or pink nests of alunite of 2 to 10 cm in size in yellow-veined rust-red bauxite. There were even some exceptionally big nests of 30 to 80 cm in size, consisting predominantly of alunite accompanied by a few percent gibbsite, boehmite and kaolinite. In the open pit working Lens VI of Nyirád–Darvastó, the writer has found white nests of 5 to 20 cm size, which contained 75 to 95% alunite accompanied by gibbsite, gypsum and kaolinite. In the Malomvölgy open pit at Halimba, the writer has observed nests of a few centimetres size consisting of alunite and halloysite. In open pit no. I at Eplény, he has found in the upper part of the ore body (fig. 149) a meat-red white-spotted bauxite which contained 25% alunite in its red parts and 45% in its white

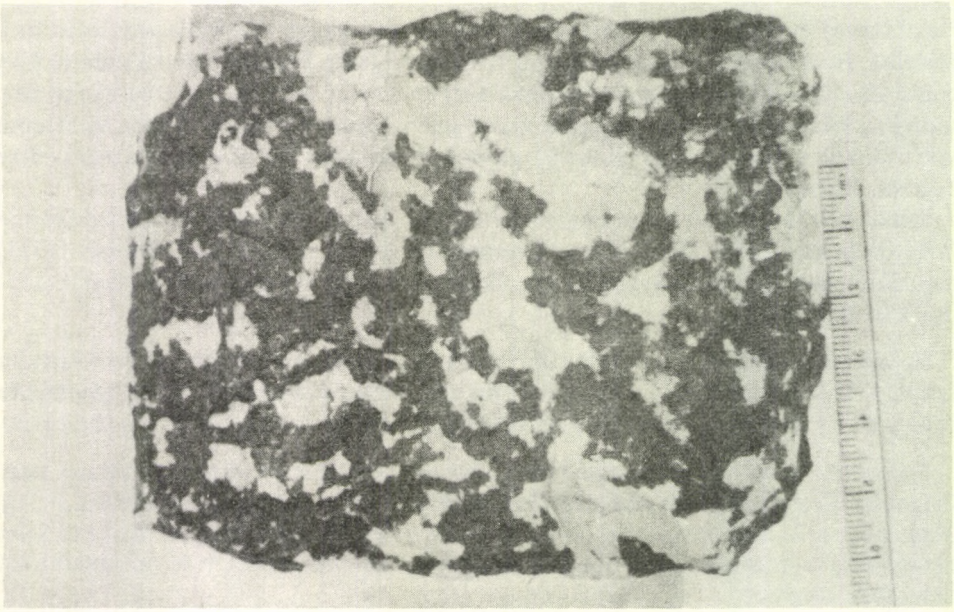


Figure 149. Alunitic bauxitic. Eplény I open pit, Hungary

ones. In all these deposit groups, the alunitic nests are found in the upper parts of the deposit, in the proximity of partly or totally oxidised pyritic bauxite types.

In the bauxitic clay deposits of Cserszegtomaj, Bárdossy (1961b) observed in an epigene gibbsite layer at the boundary of the bedrock dolomite some compact reniform nodes of 0·5 tq 1·0 cm in size which in addition to kaolinite and alumogel contained 20 to 30% alunite.

Chemical and X-ray analyses have proved alunite at Nyirád, Halimba, Szőc and Iszkaszentgyörgy to be predominantly sodic with a subordinate potassium content. In the Eplény and Cserszegtomaj alunites, on to the other hand, potassium is more abundant than sodium. There is no pure sodic or potassic alunite in any of the Hungarian bauxites.

In agreement with the opinions put forward by Gedeon (1946) and Kiss (1952), the present writer also attributes the formation of these alunite nests to the decomposition of pyrite. Epigene and supergene oxidation of pyrite gave rise to solutions containing sulphuric acid. Seeping downward, these dissolved some of the aluminium minerals, and were then gradually neutralised on their way down. Alunite was precipitated, tying down the sodium and potassium contained in the ground water solution, when the pH attained about 4 to 5. This process could take place in surface outcrops or under a thinnish overburden above all. This is why no alunite is found in any bauxite deposits deeper than 100 m, e.g. in the mines Halimba II and III. In Iszkaszentgyörgy also, it is mainly concentrated in the open pits already worked out: it is very rare in the underground Kincses II mine, and has not been observed so far in the deep underground workings of József II and III mines (oral communication by Bárdos 1973).

In a grey pyritic–marcasitic bauxite from near Cusano Mutri, Monti Matese, Italy, the writer has found a few percent of alunite finely dispersed in the matrix. Beneslavsky (1963) mentions alunite from the deposit groups of the Turgay Plateau and Mugojar, USSR, without any more detailed characterisation. Ashraf *et al* (1972) state that some clayey bauxites and bauxitic clays in the bauxite deposits of the Salt Range in Pakistan contain some sodium alunite. Calculating backward from the chemical analyses, the mineral is found to attain an abundance of 15 to 20% there.

Aluminite, $Al_2(OH)_4(SO_4) \cdot 7H_2O$. Much rarer than alunite, it was found by Gedeon (1955) in the immediate cover of Gánt bauxite, Hungary, intercalated between a brown clay and a freshwater limestone. Also in Hungary, in an open pit on the Szőlőhegy at Sümeg, Hungary, the writer has found a white layer a few centimetres thick between the bauxite and the limestone underlying it, consisting of 60% aluminite and 36% gibbsite, the remaining 4% being made up by kaolinite, goethite and calcite.

Snow-white reniform nests of 1 to 8 cm in size from a red bauxite in the Nyireskút II underground working at Szőc have been found by the writer to consist of pure aluminite. In a red bauxite in the central part of the Félix II mine at Szőc, the writer has found horizontal white aluminite segregations 1 to 4 cm thick, which could be followed for more than 20 m in a roadway. In places, these segregations were accompanied by fibrous gypsum in a thickness of 2 to 4 mm (Bárdossy and Sajgó 1968).

Gladkovsky and Rudnova (1964) identified the mineral in the Devonian bauxites of the Southern Urals, USSR, where it occurs in white compact mammillary and reniform nests of 6 to 20 cm in size in the lower parts of the deposits. It is less frequent than basaluminite, also known from those deposits. In the Študenac deposit of Crna Gora (Montenegro), Yugoslavia, there are cracks filled with white porous aluminite in red and pink bauxite, accompanied in places by segregations of pure gibbsite (Luković 1970).

Its forms of occurrence imply that aluminite is an epigene and supergene mineral beyond any doubt, whose origin is connected with the oxidation of pyrite and marcasite. It was precipitated in the course of the neutralisation of downward-seeping sulphuric acid solution. For aluminite to form, the ground water must contain no sodium or potassium, which would permit the formation of alunite. Precipitation presumably took place as a rule where the acidic solutions reached either the bedrock or the ground water table, the ground water being weakly alkaline under the influence of the carbonate bedrock.

Basaluminite, $Al_4(OH)_{10}(SO_4) \cdot 5H_5O$. Vendel *et al* (1971) identified a little basaluminite accompanying gypsum in a layer 1 to 2 cm thick in the top part of the deposit in the Gánt–Harasztos open pit, Hungary. The authors hold both minerals to be of epigene origin. Gladkovsky and Rudnova (1964) identified basaluminite in the Devonian deposits of the Southern Urals, USSR, where it forms nests of several centimetres in size in the lower part of the bauxite body. Most frequent at a height of a few centimetres above the top of the underlying limestone, it is accompanied by gibbsite and allophane (fig. 150). It is dove-grey, hard and columnar in places, soft and snow-white elsewhere. In dry air it gradually loses its crystal water, crumbling into a white powder. According to the authors cited, it is the product of the epigene and supergene oxidation of grey pyritic bauxite in the upper part of the deposits. The

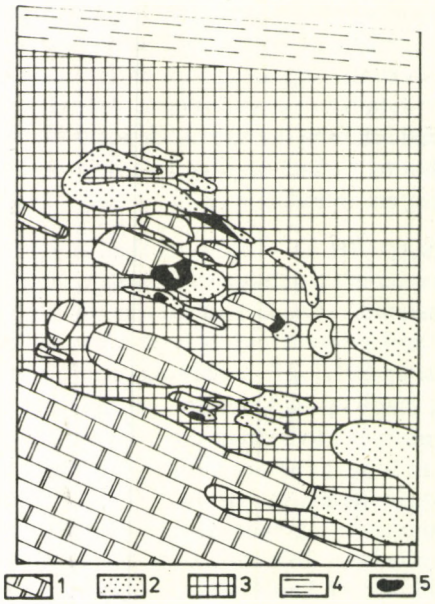


Figure 150. Basaluminite nests in bauxite. Mezhevoy Log deposit, Southern Urals, USSR. After Gladkovsky and Rudnova (1964)

(1) bedrock limestone; (2) basaluminite nests; (3) red, hard bauxite; (4) multicoloured bauxite; (5) allophane nests

sulphuric acid formed in the process dissolved some aluminium out of the rock in the course of its downward seepage. Arriving near the bedrock, these solutions were neutralised so that basaluminite could precipitate. This is when, according to those authors, the gibbsite and allophane accompanying the basaluminite also came to exist.

Lahodny-Šarc *et al* (1969) found a node consisting of basaluminite accompanied by a little gypsum in the Paprati deposit near Nikšić in Crna Gora (Montenegro), Yugoslavia. The mineral was highly crystalline. It is held to be an epigene product of pyrite oxidation.

The formation of basaluminite also presupposes the absence of alkalies from ground water. The reason why basaluminite was formed in some cases and aluminite in others is unclarified as yet.

Metabasaluminite, $Al_4(OH)_{10}(SO_4)$. This mineral was identified by Bárdossy *et al* (1978a) in one of the bauxite deposits near Tükrösmajor, Hungary, filling cracks 1 to 2 mm wide in ochroid bauxite, accompanied by smaller amounts of bassanite. Forming well-developed columnar crystals of several hundred μm in size. It presumably is the oxidation product of the pyrite found in the top part of the deposits. It is not understood why aluminite is formed in one case, basaluminite in the next, and metabasaluminite elsewhere. Metabasaluminite may possibly be the product of gradual dehydration of hydrous aluminium sulphates.

Melanterite, $FeSO_4\ 7H_2O$. First identified by Bárdossy (1954) in bauxite from Szőc –Határvölgy, Hungary, it forms pale green, fibrous platelets a few millimetres thick in

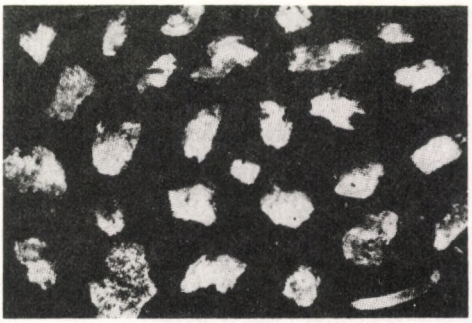


Figure 151. Segregations of melanterite in a grey bauxite. Halimba, Cseres mine, Hungary. Natural size

the lower part of a grey pyritic bauxite. The writer has since identified it in the deposits of Halimba, Szőc, Nyirád and Kislőd, invariably in grey pyritic bauxite (fig. 151). According to Komlóssy (1966), it is ubiquitous also in the grey pyritic bauxite of Iszkaszentgyörgy, Hungary.

Melanterite is a product of the epigene and supergene oxidation of pyrite and marcasite. It is being formed even today in the roadways wherever the air is moist enough, it the shape of hair-thin acicular crystals of several centimetres length on free rock faces and in cracks and cavities in grey bauxite. In dry air, these crystals soon crumble into a white powder.

According to Beneslavsky (1963), melanterite is formed in a similar fashion and at a very fast rate in Devonian grey pyritic–marcasitic bauxite in the Northern Urals, USSR. Lumps of ore containing sulphides are decomposed within half a year, their pyrite content being converted into melanterite. Gladkovsky and Rudnova (1964) identified melanterite in strongly altered pyritic bauxite in the Devonian deposits of the Southern Urals, USSR, where it forms pale green, translucent coatings and veins. It is held to be a supergene alteration product of pyrite.

Rosenite, $FeSO_4 \cdot 4H_2O$. The writer demonstrated its presence by X-ray diffractometry in an abundance of a few percent in some samples of grey pyritic bauxite from the József III ore body of Iszkaszentgyörgy. Accompanied by alunite and szomolnokite, it is a product of the epigene and supergene oxidation of pyrite.

Szomolnokite, $FeSO_4 \cdot H_2O$. It was in grey pyritic bauxite from József III ore body, Iszkaszentgyörgy, Hungary, that the writer could identify it in the company of rosenite, in abundances of 1 to 14%, highest at the bottom of the grey pyritic bauxite. Likewise a product of epigene and supergene pyrite oxidation.

Butlerite, $Fe^{3+}(OH)SO_4 \cdot 2H_2O$. A very rare sulphate mineral, it was identified by Beneslavsky (1963) in some Devonian bauxites of the Northern Urals, USSR, partly in the form of microcrack fillings, partly as very fine films coating pyrite grains. A supergene mineral, it is found predominantly in the outcrops of bauxite.

Parabutlerite, $Fe^{3+}(OH)SO_4 \cdot 2H_2O$. The writer identified it in quantities of a few percent in grey pyritic bauxite from the deposits Nyirád–Izamajor and

Iszkaszentgyörgy–József deposit, Hungary. Finely distributed in the matrix, it is accompanied by alunite and gypsum at Nyirád, and by alunite and szomolnokite at Iszkaszentgyörgy. A product of epigene and supergene pyrite oxidation.

Fibroferrite, $Fe^{3+}(OH)SO_4 \cdot 5H_2O$. In a light grey bauxite from the upper part of the Regia Piana deposit, Monti Matese, Italy, the writer has assayed 3% fibroferrite accompanied by melanterite and gypsum, formed in the course of the supergene weathering of a pyritic bauxite.

Halotrichite, $Fe^{2+}Al_2(SO_4)_4 \cdot 22H_2O$. Gladkovsky and Rudnova (1964) state it to be present in the Devonian bauxites of the Northern Urals, USSR, in the company of melanterite. It is a supergene oxidation product of pyritic–marcasitic bauxite.

Gypsum, $CaSO_4 \cdot 2H_2O$. It is found in bauxites the pyrite and macrasite content of which or of whose immediate cover got oxidised. In Hungary, in the upper parts of the Gánt deposits, there are automorphic gypsum crystals of several centimetres in size. Rare in the bauxite proper, they are more frequent in the grey and multicoloured clay beds of the immediate cover. Dudich (1965) states gypsum to attain a content of 20 to 25% in places in grey pyritic–marcasitic bauxite in Nyirád–Darvastó, whereas in the red bauxite it does not exceed one percent. In the deposit Szőc–Félix II (Bárdossy and Sajgó 1968), fissure fillings of gypsum were found to accompany aluminite nodes. The fissures are 1 to 10 mm wide, and the gypsum crystals are grown perpendicularly on their walls. This gives the fissure filling a fibrous appearance.

In the USSR, Budnikov (1971) states the clayey bauxite deposits of the Taimyr Peninsula to be traversed by veins of secondary gypsum, which in places makes up two thirds of the bauxite. This is the greatest concentration of gypsum in bauxite known from any deposit so far. In the exposures of the Northern Urals deposit groups, gypsum, associated with a number of other sulphate minerals, is a product of the supergene alteration of pyrite (Beneslavsky 1961, 2nd ed. 1974).

In the Fuentespalda deposit of Spain, Combes (1969) found a crust 1 to 2 cm thick on the footwall contact made up of gypsum and goethite. He found gypsum to penetrate the underlying dolomitic limestone in the form of fine veins to a depth of 5 to 10 cm. In this case, then, some of the sulphate solutions kept seeping downwards even after the neutralisation of iron sulphate and the precipitation of goethite, and it was only in the bedrock that all the gypsum got precipitated out of them.

Bassanite, $2CaSO_4 \cdot H_2O$. Identified by Bárdossy *et al* (1978a) in bauxite from Tükrösmajor, Nagyegyháza deposit group, Hungary, it occurs in the form of flat crystals of rhomboidal form in fissures 1 to 2 mm wide together with metabasaluminite in ochroid clayey bauxite. Its abundance gradually decreases downward. It is presumably an oxidation product of a grey pyritic bauxite forming the top part of the deposit. The oxidation of pyritic bauxites usually results in the formation of gypsum. The peculiar and obviously rare conditions which are needed for bassanite to form instead of gypsum are not understood as yet.

Anhydrite, $CaSO_4$. In the Szőc–Félix I mine, Hungary, there is a pink bauxite with red spots traversed by dense veins of gypsum right underneath the grey pyritic bauxite.

X-ray analysis by the writer has identified in the veins a few percent of anhydrite in addition to gypsum. It is obviously a co-product of gypsum, resulting from pyrite oxidation.

Barite, $BaSO_4$. A little barite was found by Kiss (1955) in the top parts of the Gánt deposits, Hungary, and in the clay immediately overlying it. Its tabular crystals are translucent and sometimes pale blue, which implies a content of strontium. Kiss ascribes it together with alunite and gypsum to the solutions containing sulphuric acid which result from pyrite oxidation. Barium was probably dissolved out of feldspar grains by the sulphatic solutions. It was soon precipitated in its downward travel: this is why it is absent from the middle and lower parts of the deposit. Vörös (1958) found a few barite grains of 300 to 1000 μm in size in the Iszkaszentgyörgy–Bitó deposit. He holds them to be epigene, precipitated out from downward-seeping solutions of sulphuric acid.

Coelestite, $SrSO_4$. Identified by X-ray diffractometry in an abundance of 1 to 2% in an alunitic bauxite sample from Eplény, Hungary by Bárdossy (1968a), it is presumably a co-product of alunite, a result of the epigene oxidation of pyrite.

Chalcanthite, $CuSO_4 \cdot 5H_2O$. It is mentioned without any detailed characterization by Beneslavsky (1963) from some bauxites in the Northern Urals, USSR.

Svanbergite, $SrAl_3(OH)_6(SO_4)(PO_4)$. It was identified in the USSR in the deposits of Severokolchedansk, Sokolovsk and Barzassk (Beneslavsky 1958) and in the Odinsk deposits by Lukanina (1959). In the latter, it forms pore and fissure fillings. Some of the grains are cryptocrystalline, others are macrocrystalline; all are presumably of epigene or supergene origin.

Sokolovite, $(Ca, Sr)Al_3(OH)_6(SO_4)(PO_4)$. Intermediate between woodhousite and svanbergite, it was identified by Gladkovsky and Sharova (1960) in the Kushmurunsk deposits of the Central Urals, USSR.

Tikhvinite, mentioned by Beneslavsky (1963) from the bauxites of the Northern Urals, is no independent mineral; it is a variety of svanbergite.

5.3.6. PHOSPHATES, VANADATES AND ARSENATES

Apatite, $Ca_5(F, Cl, OH)(PO_4)_3$. One of the most frequent phosphate minerals of karst bauxites, it is non-clastic in its majority. Slightly worn apatite paths larger than 60 μm were demonstrated by microscopy by Kiss (1955) in Gánt bauxite, Hungary. A little apatite of clastic origin was found in most of the bauxite deposit groups of the USSR by Beneslavsky (1963). Most of it is syngenetic or diagenetic and finely distributed in the bauxite matrix. The writer has identified it in a majority of Hungarian bauxite deposits, in abundances of 0·1 to 0·8%. According to Strahl (1971), hydroxyl apatite is frequent in Jamaican bauxite. He derives some of it from the weathering residue of the bedrock limestones, but holds the rest to be newly formed.

Epigene apatite is comparatively rarer. It tends to form pore and cavity fillings, as observed e.g. by Beneslavsky (1963) in the Berezovsk deposits, USSR. Todorova and Stefanov (1973) found pale yellow tabular and lenticular nests of 10 cm diameter at

most in the small Paramun open pit near Tran in Bulgaria. These wore a haematito–goethitic crust a few millimetres thick. In two samples received by courtesy of those authors, the present writer established by X-ray diffractometry the nests to consist predominantly of fluoroapatite. The nest-like appearance of the apatite segregations suggests an epigene origin.

Carbonate apatite (francolite), $Ca_5[F(PO_4)(CO_3)(OH_3)]$. It was identified in the form of minute automorphic grains, pore and vacuole fillings, as a secondary formation in the Devonian bauxites of the Northern Urals, USSR (Beneslavsky 1963). According to Bushinsky (1971), bauxite is strongly phosphatised near the bedrock of one of the deposits there, containing 23% P_2O_5 on average in the form of francolite.

Evansite, $Al_3(OH)_6(PO_4) \cdot 6H_2O$. Found together with francolite in the Devonian deposits of the Northern Urals, USSR, it is a secondary formation there, filling vacuoles and cavities, or constituting reniform aggregates (Beneslavsky 1963).

Fermorite, (Ca, Sr) $F(PO_4)$. It has been identified by the author using the electron microprobe in the bauxite of Iszkaszentgyörgy–Kincses, Hungary, where it forms round and oval segregations of 10 to 20 μm in size in iron-rich roundgrains and bauxite pebbles. Some of the calcium is substituted by cerium; the mineral even contains a minute amount of uranium. The writer regards it as diagenetic on the strength of its appearance.

Augelite, $Al_2(OH)_3(PO_4)$. Demonstrated by Bárdossy and Mack (1967) in abundances of 0·4 to 0·8% in the Malm deposits of the Parnassos and Kiona ranges, Greece, finely dispersed in the bauxite matrix, it is held to be syngenetic. It was identified by them also in the Turonian bauxite horizon of the same region, but it is not ubiquitous there.

Augelite was first identified by Slansky *et al* (1964) in the upper, secondarily lateritised part of a phosphate deposit in the Senegal. According to them, lateritisation transformed the original fluoroapatite into augélite and crandallite. On the basis of this finding, the present writer holds it probable for augélite to be present also in other bauxite deposits: its small abundace makes its identification very difficult, however.

Crandallite, $CaAl_3H(OH)_6(PO_4)_2$. After apatite, this is the most important phosphate mineral of karst bauxite. The writer's X-ray diffractometric analyses have proved it to be present in all Hungarian deposit groups, in abundances of 0·1 to 1·0%. Its bulk tends to be finely disperse in the bauxite matrix; there is very little of it, if any, in the ooids and pisoids. The writer has found in places 1 to 4% crandallite in the multicoloured bauxites lying below the cover in the Iszkaszentgyörgy deposits. It was most abundant in the light spots of epigene deferrification.

Nagyegyháza bauxite contains 1 to 5% crandallite, partly finely disperse in the matrix, and partly in the form of yellowish veins 0·1 to 1·0 mm wide, entirely filled with crandallite. Crystal size in the veins is 10 to 20 μm. The veins traverse the textural elements of the bauxite (Gecse 1975, oral communication). In the Hungarian bauxite deposit groups, then, crandallite is present in syn- and diagenetic as well as in epigene form.

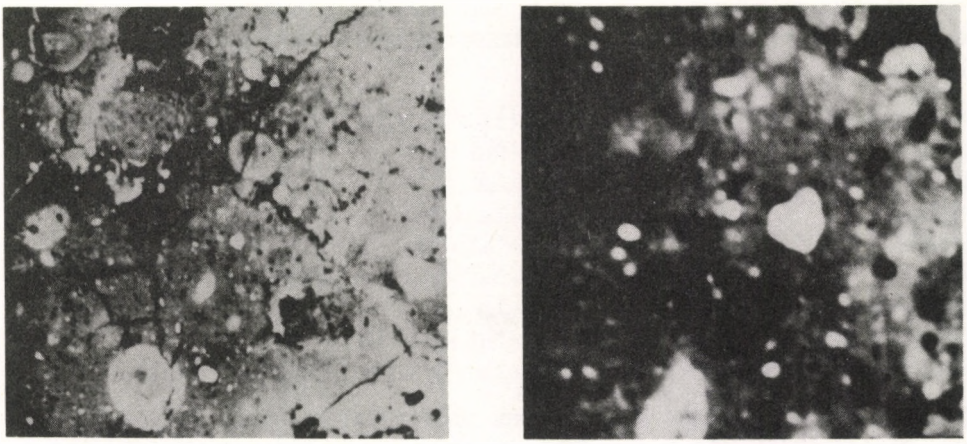


Figure 152. A worn, clastic monazite grain in a bauxite. Iszkaszentgyörgy, Kincses ore body, Hungary. Electron microprobe, electron images
(A) 300×; (B) 1200×

Anderson (1971) states it to be the principal phosphate mineral of Jamaican bauxite; its typical abundance is in the 1 to 1·5% range there. In some valley-bottom deposits, however, it attains 3 to 4%. Crandallite is most abundant in the karst bauxites of the Pacific islands. The writer has found e.g. 1 to 13% crandallite in samples received from the Loyalty Group, and 2 to 8% in samples from Niué Island, East Tonga Group. The phosphatic bauxite described from Lau Island presumably is also of crandallitic constitution. Finally, the writer found 3 to 5% crandallite in a bauxite sample from Rennell Island, a gift of de Weisse. This is in a good agreement with the finding of de Weisse (1970) who states the oolitic phosphate deposit on nearby Bellona Island to have crandallite for its principal mineral.

Monazite, $CePO_4$. It has been identified in many deposit groups by the electron microprobe investigations of Bárdossy and Pantó, in abundances ranging from a thousandth to a hundredth of a percent. Monazite grains are frequent in Gánt, Iszkaszentgyörgy and Nagyharsány bauxite, Hungary; they are seldom observed in Kislőd bauxite, and sporadic in Alsópere bauxite. Typical grain size is 1 to 10 μm; exceptionally, the grains may attain 10 to 20 μm in size. On the basis of their more or less rounded outlines, they are held to be clastic (fig. 152).

Bárdossy and Pantó (1973) could also demonstrate that monazite grains of clastic origin existed in the Provence deposits of Southern France. They found the number of monazite grains and their modal and maximum grain size to decrease gradually from west to east. This distribution is held to reflect the distance from the parent rock and the length of transport. In Italy, Bárdossy *et al* (1977) found comparatively large numbers of monazite grains in the deposits of Campo Felice in the Abruzzi and of San Giovanni Rotondo in Apulia. In both places, the grains were of 2 to 8 μm in size and of a clastic appearance.

Most monazite grains contain other rare earths elements as well as cerium. The most frequent successions of decreasing abundance observed in the deposits studied were as follows:

Nagyharsány	Ce > Nd > La > Pr > Sm > Dy
Gánt	Ce > La > Nd
Iszkaszentgyörgy	Ce > Nd > La
Alsópere	Ce > Nd > La > Pr
Kislőd	Ce > La > Nd > Th
Allauch	Ce > La > Nd > Pr > Sm > Ho > Dy > Pm > Er
Mazaugues	Ce > Nd > La
Campo Felice	Ce > La > Nd > Pr > Th
San Giovanni Rotondo	Ce > Nd > La > Pr > Sm > Gd > Ho > Th

In all the deposits groups studied, then, neodymium and lanthane are the most abundant rare earths after cerium. The others are not ubiquitous, and their abundances are also fluctuating.

It is probable that some monazite is present in many other deposits: it is only for the lack of adequate investigations that its presence has not so far been detected.

Cheralite, (Ce, Ca, La, Th)(PO_4). Much rarer than monazite, it could be identified only in Italian bauxites by Bárdossy *et al* (1977) in some samples from Campo Felice and San Giovanni Rotondo. The grains are of 2 to 8 μm in size and of clastic origin.

Xenotime, YPO_4. It was identified by the electron microprobe in quantities from a thousandth to a hundredth of a percent by Bárdossy and Pantó in several bauxites. In the Hungarian deposits, most particles are of 1 to 10 μm in size, rounded, of clastic origin. It was only in Nagyegyháza bauxite that a few grains in the 10 to 20 μm size range were discovered (fig. 153). Xenotime grains were found to be comparatively abundant in the Gánt–Harasztos and Gánt–Újfeltárás deposits. They are less frequent at Kislőd, sporadic at Iszkaszentgyörgy, Nézsa and Nagyharsány.

In the deposits of the Provence, Southern France, the number, the modal and greatest size of the grains increase from west to east, opposite to monazite. In the easternmost deposit, Pas de Recoux, the largest grains were of 22 by 30 μm in size (Bárdossy and Pantó 1973). In Italy, in the deposits examined by those authors, xenotime is less abundant than monazite: it was identified only in a few samples from Campo Felice and San Giovanni Rotondo. The grains are of 2 to 10 μm in size, rounded, of a clastic appearance.

The xenotime grains contained numerous other rare earths elements in addition to yttrium. Their decreasing order of abundance was the following in the deposits studied.

Nagyharsány	Y > Dy > Er > Gd > Ho > Yb
Gánt	Y > Dy > Er > Ho > Yb > Gd > Tb > Tm > Hf > Lu
Nézsa	Y > Dy > Yb > Er > Ho
Allauch	Y > Dy > Sm > Gd > Yb > Ho > Er > Eu > Nd
Mazaugues	Y > Dy > Gd > Ho > Er > Yb > Tb > Lu > Sm
Pas de Recoux	Y > Dy > Er > Yb > Ho > Lu > Gd

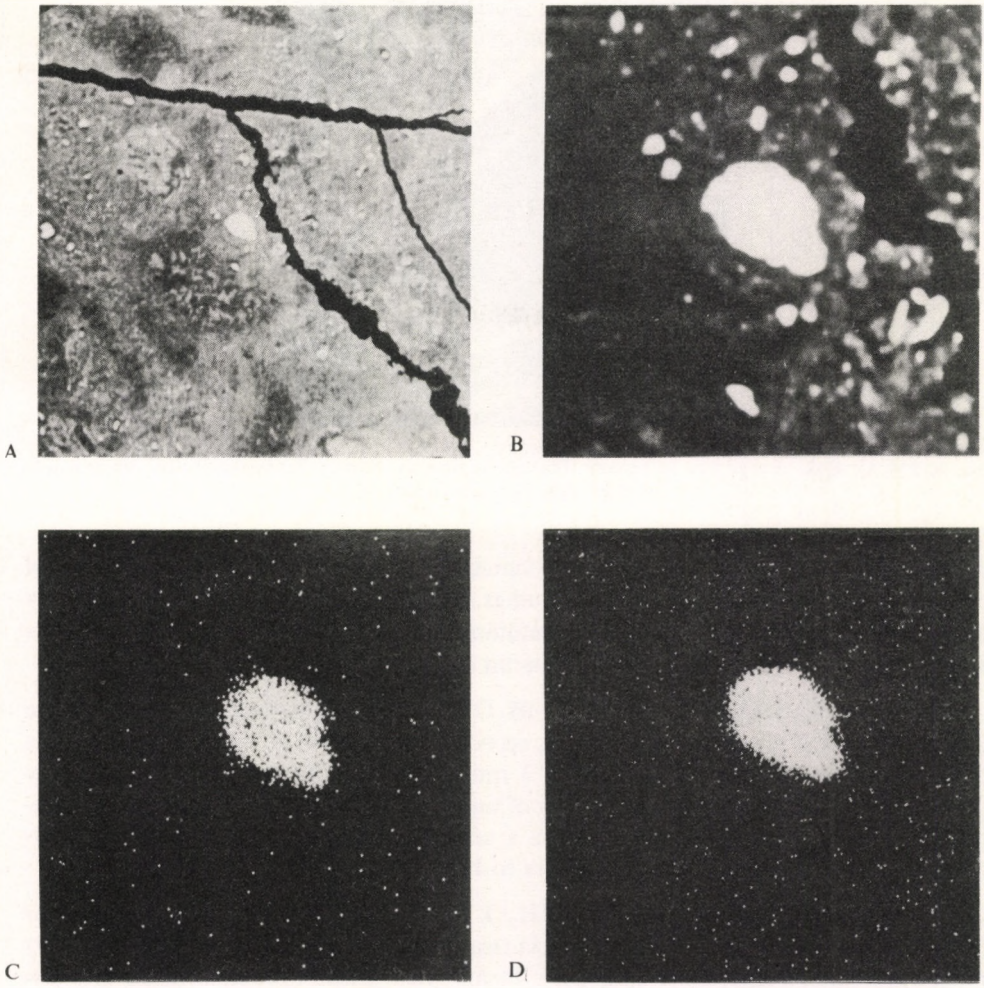


Figure 153. A worn, clastic xenotime grain in a bauxite. Nagyharsány, Hungary. Electron microprobe images (A) electron image, 300 × ; (B) *idem*, 1200; (C) X-ray image of yttrium distribution, 1200 × ; (D) X-ray image of phosphorus distribution, 1200 ×

In all the bauxites examined, then, dysprosium is the most frequent element after yttrium. The abundances of the others fluctuate from deposit to deposit.

Tkachenko and Bol'shun (1971) found a few clastic xenotime grains by microscopy in the Triassic bauxites of the Northern Urals, USSR. They did not describe the size or the distribution of the grains. According to Sharova and Glazunova (1966), some bauxites of Vietnam also contain a very few clastic grains of xenotime.

Metatujamunite, $Ca(UO_2)_2(V_2O_8) \cdot 4H_2O$. Some vivid yellow nests of peanut size in a reddish-brown bauxite were found by Köhler in 1955, in the deposit called Gräser of the Unterlaussa group in Austria. That author proved them by chemical methods to be

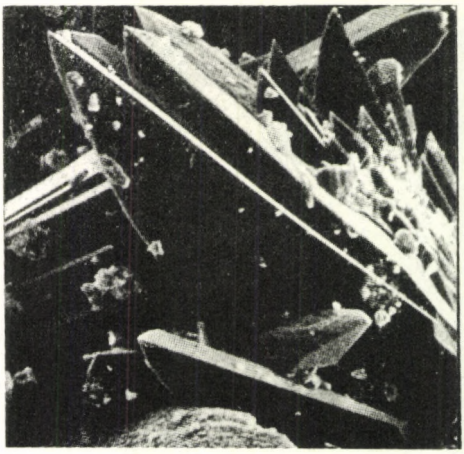


Figure 154. Crystals of erythrine in bauxite. Blanquette deposit, France. Scanning electron micrograph, 400×. After Dietrich *et al* (1976)

significantly enriched in uranium and vanadium, and assumed them to consist of carnotite on the strength of the optical studies. Brauner and Grögler (1957) using X-ray methods proved the mineral to be metatujamunite in fact, rather than carnotite. Its appearance in the form of nests suggests an epigene origin.

Tujamunite, $Ca(UO_2)_2(V_2O_8) \cdot (5—8)H_2O$. In a bauxite sample likewise from the Gräser deposit of Unterlaussa, Austria, received courtesy of W. Siegl, the writer has observed yellow segregations of 2 to 3 mm size which he has proved by X-ray diffractometry to consist predominantly of tujamunite accompanied by less metatujamunite. Metatujamunite presumably is a secondary dehydration product of tujamunite. The writer holds the tujamunite to be epigene.

Metazeunerite, $Cu(UO_2)_2(AsO_4)_2 \cdot 8H_2O$. It was identified by Brauner and Grögler (1957) in the Gräser deposit of Unterlaussa, Austria, where it is accompanied by metatujamunite. Presumably epigene, it forms grass-green, tabular crystals of a few millimetres in size.

Erythrite, $CO_3(AsO_4)_2 \cdot 8H_2O$. It was identified by Dietrich *et al* (1976) in the Blanquette deposit near Brignoles, France, where it forms well-developed crystals of a few tenths of a millimetre size in the bottom part of the bauxite (fig. 154). It has been observed also in the form of spherulites of less than 2 mm in size. Presumably secondary.

5.3.7. NATIVE ELEMENTS

Native sulphur, S. Kormos (1930) described a grey pyritic bauxite from Istria, Yugoslavia in which 1·5% sulphur was left over after subtracting from the analytical result of sulphide sulphur the amount needed for pyrite. Kormos surmised this excess to be in the form of native sulphur. Caillère and Pobeguin (1963, 1965) identified native

sulphur in an abundance of 1 to 5% in grey, pyritic clayey bauxite from the Durban deposit, Ariège, France, where it is finely disperse in the bauxite matrix.

Native sulphur is a product of the diagenetic reduction of bauxite. Sulphur-separating bacteria presumably played an important role in its formation.

Native copper, Cu. In the Rohrer open pit in the Carboniferous deposit group of Missouri State, US, in a bauxitic clay of ooidic texture, Allen (1955) found segregations of native copper about one centimeter thick in leaf-shaped forms, surrounded by a black carbonaceous bauxitic clay. According to Allen, the strongly reducing medium implied by the black carbonaceous lumps may have played an important role in the segregation of native copper. It seems a diagenetic or epigene formation.

Native gold, Au. Kiss (1955) identified native gold grains of clastic origin, of 90 to 100 μm in size and worn outline, in some bauxite samples from Gánt, Hungary. Their abundance is very small, probably less than a thousandth of a percent.

5.3.8. ORGANIC MINERALS

Chemical analyses indicate some hundredths or tenths of a percent of organic matter in most bauxites. Very little is known about the mineralogical forms taken by this organic matter.

Vörös (1958) found a few organic mineral grains in each of his samples in the course of his micromineralogical examination of one profile of Bitó open pit and of József II adit, Iszkaszentgyörgy, Hungary. These grains were of 40 to 1500 μm in size, yellowish-brown, translucent. In the flame, they burnt up, leaving behind some ash. They did not dissolve in water, hydrochloric acid, motor spirit or xylene. Their refractory index is about 1·5. Vörös assumed them to be some derivative of bituminite, and tentatively identified them with *succinite*.

5.4. INDIVIDUALISM AND HETEROGENEITY OF BAUXITE MINERALS

It was the refinement in mineralogical methods which led to the recognition that the real crystal structures of the minerals found in bauxites differ more or less from the ideal structures inferred from theory. These deviations give rise to some individual mineralogical features. This phenomenon was named *mineral individualism*, and quantified in terms of a variety of X-ray, infra-red, and physico-chemical parameters, by Bárdossy *et al* (1976). The measurements of the half-amplitude widths of the X-ray peaks, the X-ray constants referred to those of boehmite, and the infra-red specific extinction turned out to be especially useful. The authors found the parameters of kaolinite to be the most diverse, followed by haematite and goethite, then gibbsite and boehmite, and finally diaspore and anatase. Substantial differences were also observed in the solubility activity of the iron minerals.

The above parameters tend to vary within rather narrow limits in one and the same deposit group: it is only the parameters of the minerals of secondary formation—nests,

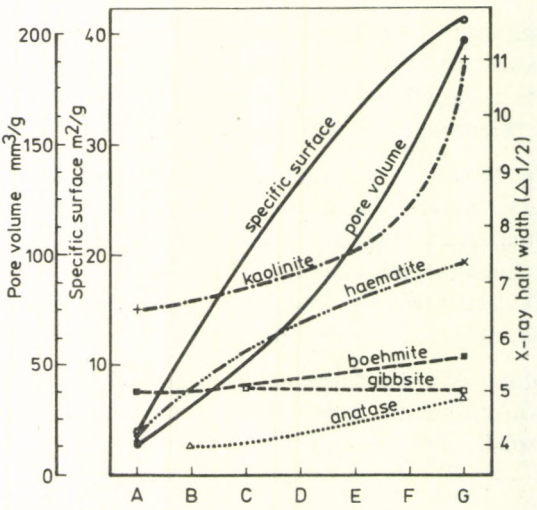


Figure 155. Changes in mineral individualism, specific surface, and pore volume of bauxite in a sequence of increasing tectonic dislocation. Locations and degree of tectonic dislocation are

(A) Chalkidike, slight thermal metamorphism plus strong tectonic disturbance; (B) Campo Felice, strong; (C) Mazagues, medium; (D) Malomvölgy, slight; (E) Halimba III mine, slight; (F) Spinazzola, slight; (G) Haiti, no tectonic disturbance (after Bárdossy *et al* 1977)

fissure fillings etc.—that differ essentially from those of the minerals in the main mass of the bauxite.

The relationship between the said parameters and the geological age of the bauxite deposit is a very vague one. The minerals of the older deposits tend in a general way to be more crystalline than those of the younger ones, but there are numerous exceptions to this rule.

The authors cited have, on the other hand, established a very close connection between the crystallinity of the bauxite minerals and the degree of structural deformation of the deposits. The crystallinity and crystal size of minerals increase unequivocally with increasing structural deformation. Some selected examples are presented in figure 155. The half-height widths of the peaks of kaolinite and haematite decrease markedly with increasing tectonic deformation of the deposits; those of boehmite and anatase decrease much less. Remarkably enough, the crystallinity of gibbsite remains unchanged. Beyond a certain degree of structural deformation, however, the mineral disappears by conversion into boehmite or diaspore. The specific extinction parameter exhibits a similar trend. Figure 155 further reveals that specific surface and microporosity (pore volume constituted by pores of less than 100 μm in size) also decrease monotonically with increasing structural deformation. This implies both an increase in crystal size and a closer packing.

The authors cited also studied some deposits unburied since their origin and unexposed to any structural stress. The minerals in the karst bauxite deposits of this type were found to be very small (0·05 to 0·3 μm), and their crystallinity was found to be very far from the ideal. In some lateritic bauxite deposits, on the other hand, they observed an extremely strong secondary recrystallisation, which had greatly increased the grain size and crystallinity of the minerals. This is illustrated by figure 156, in which a karst bauxite from Haiti is compared with a laterite bauxite sample from Weipa,

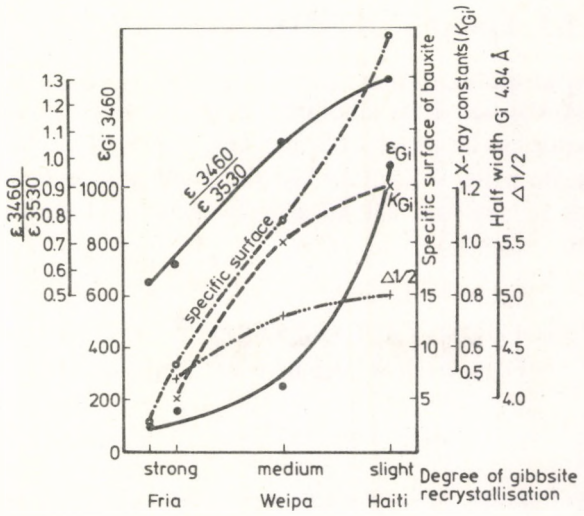


Figure 156. Changes of gibbsite individualism in surface deposits due to secondary recrystallisation. After Bárdossy *et al* 1977

Australia (of moderate recrystallisation) and two samples (of strong recrystallisation) from Fria, Guinea. The samples' gibbsite contents were as follows: 43·9% in the Haiti sample, 77·8% in the Weipa sample, and 82·7 and 87·7% respectively in the two samples from Fria. The parameters of gibbsite are seen to have shifted quite substantially during recrystallisation. The conditions for crystallisation on the surface are more favourable in the laterite than in the karst bauxite deposits, where the limestone environment exhibits a strong inhibiting effect.

Mineral individualism according to the authors' measurements induces significant differences also in the behaviour of the bauxite minerals during alumina processing, which makes a detailed investigation of this feature a task of practical importance.

The reasons for mineral individualism are thought to be connected with differences in the processes and circumstances of mineralization. The real structures and individual properties of the minerals of primary formation may be modified by the secondary processes. Experience collected so far implies the secondary processes to act as a rule in the direction of an improved crystallinity and larger crystal size.

By means of selective dissolution tests and repeated centrifuging, Bárdossy *et al* (1976) could prove that there may be several bauxite-mineral varieties side by side in one and the same bauxite. This phenomenon was originally given the name mineral heteromorphism, but the term *mineral heterogeneity* seems to be more to the point.

The reason for mineral heterogeneity in bauxite may be some small changes over time in the conditions of formation, resulting in temporal successions of differing mineral varieties. These varieties seem to be sufficiently stable to persist side by side over time span fairly long even in the geohistorical sense of the term. Increasing pressure and temperature tend to reduce the heterogeneity of the mineral associations: the minerals of the metamorphic bauxites tend to be predominantly homogeneous as to crystal structure and constitution.

5.5. ISSUES OF MINERAL GENETICS

In this chapter, we outline at first the conditions under which the principal bauxite minerals can be synthesised in the laboratory or transformed into other minerals. As a next step, the theories so far proposed concerning the origin of the minerals are to be reviewed. Finally, the writer will outline his own genetic system, developed on the basis of all the results of analyses, statistical processings and laboratory experiments accessible to him.

5.5.1. CONDITIONS OF LABORATORY SYNTHESIS OF THE ROCK-FORMING BAUXITE MINERALS

5.5.1.1. THE SYSTEM Al_2O_3–H_2O

In the last two decades, many authors have been concerned with constructing a phase diagram for this system. The phase diagrams of Ervin and Osborn (1951), Kennedy (1959) and Torkar and Krischner (1964) are shown in figure 157. Although their results conflict in many details, they agree in that the varieties of $Al(OH)_3$ are stable below about 90 to 120 °C. Torkar and Krischner state that, in a pure Al_2O_3–H_2O system, bayerite is invariably formed; gibbsite is formed only in the presence of Na^+ or some other alien ions. These authors accordingly regard bayerite as the stable modification of $Al(OH)_3$, and consider gibbsite and nordstrandite to be varieties stabilized by alien ions. Their opinion is confirmed by Ginsberg's (1962) experiments who found that the gibbsite lattice can accommodate 0·1 to 1·0% alkalies, and that the alkali content stabilizes the lattice.

According to Schoen and Roberson's experiments (1970) only gibbsite can form at room temperature at a pH less than 5·8. In a neutral or weakly alkaline medium, bayerite is the stable modification; metastable gibbsite may also form, however. A very slow ageing of Al-hydroxide gel—over a year or a year and a half—permits bayerite to recrystallise into nordstrandite if the medium is alkaline. Nordstrandite is held to be the stable modification in an alkaline environment by Schoen and Roberson.

Above the trihydrates range of stability, there is the stability range of the almonohydrates. According to both Ervin and Osborn and Torkar and Krischner, boehmite is the stable phase below about 150 bar and diaspore at all pressures higher than that. Kennedy on the other hand found diaspore to be stable over the entire pressure range and regards boehmite as a metastable phase everywhere. Its rate of transition into diaspore, however, isvery slow; the phase diagrams thus do not in fact express true equilibrium conditions. This conjecture was confirmed by the experiments of Neuhaus and Heide (1965), who could substantially accelerate the process by seeding with diaspore crystallites, and could in that way procedure diaspore all over the boehmite range. Above some 380 °C, the stability range of the dehydrated Al_2O_3 modifications follows, α-Al_2O_3 (corundum) being the stable final product.

In the geological respect, the tests performed at Earth-surface temperature and pressure are of considerable importance. According to Papée *et al* (1958), amorphous

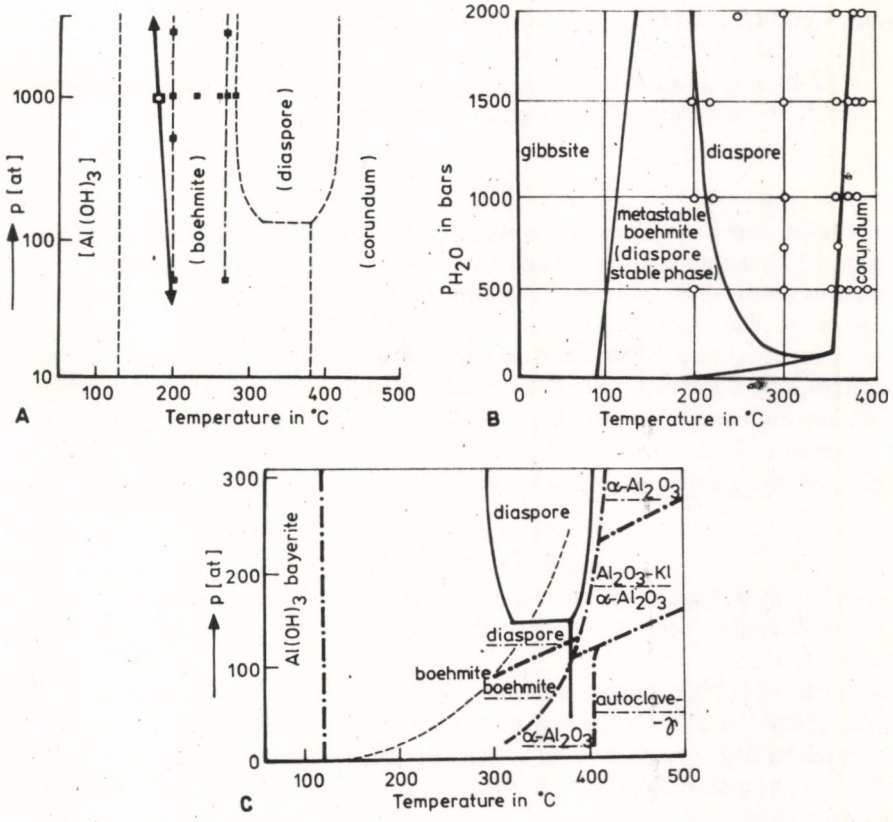


Figure 157. Phase diagrams of the Al_2O_3–H_2O system

(A) according to Ervin and Osborn (1951); (B) according to Kennedy (1959); (C) according to Torkar and Krischner (1964)

aluminium hydroxide precipitated in a weakly alkaline medium crystallises into beohmite if the medium is maintained alkaline throughout. Pa Ho Hsu and Bates (1964) precipitated aluminium hydroxide out of aluminium sulphate and chloride solutions by modifying the mole ratio $NaOH/Al^{3+}$. The phase crystallising out of the precipitate was gibbsite, bayerite or nordstrandite or a mixture of them at a mole ratio of 3 to 3·3 (at a pH of 8 to 10); on the other hand, at a mole ratio in the 2·1 to 3·0 range (between a pH of 4 and 7), an X-ray-amorphous $Al(OH)_3$ precipitate was formed, which remained even after six months of ageing.

Schellmann (1964a) obtained an amorphous aliminium hydroxide gel on adding potassium hydroxide to a solution of $AlCl_3$. He let it age for seven months at a pH = 7·5, and kept it at 80 °C for the last four days, during which it crystallised into gibbsite. The same process of ageing at the same pH gave rise to boehmite in the presence of the complex anion HCO_3. The great importance of this experiment is in that it provides direct proof of the mineralising influence—preferential for boehmite—of a carbonate environment. The presence of Mg ions does not affect significantly the equilibrium conditions of the phases (Roy *et al* 1953). The same holds also for CaO (Peppler and Wells 1954).

Garrels and Christ (1965) studied the dissolution stability of gibbsite by means of thermodynamic calculations. If it is accepted that gibbsite dissolves in ionic form in an acidic medium and as aluminate ions in an alkaline one, then it must be stable at a pH of 5·1 where its minimum of dissolution activity lies.

5.5.1.2. THE SYSTEM Fe_2O_3–H_2O

The equilibrium conditions of the phases of this system have in the last decade been investigated by many workers. Their findings are contradictory in several respects, presumably owing to the fact that, at room temperature, no equilibrium sets in even in several years. The most important in this respect are the investigations of Wefers (1966), who found that X-ray-amorphous ferrihydroxide may crystallise into goethite or haematite or a mixture of the two, with goethite being the preferred phase in an alkaline medium and haematite in an acidic one. Above 70 °C, goethite recrystallises into haematite. This reversible reaction involves a very slight energy change: the enthalpy of the reaction ($\Delta_x H_r$) is a mere 4 kcal for the reaction equation

$$\underset{\text{goethite}}{2FeOOH} \rightleftarrows \underset{\text{haematite}}{Fe_2O_3} + H_2O.$$

According to Wefers, this is the reason why the two minerals so often occur together in bauxite. (Lepidocrocite is a metastable modification: it recrystallises into goethite above 25 °C.)

Schellmann's (1959) findings were similar: according to him, the presence of alien ions also affects the crystallisation of amorphous ferric gel. The formation of haematite is promoted by Mg^{2+} and Ca^{2+} ions; that of goethite, by the complex anions HCO^{3-} and SO_4^{2-}. The presence of K and Na ions, on the other hand, does not affect the equilibrium either way.

5.5.1.3. THE SYSTEM Al_2O_3–Fe_2O_3–H_2O

As described in the mineralogical chapter, the Al^{3+} and Fe^{3+} ions may replace each other isomorphously within limits in certain Al and Fe minerals. In view of the importance of the issue for alumina plant technology, Thiel (1963) and Wefers (1967) performed detailed experiments into the equilibrium conditions of this system. The findings of the latter author are presented as a phase diagram in figure 158.

Wefers found that, below one kilobar phase boundaries are only shifted insignificantly by pressure changes. The diagram accordingly refers to conditions corresponding to the isothermal vapour pressure of the saturated solution. At room temperature, bayerite is the stable aluminium mineral in a pure Al_2O_3–Fe_2O_3–H_2O system. The iron mineral in equilibrium with bayerite is alumogoethite. Neither bayerite nor gibbsite forms any mixed crystals containing ferric ions. The goethite–haematite transition takes place at 50 °C, that is, at a temperature lower by 20 °C than in the pure system Fe_2O_3–H_2O. Wefers attributes this phenomenon to the isomorphous insertion of aluminium.

Between 50 °C and about 90 to 100 °C, then, alumohaematite and bayerite are phases in equilibrium. At higher temperatures, bayerite is replaced by diaspore. The presence of diaspore at so low a temperature is made possible by the epitactic

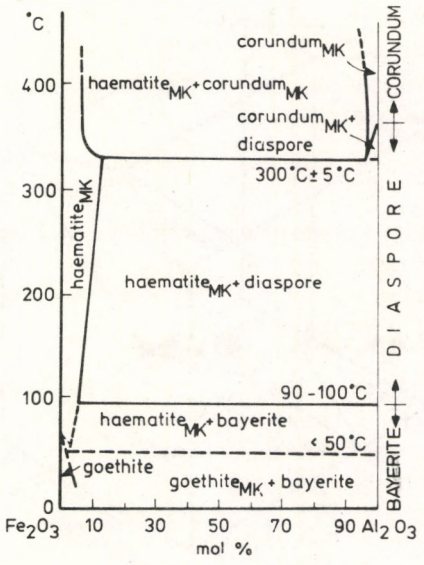


Figure 158. Phase diagram of the system Al_2O_3–Fe_2O_3–H_2O. After Wefers (1967)

nucleating influence of minute crystallites of goethite. The anionic lattice of goethite differs so little from that of diaspore that diaspore can keep growing on goethite seeds even below its range of thermodynamic stability. According to Wefers, diaspore may form in nature even at temperatures lower than that if more time is available (a time span so long as to be impossible to reproduce in the laboratory). The most important preconditions of diaspore formation in Nature (on the surface), then, are conditions ensuring nucleation and a fair degree of solubility.

Above 330 °C, the phases in equilibrium are alumohaematite and iron-bearing corundum. This temperature is also much lower than that observed in the pure system Al_2O_3–H_2O. According to Wefers, the formation of both alumogoethite and alumohaematite is promoted by a high concentration of iron and a comparatively low one of aluminium. On increasing the concentration of aluminium, mixed crystal formation is replaced by orientated intergrowths of pure Al and Fe minerals.

5.5.1.4. THE SYSTEM Al_2O_3–SiO_2–H_2O

In the course of clay minerals research in recent years, this system has been investigated a great deal. Our concern here is primarily with the behaviour of kaolinite, the phase of the system which is most abundant in karst bauxites. We shall accordingly concentrate here on the experiments of Garrels and Christ (1965), in which those authors determined at room temperature and atmospheric pressure the stability ranges of kaolinite and gibbsite as functions of aluminium and silica concentration and of the pH (fig. 159).

The most important fact is that, as the concentration of H_4SiO_4 increases, kaolinite is replaced by gibbsite at a constant weakly acidic pH. If, then, dissolution takes place

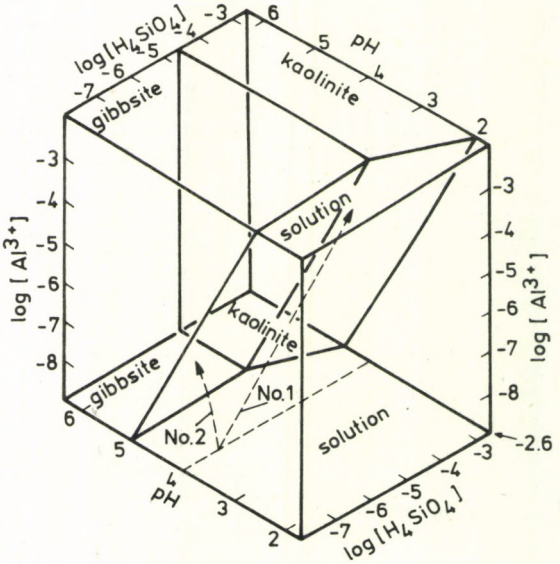


Figure 159. Stability ranges of kaolinite and gibbsite as functions of silica concentration and pH, at 25 °C and atmospheric pressure. After Garrels and Christ (1965)

in a buffered acidic medium, the composition of the solution is displaced in the direction of the arrow marked 1, towards the stability range of kaolinite. This means that kaolinite is dissolved congruently, the substance left behind being also kaolinite. Such a buffering effect may be exerted in nature e.g. by the humic acids resulting from the decomposition of a tropical vegetation. If, on the other hand, this buffering effect is weak or absent, then the composition of the solution is displaced as indicated by the arrow marked 2, towards the stability field of gibbsite. In that case, the dissolution of kaolinite is incongruent, and what is left behind is gibbsite.

In the writer's opinion, *micro-organisms* may also play a decisive role in deciding whether the first or the second process gains the upper hand, as their activity may substantially influence the pH, and accelerate or retard the oxidation of the decomposing organic matter. Gardner (1970) on a thermodynamic basis (utilising free energies of formation) gave a theory for the dissolution conditions of kaolinite as a function of the above three factors. His findings, plotted in figure 160, are in a fair agreement with the data of Garrels and Christ, but afford a better insight into the role played by the concentration of Si in the solution. The dissolution of kaolinite is congruent at any pH less than 3·9. At any pH greater than that, the dissolution residue may be gibbsite only if the solution contains less dissolved Si than $10^{-4.6}$ mole per litre. If the aluminium hydroxide is amorphous, the limiting pH is 5·6 and the maximum admissible Si concentration of the solution is even less, $10^{-6.3}$ mole per litre at most. Natural rainwater satisfies all these conditions, and so pH is not in itself a decisive factor!

If the system contains also quartz in addition to kaolinite, then, according to Gardner's calculations, all the quartz must get dissolved first at a pH greater than 3·8. It

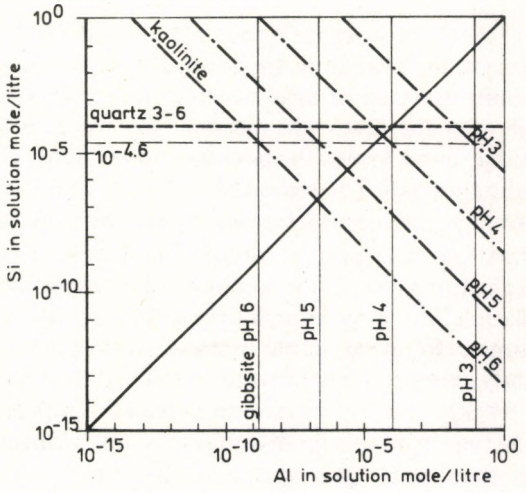


Figure 160. Solubility of kaolinite as a function of pH and concentration. After Garrels (1970)

is only thereafter that the incongruent dissolution of kaolinite can begin to leave behind a gibbsite phase. In any medium more acidic than that, the kaolinite gets dissolved first, except if the quartz grains wear films of clay minerals, which retards their direct dissolution.

The above experimental findings were confirmed and further refined by the thermodynamic calculations of Fritz and Tardy (1973) who found that the stability ranges of gibbsite, kaolinite and quartz are also substantially affected by these phases' degree of crystallinity. For instance, the more crystalline the kaolinite, the broader its range of congruent dissolution.

Kaolinite was synthesised by de Kimpe *et al* (1964) at surface temperature and pressure, as an experimental proof of the possibility of kaolin formation under Earth-surface conditions.

5.5.1.5. TITANIUM OXIDES

The stability ranges of titanium oxides were investigated by Kessmann (1966) who found anatase to be the stable phase at low temperatures and in the presence of low alkali concentrations. Rutile does not become stable below 400 °C, nor does brookite until the alkali concentration becomes higher. This is a plausible explanation for the predominance of anatase in karst bauxites.

5.5.1.6. MODEL EXPERIMENTS

Of the wide range of model experiments performed in the last decades, let us discuss here only those one concerned with kaolinite decomposition, in view of the fact that many eminent scientists—e.g. Erhart (1956)—refuse to accept the possibility of kaolinite getting transformed into aluminium minerals under Earth-surface conditions.

Pedro and Berrier (1966) let water of about pH = 6 and 65 °C drip continuously through some highly crystalline kaolinite. In the upper part of their sample holder tube, some 15% of the kaolinite was transformed into boehmite after one year of continuous percolation, although after the first eight months only traces of boehmite could be detected. X-ray examinations revealed the boehmite to be highly crystalline, although its grains were exceedingly small. The remarkable thing is that the phase thus formed was not gibbsite, although any considerations of the structure of the lattice would suggest that the formation of a gibbsite lattice would have been simpler after the liberation of the octahedral layer in the kaolinite. The experiment proved that an uninterrupted percolation for a long enough time will in fact decompose the kaolinite lattice, which is assumed to be so stable, even at atmospheric pressure and at a temperature but slightly above room temperature, and that the decomposition product is boehmite. In this respect, the finding is of fundamental importance.

In nature, where the time available for dissolution is much longer, the process can in the writer's opinion take place also in the temperature range of 35 to 50 °C, corresponding to a tropical climate.

Recent experiments by Pedro *et al* (1970) modified the above results in the sense that decomposition will proceed fastest if the kaolinite lattice is disordered (rather than ideally crystalline), and that the conditions most favourable to the crystallisation of boehmite are a neutral or sligthly alkaline medium and the presence of Ca ions.

5.5.2. EARLIER HYPOTHESES CONCERNING THE ORIGIN OF THE ROCK-FORMING BAUXITE MINERALS

The most widely accepted view of earlier decades may aptly be called the *dehydration theory*. It was developed by Harrassowitz (1926), according to whom gibbsite, the primary bauxite mineral, was transformed into boehmite by overburden pressure and the structural strains of folding. Weak metamorphism produces diaspore; stronger metamorphism produces corundum. The succession of transformations, then, is

gibbsite → boehmite → diaspore → corundum.

De Weisse (1948, 1964) held similar views, with the difference that, according to him, gibbsite could form as a primary phase only where a deposit remained unburied for some time, and the climate to which it was exposed resembled the conditions of lateritisation. If, however, the bauxite was inundated immediately on its deposition by a lake or sea, and buried by the sediments there, then boehmite would be the primary phase to form. Gibbsite is transformed into boehmite by overburden pressure. According to de Weisse, diaspore could form out of boehmite and/or gibbsite only at temperatures higher than 370 °C and under an increased pressure. An even stronger metamorphizing influence will convert even diaspore into corundum.

According to A. D. Arkhangelsky (1937) and also Goretsky *et al* (1949), it takes a dynamometamorphic influence for diaspore to form. Vishnyakov (1953) also held the above succession of transformations to be the only possible one, and even stipulated that for the formation of boehmite a weak metamorphism or a protracted epigene transformation would be required. In his opinion, diaspore and corundum are

invariably metamorphic. A similar idea was put forward by Marić (1964b), who assumed the Yugoslav bauxites to be pronouncedly metamorphic, their minerals having been produced under the conditions of a low-temperature hydrothermal process. He considered the Yugoslav bauxites to be a special subfacies of the 'zeolite facies'.

Terentyeva (1958) states the phase first crystallising out of an initial gel to be gibbsite as a rule, boehmite being produced much more seldom. Protracted epigene processes will recrystallise gibbsite into boehmite. This process according to her, takes place more readily in higher-grade bauxite than in types richer in kaolinite. She also attributes the formation of diaspore to dynamic metamorphism or at least to a strong structural stress. Such recrystallisation, too, takes place more readily in pure than in clayey bauxite. This is why in the Devonian deposits of the Northern Urals, USSR, the high-grade types of bauxite tend to be so much richer in diaspore. She gave no justification for the assumption that the presence of clay minerals would retard recrystallisation. Corundum according to her would form under a strong metamorphic influence only. All the minerals thus formed are stable and are not changed further by any supergene processes.

According to Kiskyras (1960a), boehmite in Greece was recrystallised into diaspore only where the deposit was exposed to a strong structural stress. Kiskyras, however, also pointed out that not all the diaspore in bauxite can be explained by the assumption of high temperature and strong pressure, finding as he did that pisoidic bauxites are higher in diaspore as a rule than pelitomorphic ones, quite irrespective of their structural position. He did not attempt to answer the question as to how diaspore could form under surface conditions.

This is where we should cite the opinion of Kiss and Vörös (1965), according to whom gibbsite used to be the predominant primary aluminium mineral in Gánt bauxite, Hungary: part of it recrystallised subsequently into boehmite under the influence of overburden pressure.

The other principal group of views could be subsumed under the heading of '*surface theories*', deriving as they do all the principal bauxite minerals by processes acting under surface temperature and pressure. This theory has its opponents even today; these have a tendency to invoke the above-cited findings of the laboratory experiments according to which boehmite, diaspore and corundum may only form at temperatures and pressures higher than the surface-normal. The adherents of the theory, on the other hand, strive to explain the formation of the different minerals in terms of differences in Earth-surface physico-chemical and paleogeographic conditions.

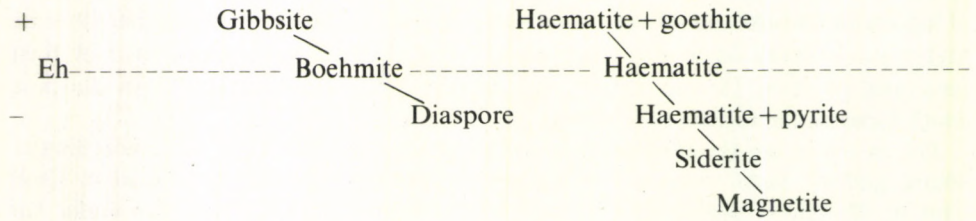
De Lapparent as early as 1935 came to the conviction that gibbsitic bauxite is formed above the ground water table, boehmitic bauxite more or less flush with it, and diasporic bauxite below it. In the same year, Allen by a very thorough microscopic study of the diasporic bauxite deposits of Missouri State, US, proved that the diaspore there must have formed out of kaolinite and illite under surface temperature and pressure. Diasporization according to him could take place wherever the bedrock was karsted strongly enough to permit a sufficiently intense percolation of ground water.

This idea was refined likewise on the basis of findings concerning Missouri bauxite, by Keller *et al* (1954), who explained the formation of diaspore under surface temperature and pressure in terms of the principles of '*structural inheritance*' and

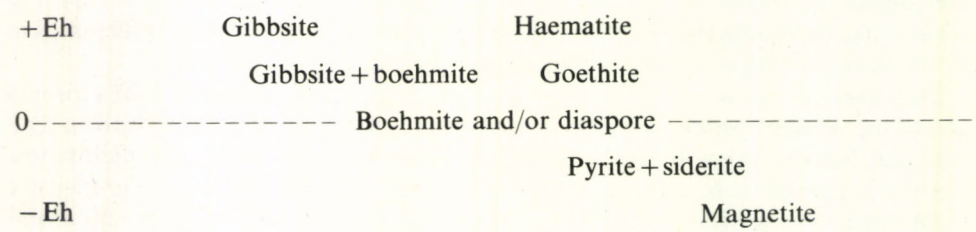
'*energy inheritance*'. If silica is leached very slowly out of kaolinite in a process of hydrolysis or dialysis, then the octahedral layer, instead of collapsing, changes into diaspore, requiring no input of energy. That is, the diaspore inherits, this way, the energy of formation of kaolinite. In Missouri bauxite, this process took place under reducing conditions in a weakly acidic medium. If kaolinite is entirely dissolved, on the other hand, gibbsite or boehmite will form: their crystallisation requiring less energy than diaspore formation. Gibbsite is the preferred phase under oxidising conditions (percolation by fresh ground water), and boehmite under the more reducing conditions of a poorer drainage.

Köster (1956) demonstrated that small amounts of diaspore exist in some laterite profiles of India; this is a direct proof of diaspore formation on the Earth surface, as these laterite profiles have never been buried or exposed to any structural stresses. According to Chang Yuan-Lung (1956) and Schüller (1957), diaspore must have formed under surface temperature and pressure in the Chinese bauxites, as none of the deposits exhibits any evidence of a higher temperature or pressure.

Valeton (1965) in the course other mineralogical studies on French bauxites gained the conviction that the oxidation–reduction potential influences the formation not only of the iron but also of the aluminium minerals. She distinguished as a function of the redox potential the following 'mineral facies':



According to her, the boehmitic and boehmito–gibbsitic bauxites of the Provence were formed as a result of strong ground water percolation, whereas the diasporic bauxites of the Pyrenees were formed under conditions of prolonged saturation with stagnant ground water. A similar relationship was proposed by Combes (1969) for the bauxites of the Ariège:



Combes's statements gain an exceptional importance if the deposits he has in mind are admitted to be the products of *in situ* bauxitisation. According to him, boehmite and diaspore had simultaneously formed there out of kaolinite. In view of haematite being the associated iron mineral, he attributed to the medium an Eh of zero or a very

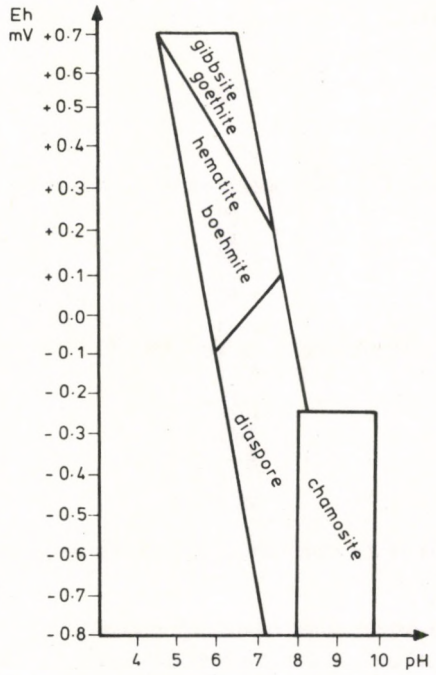


Figure 161. Ranges of formation of aluminium and iron minerals as functions of pH and Eh. After Komlóssy (1976)

slightly oxidising one. He considered the diaspore–pyrite–siderite association posited by Valeton to be incorrect because, in the Péreille deposit, referred to as an exemplary case by Valeton, siderite and pyrite are secondary minerals, the products of the influence of a carbonaceous clay in the immediate cover. Some red bauxite containing the original diasporo–haematitic association can in fact still be found in the lower part of the deposit. In the stability ranges of haematite and goethite, even very slight changes in Eh or pH could trigger the formation of either diaspore or boehmite, or of both at the same time. Under slightly more oxidising conditions, boehmite and gibbsite were formed, pure gibbsitic bauxite required a more strongly oxidising environment.

Similar conclusions were arrived at by Komlóssy on the strength of his mineralogical investigations on Iszkaszentgyörgy bauxite, Hungary (1970) and on the bauxites of Vietnam (1976). He plotted in a diagram the fields of formation of the aluminium and iron minerals as functions of the Eh and the pH (fig. 161). According to him, a gibbsite–goethite association may form under the most strongly oxidising Eh in a weakly acidic medium: boehmite and haematite under a weakly oxidising Eh in an even less acidic medium; diaspore and haematite at an Eh about zero in a neutral medium; and a diaspore–chamosite association in a reducing and slightly alkaline medium.

Nia's (1968) findings concerning the mineral constitution of the Turonian bauxite horizon of Central Greece were described in Section 5.2.4. According to him, the diaspore there is a diagenetic derivative partly of kaolinite and partly of boehmite. The Eh and pH had played no role in this process. Diasporisation took place where ground

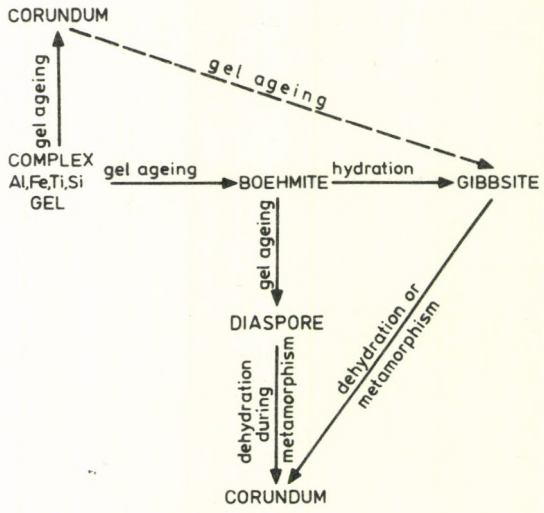


Figure 162. Genetic system of the aluminium minerals, according to Beneslavsky (1963, 1974)

water percolation was most intense. Invoking Wefers's experiments, Nia attributed great importance to the role of alumogoethitic seeds for the diaspore formation. Diasporisation was weaker in more clayey bauxites, owing to their lower permeability.

An adherent of the surface theory in a general way, Beneslavsky nevertheless held views different from the above ones in many particulars, attributing the lion's share to the colloidal properties of bauxite (1952, 1961, 2nd ed. 1974, 1963). According to him, the initial material of bauxite formation was an amorphous ('primary') complex gel of Al, Fe, Ti and Si compounds, in which cryptocrystalline boehmite came to be formed in an overwhelming majority of cases (fig. 162). Depending on the environmental conditions, this could get hydrated into gibbsite or recrystallised into diaspore. According to his diagram, neither boehmite nor diaspore can form out of gibbsite directly, the only direct derivative of gibbsite being corundum. In the primary gel, cryptocrystalline corundum may form by gel ageing under surface conditions. Continued ageing may recrystallise it into gibbsite. Beneslavsky held the main period of mineralisation in bauxite to be the diagenetic phase. The physico-chemical conditions controlling these mineral transformations do not differ much: there presumably were but very slight differences in Eh, pH, solution composition and concentration.

Beneslavsky felt that the phase diagrams established in the laboratory are unsuited to explain all the particulars of the mineral associations found in the bauxites, as some of the processes of mineral formation cannot as yet be adequately reproduced in the laboratory.

Gladkovsky and Ushatinsky (1964) retained the genetic succession gibbsite–boehmite–diaspore–corundum, but assumed that it may have operated in two different ways (fig. 163). Firstly, under surface conditions, in laterite profiles, hand in hand with a depletion of the water content, and secondly, under the influence of metamorphism. The resulting corundum and diaspore are cryptocrystalline in the first case and

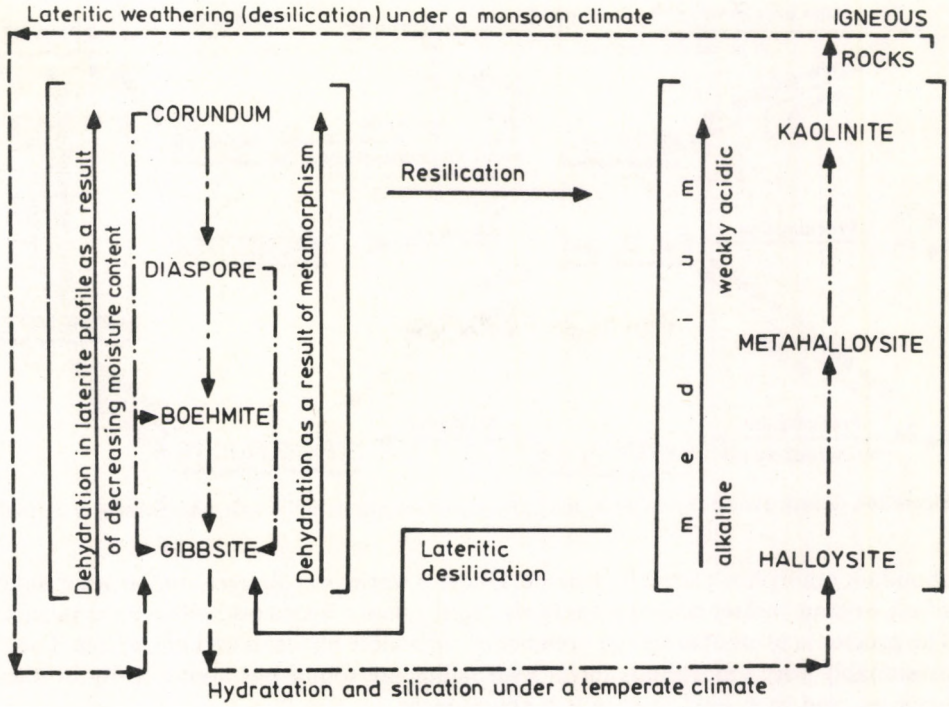


Figure 163. Genetic system of aluminium oxide, aluminium hydroxide and phyllosilicate minerals in the bauxites and laterites, according to Gladkovsky and Ushatinsky (1964)

coarsely crystalline in the second. They assumed that, in the laterite profiles, dehydration may be due to solar energy, but specified no mechanism to account for the process.

They attributed an especially great importance to epi- and supergene changes in the bauxite, which according to them initiated a reverse process, gradually hydrating corundum, diaspore and boehmite into gibbsite. They called this process the 'weathering' of bauxite which, according to them, could take place even under a temperature climate if it was humid enough. The present mineral constitution of a bauxite deposit, then, should be independent of its age (provided the deposit has not been exposed to metamorphism), and should depend only on the composition of the original bauxite and on its supergene 'weathering'.

Bauxite literature is much less concerned with the conditions of origin of the *iron minerals:* it is usually the general findings derived from sedimentary petrography and from the study of iron ores that are referred to. Scheffer *et al* (1957) developed the diagram shown as figure 164, based largely on pedological evidence, to explain the origin of ferrous and ferric oxide and hydroxide minerals. The great importance attributed by the authors to the issue whether the ferric and ferrous ions were precipitated jointly or separately is clearly apparent in the diagram.

According to Schwertmann (1969), organic matter inhibits the crystallisation of ferric oxides and hydroxides. This is why soils rich in organic matter tend to contain

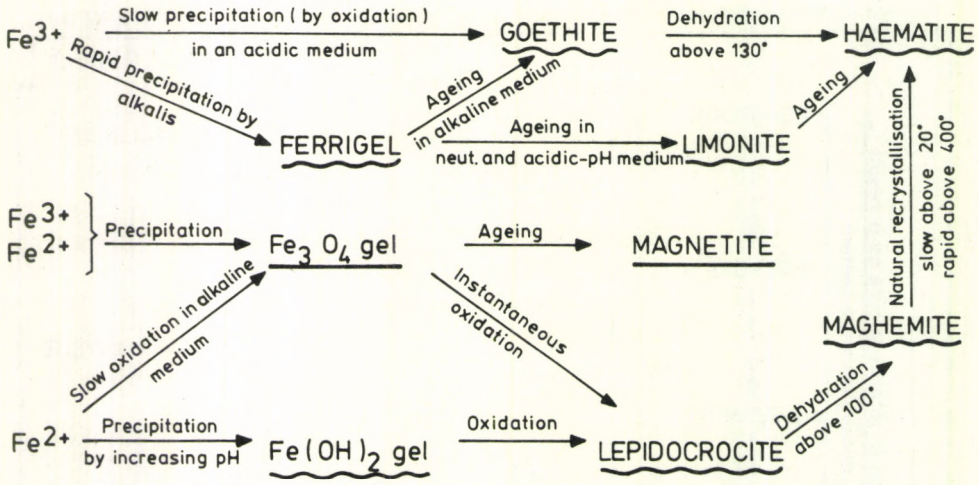


Figure 164. Genetic system of iron oxide and iron hydroxide minerals in the soils. After Scheffer *et al* (1957)

abundant amorphous ferric hydroxide. Under a warm, wet climate, on the other hand, where organic matter decomposes fairly rapidly, ferric hydroxide will soon crystallise. The process is promoted by the presence of carbonate minerals and of Ca ions. This is presumably why amorphous ferric hydroxides are found but sporadically in karst bauxites, and then only in very low abundances.

Chukhrov (1973, 1975) described using the name *ferrihydrite* a new metastable ferric hydroxide mineral ($2{\cdot}5Fe_2O_3 \cdot 4{\cdot}5H_2O$) which is converted into goethite in a ground water containing ferrous ions, and into haematite under any other conditions. According to Chukhrov, ferrihydrite is formed under the influence of certain bacteria (*Gallionella, Leptothrix, Toxothrix*) in the course of the weathering of iron-containing minerals. Under conditions of abiogenic weathering, on the other hand, goethite and lepidocrocite are formed directly (the former if oxidation is rapid and the latter if oxidation is slow).

Ferrihydrite has been identified so far in Recent and sub-Recent sediments only, but according to Chukhrov its formation was also widespread in time and space in the geohistorical past. It played an important role in the formation of haematitic iron ores and of laterites, as proved by the fact that traces of the above-named bacteria have been identified in numerous ancient sediments and sedimentary iron ores. The present writer deems it to be probable for ferrihydrite to have played a role in the formation of the bauxites' iron minerals, but proving that assumption will require a good deal of purposeful geo-microbiological research.

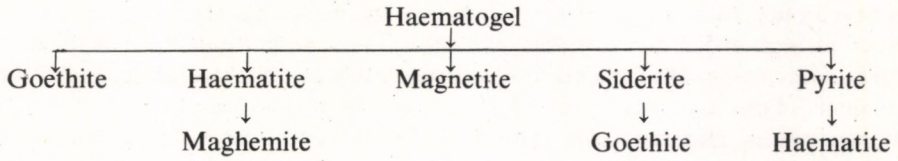
If we intend to extend our considerations also to the ferrous sulphides and carbonates, it will be necessary to devise sets of Eh–pH diagrams each of which corresponds to a given concentration of sulphur and carbonate ions. The relevant calculations were performed by Garrels and Christ (1965). The essence of their findings is that, in the presence of sulphur ions, pyrite and marcasite are formed if the medium is acidic or neutral and the Eh is a reducing one, whereas magnetite is formed if the medium is strongly alkaline. If there are carbonate ions present, siderite is formed in the

pH range from 6 to 10. According to Barner (1970), the optimum conditions of siderite formation are a low S^{2-} concentration, a high CO_2^- and Fe^{2+} concentration, a low Eh and a near-neutral medium. He conjectured these conditions to be optimally satisfied by fresh swamp water. Most sedimentary siderite deposits came to exist under conditions such as these.

The thermodynamic parameters of iron silicates being unknown as yet, it is impossible to plot the boundaries of their ranges of stability in the Eh–pH diagrams. Chamosite owing to its varying Fe^{2+} and Fe^{3+} content may form in environments of different Eh values. According to Harder (1973), chamosite is formed out of a Fe–Al–Si gel under a reducing Eh by processes of early diagenesis. If the Eh increases, iron cannot enter the silicate lattice any more: it will form ferric oxide or hydroxide minerals.

Porrenga (1967) in his mineralogical studies of recent marine sediments found that chamosite can form close inshore at depths less than 60 m under a tropical climate, in sea water of 20 °C or more. At greater depth or in cooler water, chamosite is replaced by glauconite. Schellmann (1966) identified chamosite in recent clayey sediments, the derivatives of laterite erosion, off the Guinean coast. It was formed by the chamositisation of minute grains of goethite under the influence of sea water containing Al, Si and Mg, saturating the pores of the sediment. According to Schellmann, chamosite rather than kaolinite is the stable phase under slightly reducing conditions in near-shore marine sediments. Iron-rich leptochlorites will also form in a marine environment under reducing conditions according to most handbooks of mineralogy.

As far as the bauxite deposits are concerned, Beneslavsky assumed the following succession of iron-mineral transformations:



Remarkably enough, Beneslavsky even derived siderite and pyrite from the original haematogel, relegating to a place of minor importance the secondary reduction of haematite and goethite. According to his diagram, only goethite was formed by the secondary oxidation of siderite, and only haematite by that of pyrite. According to Beneslavsky, siderite and goethite should be found in the trihydratic bauxites and pyrite and haematite in the monohydratic ones.

The present writer attributes a great deal of importance to Beneslavsky's statement that alumogoethite, alumohaematite and ferridiaspore may each be formed by gel ageing, whereas the Al and Fe minerals precipitating out of ionic solutions are stochiometrically pure.

The conditions of formation of the *clay minerals* in bauxite are known largely thanks to clay mineral research. Let us therefore confine our discussion here to a few hypotheses directly relevant to bauxite.

According to Erhart (1956), the kaolinite lattice, formed by surface weathering, is so resistant as to defy decomposition even in the course of lateritisation and bauxitization.

According to Beneslavsky (1959), the vast bulk of the clay minerals found in bauxite is not derived from the outside (i.e. it is not continental-clastic), but of syngenetic origin, crystallised out of the complex primary gel already referred to. In his opinion, the affinity of silicon to aluminium in the primary gel was so great as to prevent the formation of any free Al minerals until all the silica present had been used up to form some clay mineral. This is why bauxite contains no syngenetic quartz or cristobalite. In the course of crystallisation, halloysite or pyrophyllite may form: the former may be converted into kaolinite. Pyrophyllite and the hydromicas of continental origin may also get kaolinitised by secondary processes. In epigene resilification, on the other hand, kaolinite and halloysite may build up out of silica dissolved in ground water plus the aluminium minerals of the bauxite.

Gladkovsky and Ushatinsky (1964) assumed mineral transformations similar by and large to those proposed by Beneslavsky. They attributed a great importance especially to supergene resilification, which produces halloysite by preference in an alkaline medium and kaolinite in a weakly acidic one (cf fig. 163).

Titanium minerals have attracted the least attention so far. According to Valeton, the principal titanium mineral of most laterite profiles is cryptocrystalline anatase. In Hawaiian laterites, Matsusaka *et al* (1965) found authigenic titanomagnetite containing as much as 21 to 25 mol % TiO_2. The oxidation of Fe^{2+} ions, on the other hand, may trigger the formation of titanomaghemite. The presence of these minerals must also be reckoned with in the karst bauxites.

5.5.3. A GENETIC SYSTEM OF ROCK-FORMING BAUXITE MINERALS

In developing his own minerogenetic system, the writer has started from the facts outlined in Section 5.3, and from his own statistical computations. He has taken the findings of laboratory experiments into account only where they were compatible with the evidence of observations in Nature. He agrees with Beneslavsky in that some of the part-processes of minerogenesis cannot as yet be reproduced in the laboratory, especially in view of their extremely slow reaction rates. In the following, let us review mineral formation in the temporal succession of the processes involved, as illustrated by the diagram in figure 165.

As opposed to what has been assumed by many authors including Beneslavsky, the primary material accumulated at the deposit sites was not necessarily and exclusively an amorphous complex gel: in the bauxites of clastic texture above all, it presumably included abundant minerals of allothigenic origin, as follows:

(i) The most resistant minerals of a sedimentary or igneous parent rock: quartz, zircon, rutile, corundum, magnetite, ilmenite etc. Their aggregate abundance in karst bauxites, however, is only in the tenth-of-a-percent range.

(ii) If the parent rock was a clay or a rock rich in clay minerals, then minute grains of clay minerals (kaolinite, halloysite, sericite and illite) could be delivered to the site of accumulation of the bauxite deposit in substantial quantities. Under the climatic conditions envisaged, most of the montmorillonite, vermiculite, chlorite and mixed-lattice clay minerals were decomposed in the course of weathering and transport.

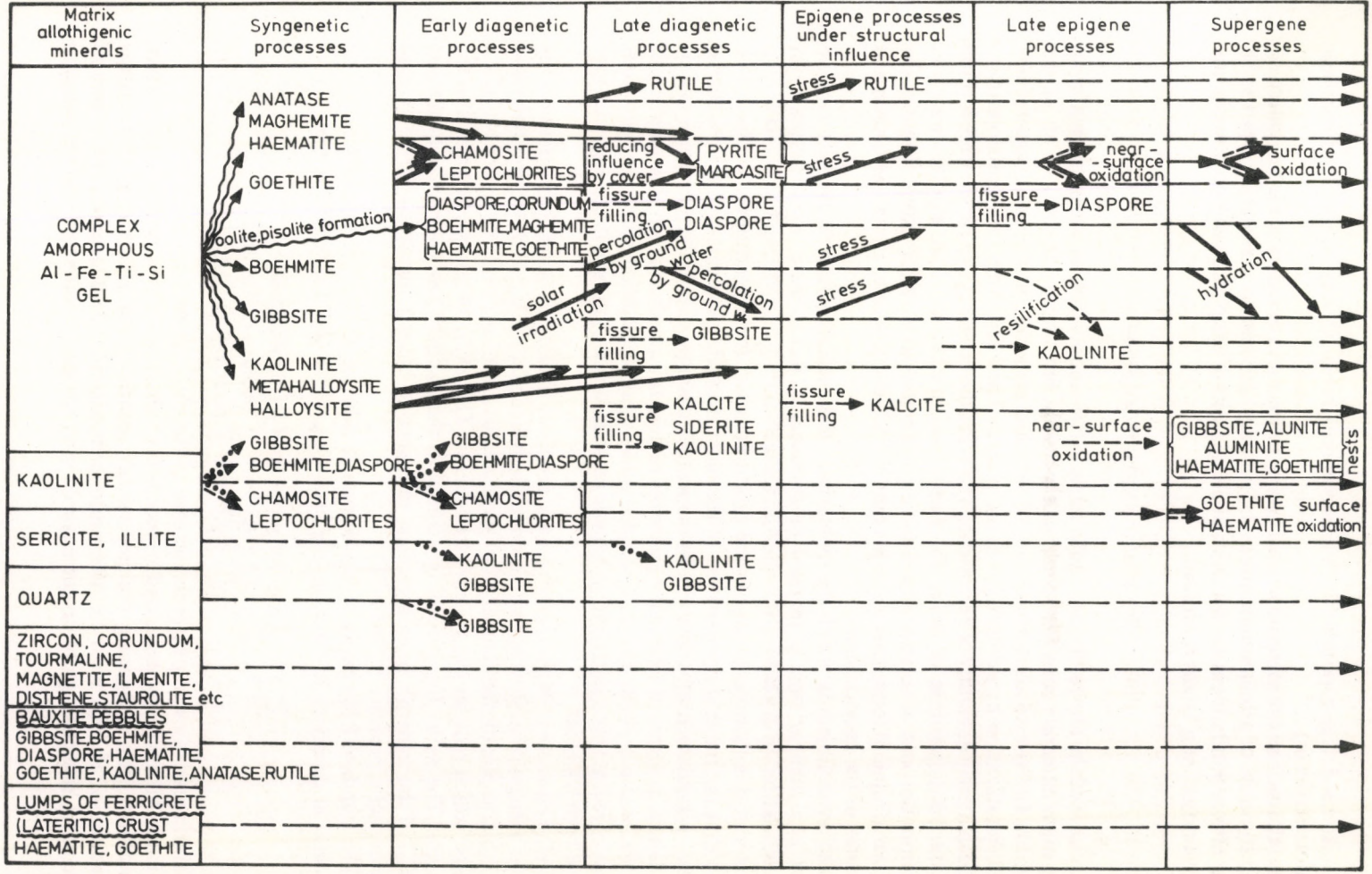


Figure 165. Genetic system of the most frequent minerals of karst bauxites. Wavy line: crystallisation by gel ageing. Dotted line: decomposition of existing minerals, partial leaching. Short-dashed line: precipitation out of a solution. Full bold line: recrystallisation, hydration, dehydration. Long-dashed line: minerals left unchanged

(iii) In the Ariège-type deposits, the parent rock probably was a normal sedimentary clay or clay marl.

(iv) Clastic bauxite material (clastic grains, pebbles and boulders of bauxite), derived from some site of primary bauxitisation. This may have been of the lateritic type too, in view of the abundance in some karst bauxite deposits of lumps derived from a ferricrete, the 'cuirasse' of a lateritic profile.

5.5.3.1. SYNGENETIC PROCESSES

In view of the above-outlined differences in the composition of the initial material, *syngenetic mineral formation* could not possibly have been identical in all of the deposits. In those deposits where most of the initial material consisted of an amorphous gel, crystallization by gel ageing was the principal process. Where the initial material consisted predominantly of pre-existing minerals, the syngenetic minerals were formed by the decomposition of those. In cases of sedimentation in an aqueous medium, precipitation out of solutions may have contributed to a minor extent. Finally, the nature of mineral formation was considerably influenced also by the physico-chemical features of the environment.

The environment in which the Quaternary and even the Pliocene deposits came to exist is a familiar one: a tropical or subtropical, wet climate; a lush vegetation; accumulation on dry land with temporary inundations; an oxidizing Eh. We have seen what sort of mineral associations are produced under these conditions. We have also seen that the mineral composition of the older deposits—especially of the Paleozoic ones—is fundamentally different. In the writer's opinion, not all the differences can be explained in terms of secondary processes of mineral transformation: the majority of the Paleozoic deposits had accumulated under conditions different to start with from those of recent times, and their syngenetic mineral formations must in consequence also have been different.

In the assessment of the minerogenetic processes of bauxite, insufficient attention has so far been paid in the writer's opinion to the evidence contributed by thermodynamics, and this despite the familiar fact that spontaneous chemical processes invariably take a course which reduces the system's free energy. This must, of course, be valid also for bauxite. The writer has accordingly compiled the standard free energies of formation (ΔF_f°) of the minerals having greatest relevance to bauxite formation for 25 °C and atmospheric pressure (cf table 41).

The standard free energy of formation is a function also of the crystallinity of the mineral in question. In the case of gibbsite, e.g. Fritz and Tardy (1973) give

$-275{\cdot}3$ kcal/mol for gibbsite of fair crystallinity,
$-274{\cdot}9$ kcal/mol for gibbsite of medium crystallinity,
$-273{\cdot}0$ kcal/mol for gibbsite of poor crystallinity.

The spread is even greater in the case of the clay minerals.

Table 41 reveals gibbsite and bayerite to be the most stable phase under surface conditions. (In comparing the free energy values of the alumina minerals equal numbers of H ions must be taken into consideration. This doubles the values of boehmite, diaspore, gibbsite and bayerite with respect to alumogel and corundum.)

Table 41. Standard free energies of formation of the bauxites' rock-forming minerals at 25 °C temperature.

Mineral	State	F_f^0, kcal mol^{-1}	Reference
Alumogel	Amorph.	− 271·3	KITTRICK, *Soil. Sci. Proc.* 1966
Boehmite	Cryst.	− 218·7 ±0·2	PARKS, *Am. Miner.* 1972
Diaspore	Cryst.	− 219·5 ±0·5	PARKS, *Am. Miner.* 1972
Gibbsite	Cryst.	− 273·5 ±0·3	ROBIE and WALDBAUM, *US Geol. Surv. Bull.* 1968
Bayerite	Cryst.	− 276·2	Techn. Rep. Bruxelles 1960
Corundum	Cryst.	− 378·1 ±0·3	E-AN-ZEN *Am. Miner.* 1972
Goethite	Cryst.	− 117·04	MELNIK, *Thermodyn. Constants. Kiev.* 1972
Haematite	Cryst.	− 177·7 ±0·3	ROBIE and WALDBAUM, *US Geol. Surv. Bull.* 1968
Magnetite	Cryst.	− 243·1 ±0·5	ROBIE and WALDBAUM, *US Geol. Surv. Bull.* 1968
Pyrite	Cryst.	− 38·3 ±0·4	ROBIE and WALDBAUM, *US Geol. Surv. Bull.* 1968
Siderite	Cryst.	− 161·0 ±0·5	ROBIE and WALDBAUM, *US Geol. Surv. Bull.* 1968
Anatase	Cryst.	− 212·63±0·3	ROBIE and WALDBAUM, *US Geol. Surv. Bull.* 1968
Rutile	Cryst.	− 212·56±0·1	ROBIE and WALDBAUM, *US Geol. Surv. Bull.* 1968
Silica gel	Amorph.	− 203·3 ±0·5	ROBIE and WALDBAUM, *US Geol. Surv. Bull.* 1968
Quartz	Cryst.	− 204·6 ±0·4	ROBIE and WALDBAUM, *US Geol. Surv. Bull.* 1968
Daphnite	Cryst.	− 822·4	GORBACHEV, *Gen. boksitov* 1966
Kaolinite	Cryst.	− 910·0	E-AN-ZEN, *Am. Miner.* 1972
Pyrophillite	Cryst.	−1260·0	E-AN-ZEN, *Am. Miner.* 1972
Illite	Cryst.	−1277·0	ROUTSON and KITTRICK, *Soil. Sci. Proc.* 1971
Muscovite	Cryst.	−1340·0	E-AN-ZEN, *Am. Miner.* 1972
Clinochlore	Cryst.	−1974·0	E-AN-ZEN, *Am. Miner.* 1972

Gibbsite must accept energy or be placed under special physico-chemical conditions if it is be converted into boehmite, diaspore or corundum. The physico-chemical conditions in question may include certain values of the Eh and pH, the presence of alien ions or of seed crystals. From alumogel, on the other hand, each of the crystalline aluminium minerals may be spontaneously derived, as far as the thermodynamics of the process is concerned.

Gel ageing under surface conditions, then, mainly produces gibbsite, as proved by the composition of Miocene and Pliocene bauxites (table 16).

The experiments of Schoen and Roberson (1970) and the geological conditions of occurrence of nordstrandite and bayerite imply that the formation of those two minerals was promoted by an alkaline medium. Both will recrystallise into gibbsite even under a minimum of overburden pressure, however. This is why they are absent from the pre-Tertiary, buried bauxite deposits. On the other hand an alkaline ground water may also result in the secondary recrystallization of gibbsite into nordstrandite (Kulikova *et al* 1974).

In the writer's opinion, the formation of boehmite in rock-forming abundance must have been confined to where the concentration of dissolved HCO^{3-} in the ground water was comparatively high (cf the experiments of Schellmann 1964a). Such conditions could prevail only at or below the ground water table, in deposits in immediate contact with carbonate rocks. This is where the mineralogical role of the carbonate bedrock comes to the force—*this is why karst bauxites are more boehmitic than lateritic bauxites.*

As we saw it in table 15, the Mediterranean-type deposits, the most boehmitic ones of all karst bauxites, are in fact the deposits most closely attached to carbonate rocks. The

reason why no boehmitic bauxite is found in the Kazakhstan-type deposits is that their peculiar deposit features hampered their penetration by hydrocarbonate solutions.

The formation of alumina minerals is a complex phenomenon which may also have been influenced by the oxidation–reduction potential and the pH. The writer, in agreement with Valeton, Combes and Komlóssy feels the preferred associations to have been gibbsite and goethite at a high enough Eh and in a weakly acidic medium, and boehmite and haematite at a somewhat lower but still positive Eh and in a weakly alkaline medium. This assumption is confirmed among others by the model experiments, referred to above, of Pedro *et al* (1970). The dependence on the oxidation–reduction potential is proved among others by the relationship detected by Bárdossy (1961a) between the boehmite/gibbsite ratio and the Fe_2O_3/FeO parameter in the Transdanubian Hills of Hungary.

With regard to the Quaternary and Pliocene bauxites' mineral associations, the writer feels that some maghemite may also have formed syngenetically. This is confirmed by the strikingly high maghemite contents of some laterite profiles (Bonifas and Legoux 1957; Schellmann 1964b; Balkay and Bárdossy 1967).

The ageing of a primary gel under a positive Eh resulted in the preferential crystallisation of titanium into anatase, and of silica partly into halloysite and metahalloysite and partly into kaolinite.

Wherever the primary material contained abundant kaolinite of allothigenic origin, the principle of 'energy inheritance' invoked by Keller *et al* (1954) permitted diaspore and boehmite to form under conditions of a slow uninterrupted percolation by ground water. This assumption is confirmed by the diasporo–boehmitic and diasporic bauxites formed out of clays and clay marls in the Ariège-type deposits. The mineral association implies a zero or slightly negative Eh to have prevailed throughout this type of mineralisation: that is, the oxidation–reduction potential also played a certain role. This process is so slow that it cannot be regarded as wholly syngenetic; it overlaps into the phase of early diagenesis.

Intense percolation by ground water and a positive Eh made even a kaolinitic primary material yield gibbsite, a situation observed in numerous laterite profiles. According to Garrels and Christ (1965), the process can be characterised in thermodynamical terms as follows:

$$\underset{\text{Kaolinite}}{H_4Al_2Si_2O_9} + \underset{\text{Water}}{5H_2O} = \underset{\text{Gibbsite}}{Al_2O_3 \cdot 3H_2O} + \underset{\text{Dissolved silica}}{2H_2SiO_4}$$

$$F_r^\circ = \Delta F^\circ_{f\,\text{Gibbsite}} + 2\Delta F^\circ_{f\,\text{Silica}} - \Delta F^\circ_{f\,\text{Kaolinite}} - 5\Delta F^\circ_{f\,\text{Water}}$$

$$\Delta F_r^\circ = -554{\cdot}6 + 2(-300{\cdot}3) - (884{\cdot}5) - (-56{\cdot}7) = 12{\cdot}7\ \text{kcal.}$$

Calculating the equilibrium coefficient of this system, one finds it to be balanced if the ground water contains 2 ppm H_4SiO_4. For silica contents lower than that, kaolinite dissolves incongruently, and gibbsite is formed. We should not forget the buffering effect already referred to, which affects ground water composition and thereby controls the option of gibbsite formation. The contribution of the micro-organisms to this control should also be pointed out once more.

As we saw it in the chapter of deposit features, bauxite accumulation, especially in the Paleozoic, often continued in seashore lagoons and bays. Syngenetic minerali-

sation was in such cases influenced by the temperature, composition, Eh and pH of the sea water and also by the organic substances present. Under a reducing pH, in the presence of abundant organic matter, syngenetic chamosite and leptochlorite was formed. The marine environment thus specified is fully compatible with Schellmann's and Porrenga's observations described in Section 5.4.2. The syngenetic aluminium minerals of this facies may have been diaspore and boehmite.

5.5.3.2. DIAGENETIC PROCESSES

Most of the processes of syngenetic mineralization continued into the early phases of *diagenesis*. The role of mineralisation by gel ageing, however, grew increasingly subordinate. Recrystallisation came to the fore, and so did the decomposition, hydration and dehydration of the minerals. A lesser role was played by precipitation out of solutions.

The most important process was the decompostition of kaolinite into gibbsite, boehmite or diaspore as a result of leaching by ground water under the diverse conditions enumerated above, by processes much the same as in the syngenetic phase. Most of sericite was decomposed into kaolinite. In the lagoons of the seashore, the conversion of kaolinite into chamosite and leptochlorites continued, the iron required being contributed by the reduction and redissolution of goethite and haematite.

One of the most typical processes of early diagenesis was *the formation of ooids, pisoids and macropisoids*. This is essentially a process of segregation, starting about a central nucleus and proceeding outward therefrom. We may, then, state fairly confidently that the original state was not a homogeneous sphere in which spherical shells started to segregate all of a sudden. Segregation presumably took place in a soft, gel-like material saturated with water, as proved by the numerous fissures and veins, produced by gel compaction, traversing the concentric particles. The process of segregation was presumably initiated by the greater adhesion of the mineral grain or bauxite particle in the nucleus: in fact, the differences in adhesion coefficient between the Al and Fe hydroxides may even have played a role in the formation of the individual shells.

The essence of the process is the colloid-chemical segregation of Al and Fe gels from a silicate gel. The shortest path of mass transport by which such a segregation could take place is defined by the surface of a sphere: this is why these particles are spherical. Al and Fe gels migrated inward: silica gel migrated outward. At the same time, the Al and Fe gels also segregated into successive low-iron and iron-rich shells. The rate of segregation presumably exhibited a rhythmic alternation, each minimum coinciding with the completion of a double shell. The entire process was probably brought to a halt by advanced crystallisation due to gel ageing, which froze in the entire system, as it were.

Of the iron minerals, those have a tendency to concentrate in the concentric particles which are also most abundant in the surrounding matrix; that is, the process is essentially a quantitative rather than a qualitative one. The only qualitative difference is the appearance of maghemite in some pisoids and macropisoids, and the appearance of magnetite in the concentric particles of some Kazakhstan-type deposits. At the aluminium minerals, the process of crystallisation did not progress to gibbsite, the

phase of lowest free energy, but usually remained in the diaspore state: in fact, the phase formed in the Kazakhstan-type deposits was corundum. The data in table 41 show the free energy of formation of corundum to be hardly less than that of amorphous alumogel. The possibility of a spontaneous chemical change does thus exist even here the question is, why did the change stop. The writer prefers to invoke the nucleating influence of iron minerals, as assumed also by Slatkine and Heller (1961) and by Lisitsyna and Pastukhova (1963). Goethite is well known to be isomorphous with diaspore, haematite with corundum; furthermore, Wefers's experiments have revealed the formation of isomorphous mixed crystals to be supplanted by orientated intergrowths of the corresponding pairs of minerals whenever the Al concentration increased and the Fe concentration decreased sufficiently. This process may have been promoted by the lowered silica content of the concentric particles.

With a view to clarifying this issue, the writer has prepared X-ray diffractograms of brownish-black pisoids and macropisoids from a laterite sample collected near Perth, Australia, and has found those to be enriched against their surrounding matrix in haematite and maghemite; some poorly crystalline corundum is also present. Gibbsite and goethite contents are much reduced, whereas kaolinite, of low abundance in the matrix, has disappeared completely in the pisoids. All this proves that the above-outlined processes may have taken place in a similar fashion in lateritic bauxites also. The laterite in question having been exposed on the ground surface since its formation and been unaffected by any structural or metamorphic influence, this is unequivocal evidence that cryptocrystalline corundum may in fact form under surface temperature and pressure in the course of pisoid and macropisoid formation.

There were also diagenetic mineral transformations other than the formation of concentric particles. For example, in the deposits which after their accumulation remained unburied over prolonged spans of time under a favourable climate, processes similar to those in the laterites took place. It is known that in the top parts of these unburied laterite profiles gibbsite is often accompanied by boehmite and more rarely even by diasprore. Balkay and Bárdossy (1967) explained this in terms of an energy input by insolation. In their opinion, the warming of the ground under insolation raises the thermal energy level there into a close proximity of the energy level of the gibbsite–boehmite transition. At the very surface, even the direct radiation energy received may be sufficient to achieve this. Wherever the energy input attains or surpasses the energy level of the transition, the crystallites of gibbsite are converted into boehmite one by one. This reaction may progress as long as the energy input is maintained. In a similar manner, but at an even lower rate, boehmite grains are converted into diaspore. This mechanism of transformation is of course promoted by the hot and dry seasons and climatic phases, during which the top of the profile can be warmed up quite substantially. In the writer's opinion, however, this mode of mineral transformation plays a subordinate role in karst bauxites as compared with the lateritic bauxites.

Instances of deposits buried under an overburden are much more frequent. In those deposits, ground water seeping through the bauxite triggered a process going in the opposite direction, that is the hydration of gibbsite into boehmite.

The transformation is an energy-liberating one; that is, hydration may come about spontaneously under the influence of ground water. Since ground water percolates the deposits from the top downward as a rule, gibbsitisation also progressed in that

direction, as observed for example in the deposits of the Southeastern Bakony Hills, Hungary. According to the findings of Goldich and Bergquist (1947, 1948), this process could be initiated by a simple increase of average rainfall in the region. Clearly, the more abundant the precipitations, the more favourable was the situation for the diagenetic hydration of boehmite (cf Section 5.2.4).

The diasporisation of some deposits of boehmitic constitution may also have been a diagenetic process (e.g. in Central Greece). In agreement with Nia, the writer attributes the principal role here to the formation of diaspore seeds in addition to an intense ground water percolation. It should not be forgotten that this process also was an energy-liberating one which could thus proceed spontaneously (cf table 41). The seeds in the present writer's opinion intervene only to accelerate the reaction which would be extremely slow otherwise.

Depending on the external physico-chemical conditions, then, boehmite may have passed into gibbsite in certain places and into diaspore in others. These two opposite processes are, of course, mutually exclusive; they cannot both act in the same deposit at the same time. It is well known that, of the three minerals, boehmite constitutes the smallest crystallites. This may also have been a factor promoting the above-outlined processes.

It was in the course of diagenesis that most of metastable maghemite crystallised into haematite, and the first-formed phases halloysite and metahalloysite dehydrated into kaolinite.

Wherever the immediate cover deposited above the bauxite was of swampy facies, the top parts of the deposits underwent diagenetic and early epigene reduction (cf Section 5.2.4). The hydrogen sulphide, formed in such cases by the decomposition of organic matter in the swamp, reduced haematite and goethite into pyrite and marcasite, and less seldom—if the concentration of dissolved CO_2 was high enough—into siderite. Part of the iron was removed in solution. The other minerals were not significantly affected, and, as contrary to syngenetic reduction, no chamosites and leptochlorites were formed, either. The silica content of the downward-seeping solutions united with the aluminium minerals to form secondary kaolinite ('re-silification'). This process, however, is not significant with respect to volume in this phase. The kaolinite thus formed has a well ordered lattice structure.

5.5.3.3. EPIGENE PROCESSES

By the onset of the *epigene processes*, mineralisation connected with gel ageing had become exceptional. Recrystallisation, hydration and dehydration came to the fore, and it is here that the role of precipitation out of solution was the greatest.

As regards the prime movers of the epigene processes, the writer does not ascribe much of a significance to overburden pressure, although according to some specialists gibbsite is gradually converted into boehmite even under an overburden of a few dozen metres. This view, however, is contradicted by the fact that boehmito–gibbsitic bauxite in the northern part of the Rákhegy ore body, Iszkaszentgyörgy, Hungary still contains 17 to 21% gibbsite at a depth of 350 to 370 m. This is the deepest occurrence of gibbsite known to this writer.

Structural (tectonic) stresses and deformations, on the other hand, may have had a profound effect on a mineral association. This is revealed for example by the data in table 17, as well as by Terentyeva's observations (1958) on the deposits at Boksonsk and in the Southern Urals, USSR (cf Section 5.2.4). Structural influences may have recrystallised gibbsite into boehmite and boehmite into diaspore, in good agreement with Gout's and Kra Kouadio's (1973) experimental findings. Those authors synthesized ferridiaspore by grinding a mixture of gibbsite and goethite for 6 to 20 h at 110 to 210 °C. The transformation will also take place at lower temperatures, but at much slower rates: for instance, 240 h grinding is required at 50 °C. The authors found the amount of diaspore produced and its crystallinity to be the higher, the more goethite there was in the mixture, other parameters being equal. The presence of goethite promotes according to them the formation of diaspore as a nucleating agent. Above 210 °C, grinding gave rise to corundum rather than to diaspore. This is a much slower temperature than the lower temperature limit of corundum formation postulated by the investigators of the phases in the Al_2O_3–Fe_2O_3–H_2O system (cf Section 5.5.1.3).

Structural (tectonic) stresses tend to recrystallise goethite into haematite, a process that is very easy to initiate in view of the slight energy change involved. This assumption is confirmed by the findings of Gout and Ferrier (1972), who converted goethite into haematite by grinding at room temperature for 20 to 60 h. This is why goethite gradually disappears from the Mesozoic and Paleozoic deposits (table 19).

Anatase partly recrystallised into rutile. This process, however, is so slow that no full recrystallisation has been achieved since the Paleozoic deposits.

The effects of structural stress are highlighted even more strikingly by a comparison of the molecular volumes of the aluminium hydroxide minerals:

gibbsite: $31{\cdot}965 \pm 0{\cdot}015\ cm^3$
boehmite: $19{\cdot}535 \pm 0{\cdot}026\ cm^3$
diaspore: $17{\cdot}760 \pm 0{\cdot}026\ cm^3$

(data published by Robie and Waldbaum 1968). Clearly, pressure tends to generate minerals of lower molecular volume. The same holds for the two principal titanium minerals, whose molecular volumes are:

anatase: $20{\cdot}52 \pm 0{\cdot}3\ cm^3$
rutile: $18{\cdot}82 \pm 0{\cdot}1\ cm^3$

Structural stress does not exert the same mineral-transforming influence everywhere, as pointed out in connection e.g. with the Eocene–Paleocene deposits (table 17). The reasons for the deviations are not yet understood. The writer deems it probable that the mineral-transforming effect could be controlled among others by the duration and frequency of repetition of structural stress and further by the impermeability of the immediate cover of some deposits.

The fillings of the veins and cracks traversing the bauxite are epigene in their majority, formed as they were by precipitation out of ionic solutions. The pore fillings, segregations of irregular shape, nests and concretions are also epigene in their bulk. From the minerogenetic standpoint, their most striking feature is that they are all monomineralic or at most di- or trimineralic.

One of the most important epigene processes is *resilification*. If ground water percolating the deposit had a high in dissolved silica contact, this latter tended to unite with some of the aluminium minerals, forming kaolinite, or more seldom halloysite. The kaolinite thus formed is invariably highly crystalline. In places, resilification was strong enough to transform an entire bauxite deposit (e.g. Ollières, France) into bauxite clay.

The near-surface oxidation of the bauxites of reduced composition is partly a late epigene, and partly even a supergene process. Oxidation was initiated by oxigen-saturated rainwater seeping into the ground. It decomposed pyrite and marcasite, forming a solution of sulphuric acid and ferric sulphate. This latter seeped downward in solution, to be reprecipitated soon enough in the form of goethite, or goethite plus haematite, forming irregular segregations and nests. The sulphuric acid in the downward-seeping solution dissolved even part of the aluminium minerals, and then precipitated the aluminium in the form of gibbsite nests. Elsewhere, uniting with the sulphate content of the solution, it was converted into nests of alunite or aluminite. In the presence of calcium ions, gypsum was precipitated out of the sulphate solutions.

The writer ascribes the formation of most of the incrustations and segregations at the bedrock boundary—an issue still disputed by many workers—also to the above-outlined processes. Systematic observations over many years have revealed these incrustations to be confined to unburied deposits and to deposits under a thinnish overburden. Under an overburden of 40 to 60 m, they grow thin and sporadic; in fact, there are none under deposits buried deeper than 100 m. Control by overburden thickness implies a supergene influence, the main factor of which must have been the downward-seeping solutions. The segregations described in the chapter on systematic mineralogy, however, require a degree of chemical mobilisation of which normal ground water is incapable. The obvious way out is to invoke the influence of the sulphuric acid solutions produced by pyrite oxidation. The assumption is borne out by the sulphate minerals identified in some crusts (alunite, gypsum), and also by the fact that there are traces of epigenically oxidised pyritic bauxite above most of the incrustations.

The writer envisages the formation of these incrustations as follows. The downward-seeping acidic solutions were neutralised when they attained the bedrock boundary. In step with the increasing alkalinity of the solutions, the first to precipitate was ferric iron in the form of goethite and haematite: it was followed by aluminium in the form of gibbsite, and finally by manganese in the form of lithiophorite and todorokite. This assumption is confirmed by the fact that the crusts are almost entirely devoid of titanium. A pH less than 2, needed for the mobilization of titanium, was obviously not attained by the downward-seeping acidic solutions.

An effect similar to the above but weaker may have been produced also by humic acids. Under the Jamaican deposits, where no pyritisation has ever taken place, lithiophoritic–todorokitic black segregations are nevertheless found at the bedrock boundary. There are, however, no haematitic, goethitic and gibbsitic segregations there. Clearly, the weaker dissolving power of the solutions involved could only mobilise manganese, but not the aluminium and iron minerals. This process is not necessarily epigene: it could in principle act uninterrupted to this day, ever since the accumulation of the bauxite.

5.5.3.4. SUPERGENE PROCESSES

In the deposits uncovered secondarily, substantial supergene mineral transformations took place. The writer agrees with Gladkovsky and Ushatinsky (1964), who stated that the transformations in these situations generate reverse successions, from corundum towards gibbsite (fig. 163). This assumption is confirmed by free energy considerations. For such a process to take place, however, the crystallites must stay in permanent contact with the ground water. This is the more difficult, the harder and the more compact the bauxite, the lower its porosity. This is why compact Palaeozoic bauxites usually undergo only superficial hydration in the exposures, with the original composition of the bauxite preserved at a depth not greater than one to two centimetres.

In summary, the writer holds bauxite to be an extremely sensitive, variable system also in the mineralogical respect, which from the time of its accumulation to the present day has always been striving for an equilibrium optimally matching it to its environment. This is how such a variety of mineral transformations, often apparently contradictory, could take place in the different deposits. The conflicting views of the earlier authors are due principally to the fact that they did not recognize the closely controlled nature in both time and space of these mineralisations, which in another deposit at another time could progress in a direction opposite to the usual.

The above considerations justify furthermore the conclusion of general validity that the principal clay minerals are no stable end-products under Earth-surface conditions, but that, adapting to changes in conditions, they may be decomposed and reformed, or converted into other clay minerals.

6. KARST BAUXITE AND TERRA ROSSA

In karst bauxite literature, the Italian term 'terra rossa' ('red earth'; plural: terre rosse) is fairly widespread. Alas, opinions differ rather widely as to what exactly this terra rossa is and what is its genetic relation to the bauxites. Let us therefore mention in a historical review first of all the research results relevant to the composition and constitution of terra rossa. Subsequently, let us survey the literature on the origin of terra rossa, also in a historical manner.

Since the second half of the last century, the term terra rossa has been used to denote a reddish or brownish-red, comparatively soft deposit of clay-like appearance, very widespread in the Mediterranean region, which covers the karsted surfaces of limestones and dolostones to a thickness ranging from a few centimetres to some two metres. Usually covered by a maquis-like vegetation, occasionally by light woods, it accordingly is considered a type of soil by the pedologists, and a geological formation by the geologists.

6.1. THE COMPOSITION OF TERRA ROSSA

The first to investigate terra rossa thoroughly was Tučan (1912) who studied samples taken from six points of the Yugoslav karst. The chemical analyses of his samples are presented in table 42. Tučan found part of the silica content of terra rossa to be soluble in hydrochlorid and/or nitric acid. (From the Zlobin sample, e.g. he could dissolve 9·0% SiO_2; 17·2% turned out to be indissoluble.) From this fact and from his microscopic

Table 42. Chemical composition of terra rossa samples from Yugoslavia, according to Tučan (1912).

Sampling locality	Al_2O_3	SiO_2	Fe_2O_3	TiO_2	MnO	CaO	MgO	ZrO_2	Alk.	CO_2	H_2O	Total
Zlobin	39·14	26·20	14·03	0·51	1·45	Trace	–	0·81	Trace	–	18·14	100·28
Plase	32·89	35·42	15·03	0·30	0·93	0·43	Trace	0·10	Trace	0·23	15·32	100·55
Grolničko polje	27·80	43·61	11·75	Trace	Trace	1·64	Trace	Trace	Trace	–	15·43	100·23
Jelenje	24·38	47·89	12·63	0·96	1·18	0·68	Trace	Trace	0·32	0·39	11·86	100·29
Karlobag	26·61	46·27	12·64	0·80	0·12	1·13	Trace	Trace	Trace	–	13·32	100·89
Županjec	20·19	26·47	18·03	Trace	1·32	13·19	Trace	Trace	Trace	9·12	11·24	99·56
Županjec	25·69	32·11	6·20	Trace	–	14·44	0·48	Trace	Trace	10·77	11·33	100·97

findings he inferred the greater part of silica to be in the form quartz, and its lesser part to be in the form of an amorphous silica gel. He assumed the bulk of terra rossa to consist of an amorphous aluminium hydroxide gel which he called *sporogelite*, and iron in the form of ferric hydroxide and oxide gels. The principal mineral of titanium is in the form of the mineral zircon. The manganese content is also in gel form (a manganese hydroxid gel). Tučan identified very subordinate amounts of the following minerals, visible under the microscope only: muscovite, phlogopite, biotite, epidote, zoisite, clinosoisite, amphibole, garnets, chlorite, disthene, microcline, albite, oligoclase, andesine, bytownite, anorthite, staurolite, titanite, tourmaline, periclase, brucite, apatite, haematite, pyrite, calcite, dolomite. He found comparatively frequent plate or irregular gibbsite grains of 0·01 mm size, and identified a few diaspore grains of 0·02 to 0·04 mm in size and a few corundum grains of 0·04 to 0·25 mm in size.

Tučan obtained dissolution residues of the carbonate rocks of the region, using dilute acetic acid to dissolve the limestones and dilute hydrochloric acid to dissolve the dolostones. He could identify in the residues all the minerals he had found in the terra rossa, including gibbsite and diaspore, and also what he thought to be sporogelite. He then drew the somewhat astonishing conclusion, strongly questionable even on the evidence of his own analyses, that terra rossa agrees with bauxite in all its physical, chemical and mineralogical properties. Tučan's investigations were based on microscopy and on chemical analyses. The X-ray, thermal and other physical methods of mineralogical analysis were not yet available at the time. This is the explanation for his overrating the role of the amorphous components and for his failure to recognise any clay minerals.

De Weisse's monograph (1948) devotes a separate chapter to terra rossa. The samples studied by him were also taken mainly from the Yugoslav karst. In addition to the petrographic features of terra rossa already referred to, de Weisse emphasised the extreme fineness of the particles constituting it, its hygroscopicity and its high plasticity when wet. He presented chemical analyses of 11 samples (table 43), which resemble those published by Tučan: they are much higher in silica and much lower in aluminium, iron and titanium than the bauxites. According to de Weisse, the composition of terra rossa can be characterised adequately in terms of the *Ki* ratio proposed by Harrasowitz (cf Section 4.1.1):

dissolution residues of limestones	$Ki > 2$
pure kaolinite	$Ki = 2$
terra rossa	$Ki = 0{\cdot}5$ to 2
bauxites	$Ki < 0{\cdot}5$

The *Ki* parameters of some terre rosse may be above or below of this range, but the majority fall within it. De Weisse found the acidic dissolution residues of the limestones in the region invariably containing a higher *Ki* parameter than terra rossa, implying a higher silica and/or a lower alumina content. There is according to him a gradual transition in composition from the dissolution residues of limestone through terra rossa to bauxite. Starting from this statement, he called terre rosse the parts richer in clay of the bauxite body in his description of several deposits. This generalisation is held to be unjustified by the present writer, on the grounds that the mineral composition of terra rossa is fundamentally different from that of the bauxitic clays and

Table 43. Chemical composition of terra rossa samples from Yugoslavia according to de Weisse (1948).

Sampling locality	Al_2O_3	SiO_2	Fe_2O_3	TiO_2	MnO·	CaO	Alk.	L.O.I.	Total
Javornik	30·26	35·21	13·20	n.a.	1·50	0·72	n.a.	19·15	100·04
Karst	20·46	60·44	6·41	n.a.	1·21	0·69	n.a.	7·37	96·58
Zlobin	39·19	26·20	14·03	n.a.	–	–	n.a.	18·14	97·51
Karlobag	32·89	35·42	15·03	n.a.	–	0·43	n.a.	15·32	99·09
Podgor	9·89	52·95	5·73	n.a.	1·41	1·52	3.34	24·80	99·64
Vranić (Hercegovina)	31·05	46·82	11·63	1·04	0·54	–	n.a.	9·46	100·54
Vranić (Hercegovina)	35·85	40·76	10·37	0·90	0·60	–	n.a.	12·12	100·60
Vranić (Hercegovina)	35·61	38·65	13·03	1·49	0·62	–	n.a.	12·22	101·62
Batim (Hercegovina)	35·25	32·13	14·92	1·90	0·68	–	n.a.	14·68	99·56
Nenadić (Hercegovina)	41·41	23·17	15·45	1·94	0·77	–	n.a.	18·03	100·77
Nenadić (Hercegovina)	45·54	21·08	16·42	2·57	0·79	–	n.a.	14·39	100·79

n.a. = not analysed.

Table 44. Chemical and mineralogical composition of terra rossa from Kephallinia Island, Greece, according to Weinmann (1964).

Position of sample in standard soil profile	Al_2O_3	SiO_2	Fe_2O_3	Mineral composition
A	28·09	39·94	11·52	Illite > Vermiculite > Kaolinite, Attapulgite Quartz > Montmorillonite > Chlorite
B_1	29·71	41·71	11·32	Illite > Vermiculite > Kaolinite, Attapulgite Quartz, Montmorillonite > Chlorite
B_2	30·57	41·99	11·44	Illite > Vermiculite > Kaolinite, Attapulgite Quartz, Montmorillonite > Chlorite

clayey bauxites of the bauxite deposits: this is why the present writer deprecates the use of the term terra rossa as a designation of those. De Weisse has published no data concerning the mineralogical composition of terra rossa.

According to Niggli (1952), the most frequent chemical composition of Mediterranean terra rossa is as follows: Al_2O_3, 10 to 40%; SiO_2, 20 to 60%; Fe_2O_3, 5 to 17%; TiO_2, 1 to 3%; loss on ignition, 7 to 20%. He states the most frequent SiO_2/Al_2O_3 ratio to be 1·5 or higher.

Vadász (1956) pointed out that many terre rosse contain bauxite pebbles. Skorić (1962) by an X-ray analysis of twenty terra rossa samples from Yugoslavia established most of those to contain significant quantities of clay minerals besides quartz. Illite, kaolinite and chlorite are frequent; halloysite is found less often. In some samples, he even detected some gibbsite.

Weinmann (1964) studied the terre rosse of Kephallinia Island in the Ionian Sea, but he unfortunately confined his examinations to the chemistry and mineralogy of the −0·002 mm fraction (table 44). This terra rossa contains no free aluminium minerals, at

least not in its finest fraction. Its clay mineral content also differs substantially from what is usual in the bauxites. For example, no vermiculite or attapulgite has been found in any karst bauxite deposit, not even in the most clayey parts of those.

Marić (1964a) from 1964 on submitted the terre rosse of the Yugoslav Karst to a profound study, the findings of which he published in several papers. Table 45 presents his chemical analyses of samples taken at different points of the Karst. He also determined semi-quantitative mineral compositions for four samples by the Debye–Scherrer method, by infra-red absorption and by DTA (table 46). His findings agree well with those of Skorić. In addition, he identified by microscopy the following minerals in most samples: diaspore, muscovite, tourmaline, zircon, rutile, garnet, feldspar. He also found sporadic calcite, dolomite, pyrite, amphibole, epidote, disthene, andalusite, titanite, staurolite and anatase. He examined the acidic dissolution residues of several limestones and identified in them essentially the same minerals as in the terre rosse, albeit in somewhat different abundances.

Subsequently, Marić collected samples of terra rossa from the highest summits of the Velebit and Biokovo Ranges and from several islands off the coast. In these places, the only agent that could introduce even minute amounts of outside material is the wind: the bulk of terra rossa must be a local weathering residue (table 47). The chemical analyses were performed at Zagreb University; the mineralogical analyses were on Marić's request carried out by the present writer at the Geochemical Research Laboratory of the Hungarian Academy of Sciences, using a joint evaluation of X-ray diffractometer and derivatographic recordings (Marić 1968, 1969, 1972).

The heterogeneity of the samples is conspicuous, but their most striking feature is the presence of *nordstrandite* in three samples. It was only in recent years that this metastable aluminium trihydrate phase has been detected in a few places elsewhere. Marić and his co-workers, by a gravity separation procedure applied to samples 1, 2 and 3 from the Velebit Range, found a substantial concentration of Al_2O_3 and SiO_2 and a substantial reduction of Fe_2O_3 in each $-2{\cdot}5$ g/cm^3 fraction. The light fraction contained 86, 85 and 87% nordstrandite plus gibbsite, respectively, the rest being largely illite. The heavy fractions were found to be made up of the minerals of iron and quartz.

Wall *et al* (1962) found the composition of a dark red terra rossa covering a karsted limestone in Western Sarawak, Kalimantan (Borneo) Island, Malaysia, to be as follows: 43% nordstrandite, 31% goethite, 11% quartz, 8% kaolinite, 2·5% pyrolusite and wad, 2% ilmenite, 0·5% diaspore and 0·5% stibiconite. The high nordstrandite content agrees with the findings referred to above.

Marić (1968) attributed the presence of nordstrandite and gibbsite in terra rossa to the fact that, on the superficial weathering of the limestone, some of its aluminium content went into an ionic or colloidal solution. The solutions slowly precipitated their dissolved substances while seeping through the terra rossa, at surface temperature and pressure and at a pH of about 8·5. Under such conditions, a mixture of gibbsite and nordstrandite was formed. Boehmite was formed in turn out of nordstrandite and gibbsite, a process for which strong insolation in the summer seasons provided a sufficient energy input. These processes took place in a comparatively water-poor environment. The iron minerals went their own way, concentration into minute ooids. This is why the gravity fractionation of terra rossa is made possible.

Table 45. Chemical composition of terra rossa samples from Yugoslavia, according to Marić (1964a).

Sampling locality	Al_2O_3	SiO_2	Fe_2O_3	TiO_2	MnO	CaO	MgO	Na_2O	K_2O	P_2O_5	SO_3	CO_2	$\pm H_2O$	Total
Jelenje	24·38	47·89	12·63	0·96	1·18	0·58	n. a.	n. a.	0·32	n. a.	n. a.	0·39	11·86	100·19
Ogulin	37·88	25·87	14·28	1·75	n. a.	n. a.	n. a.	n. a.	n. a.	n. a.	n. a.	n. a.	20·21	99·99
Zadar	23·28	50·41	8·60	n. a.	n. a.	1·78	1·22	0·61	1·09	0·11	0·04	0·10	11·46	98·70
Šibenik	19·99	49·66	6·92	0·92	0·25	1·05	1·22	n. a.	1·19	n. a.	n. a.	n. a.	19·31	100·51
Čitluk	16·47	50·10	7·36	n. a.	n. a.	1·32	0·54	0·58	1·32	0·17	0·01	n. a.	11·54	89·41
Budva	16·72	56·06	7·36	n. a.	n. a.	1·16	0·72	1·12	1·48	0·22	0·02	n. a.	14·54	99·40
Lin (Albania)	25·34	38·95	12·90	n. a.	n. a.	2·26	1·63	0·52	1·06	0·11	0·13	n. a.	16·28	99·18
Koprivno (Split)	20·02	52·02	5·60	0·72	1·10	1·12	0·88	0·50	0·75	n. a.	n. a.	n. a.	17·20	99·91
Koprivno (Split)	25·58	48·86	10·63	0·90	0·49	0·89	0·64	1·45	1·09	n. a.	n. a.	n. a.	15·43	105·96
Crikvenica	21·11	49·79	8·89	0·58	0·08	1·65	1·49	0·36	1·39	0·17	n. a.	1·20	13·61	100·32
Balići (Istra)	26·46	40·78	11·57	1·05	n. a.	0·17	0·31	1·70	0·87	n. a.	n. a.	n. a.	9·25	92·16
Golestrana (Ogulin)														

n. a. = not analysed.

Table 46. Mineral composition of terra rossa samples from Yugoslavia, according to Marić (1964a).

Sampling locality	Quartz	Kaolinite	Illite	Chlorite	Gibbsite	Boehmite	Goethite	X-ray amorphous phase
Crikvenica	40–50	15–20	<5	trace (?)	<5	–	<5	trace
Balići (Istra)	20–30	20–30	<5	20–30	?	<5	15–20	10–20
Koprivno (Split) No.1	40–50	20–30	5–10	5–10	<5	–	5–10	20–30
Koprivno (Split) No.2	15–20	20–30	<5	<5	20–30	–	5–10	–

Table 47. Chemical and mineralogical composition of terra rossa samples from Yugoslavia, according to Marić (1972).

	1	2	3	4	5	6	7	8	9	10
Chemical composition										
Al_2O_3	22·31	21·66	19·53		40·75	17·66	23·03	25·22	21·00	21·63
SiO_2	38·61	50·74	48·77		23·65	49·47	46·93	49·85	55·05	54·93
Fe_2O_3	12·87	8·82	10·62		6·50	10·40	2·39	4·02	7·53	7·55
FeO	n. a.	n. a.	n. a.		0·30	0·60	0·75	0·85	n. a.	n. a.
TiO_2	1·44	1·03	1·05		1·74	1·04	1·09	1·04	1·10	1·14
MnO	n. a.	n. a.	n. a.		0·33	0·17	0·15	0·16	0·14	0·15
CaO	1·25	0·36	0·58		1·30	1·40	3·11	1·40	0·47	0·74
MgO	1·57	0·77	2·00		0·75	1·95	2·74	1·28	1·99	1·42
Na_2O	0·57	0·19	0·27		1·81	1·07	1·42	0·78	1·34	1·35
K_2O	1·71	1·51	2·26		1·73	2·50	2·12	2·02	2·36	2·33
P_2O_5	n. a.	n. a.	n. a.		0·23	0·27	0·40	0·22	0·28	0·05
$+H_2O$	14·48	10·19	10·38		18·45	9·22	12·66	9·01	7·24	7·02
$-H_2O$	5·12	4·64	4·38		2·42	4·36	3·60	4·04	1·01	1·27
Total:	99·93	99·91	99·84		99·96	100·11	100·39	99·89	99·51	99·58
Mineral composition										
Gibbsite	4·2	5·0		3·0	36·6	–	–	–	–	–
Nordstrandite	9·0	5·5		–	–	1·1	–	–	–	–
Boehmite	–	–		–	1·0	–	–	–	–	–
Kaolinite (+sudoite)	17·2	17·4		18·0	30·2	18·2	32·6	32·9	–	–
Trioctahedral chlorite	4·5	3·7		3·2	3·6	11·0	12·5	11·1	–	–
Illite	29·9	26·7		34·3	14·3	32·0	31·0	29·5	–	–
Plagioclase	6·8	4·7		4·0	3·1	9·0	7·0	7·6	–	–
Quartz	11·4	24·9		25·8	–	14·6	11·2	12·8	–	–
Haematite	–	–		–	–	–	0·5	3·0	–	–
Goethite	5·6	5·0		4·0	7·8	2·4	2·3	1·2	–	–
X-ray amorphous limonite	8·8	4·5		5·5	–	9·3	–	–	–	–
Anatase	0·8	0·6		1·6	1·7	1·0	1·1	1·0	–	–
Rutile	0·7	0·5		–	–	–	–	–	–	–
Calcite	–	–		–	1·2	0·4	0·6	–	–	–
Organic matter	1·0	1·4		0·4	0·1	0·2	0·3	0·3	–	–
Total:	99·9	99·9		99·8	99·6	99·2	99·1	99·4	–	–

Column headings:
1. Velebit Range, Struga Summit.
2. Velebit Range, Vaganski Vrh.
3. Velebit Range, Mahoran summit.
4. Slunj (north of Plitvica Lakes).
5. Velebit Range, Baričević dolac.
6. Biokovo summit above Makarska.
7. Grab-Ubline on Hvar Island.
8. Brošica on Murter Island.
9. Brošica on Murter Island.
10. Brošica on Murter Island.

Table 48. Chemical and mineral composition of terra rossa and of bauxite pebbles from Aggtelek and Jósvafő, Northeast Hungary, after Bárdossy (1968b).

	1	2	3	4	5	6	7	8	9	10	11	12
Chemical composition												
Al_2O_3	30·18	19·85	20·80	19·75	20·63	18·59	30·04	29·05	29·25	24·79	16·22	13·57
SiO_2	40·37	55·38	49·50	54·68	55·27	67·81	4·27	4·25	4·46	7·11	12·29	11·72
Fe_2O_3	11·04	7·65	8·45	6·98	7·78	2·37	45·49	48·00	42·07	47·34	58·36	48·22
FeO	n. a.	n. a.	n. a.	n. a.	n. a.	0·26	0·17	n. a.	n. a.	n. a.	n. a.	0·24
TiO_2	0·67	0·94	0·67	0·75	0·84	0·38	0·30	0·95	0·94	0·76	1·16	0·51
MnO	n. a.	n. a.	n. a.	n. a.	n. a.	0·09	0·11	n. a.	n. a.	n. a.	n. a.	6·12
CaO	n. a.	n. a.	n. a.	n. a.	n. a.	0·48	0·24	n. a.	n. a.	n. a.	n. a.	0·35
MgO	n. a.	n. a.	n. a.	n. a.	n. a.	0·99	0·11	n. a.	n. a.	n. a.	n. a.	0·25
Na_2O	n. a.	n. a.	n. a.	n. a.	n. a.	0·12	0·03	n. a.	n. a.	n. a.	n. a.	0·04
K_2O	n. a.	n. a.	n. a.	n. a.	n. a.	2·60	0·10	n. a.	n. a.	n. a.	n. a.	0·32
P_2O_5	n. a.	n. a.	n. a.	n. a.	n. a.	0·06	0·29	n. a.	n. a.	n. a.	n. a.	0·62
CO_2	n. a.	n. a.	n. a.	n. a.	n. a.	0·00	0·00	n. a.	n. a.	n. a.	n. a.	0·00
$+H_2O$	15·08	11·80	13·77	12·75	11·57	4·81	18·19	18·06	22·32	19·05	11·38	16·41
$-H_2O$	2·67	3·19	3·12	2·91	3·04	1·08	1·02	0·75	0·56	0·61	1·27	1·37
Total:	100·01	98·81	96·31	97·82	99·13	99·64	100·36	101·06	99·60	99·66	100·68	99·74
Mineral composition												
Gibbsite	0·8	–	1·5	–	–	–	45·0	40·8	43·5	35·3	22·0	17·5
Boehmite	–	–	–	–	–	–	–	1·2	–	–	–	–
Haematite	5·5	–	1·0	1·0	1·0	–	25·1	35·1	2·0	4·0	39·2	3·0
Goethite	5·4	7·0	6·5	6·0	6·8	1·9	23·2	14·2	45·3	47·8	21·0	53·1
Kaolinite	69·4	33·0	25·8	8·0	35·3	16·2	2·0	4·0	2·0	3·0	7·0	13·0
Chlorite	5·3	15·4	18·9	29·0	10·2	5·7	2·8	2·8	3·6	3·8	3·6	1·0
Illite	6·0	10·0	17·9	21·0	13·4	32·5	–	–	–	3·6	4·0	3·4
Quartz	5·7	30·9	22·8	28·0	28·4	40·9	1·0	0·8	1·4	1·2	1·0	1·0
Plagioclase	–	1·0	2·0	3·0	1·6	1·7	–	–	–	–	–	–
Calcite	–	–	4·1	3·0	2·0	–	–	–	–	–	–	–
Lithiophorite	–	–	–	–	–	–	–	–	–	–	–	6·9
Anatase	0·3	0·4	0·3	0·4	0·4	0·2	0·2	0·6	0·5	0·5	0·5	0·4
Rutile	0·4	0·5	0·4	0·4	0·4	0·2	0·1	0·4	0·4	0·3	0·5	0·1
Total:	98·8	98·2	101·2	99·8	99·5	99·3	99·4	99·9	98·7	99·5	98·8	99·4

Column headings:
n. a. = no analysis.
1. Terra rossa, reddish-brown, Jósvafő, Red Lake.
2–5. Terra rossa, reddish-brown, Aggtelek, Yellow Lake.
6. Terra rossa, yellowish-brown, Aggtelek, Yellow Lake.
7–12. Iron-rich bauxite pebbles, dark rust-red, Jósvafő, Red Lake.

Terra rossa on Gömör County karst, Hungary, locally contains some slightly worn iron-rich bauxite pebbles of 0·5 to 20 cm in size (Bárdossy 1968b). These are much harder than the terra rossa proper. Chemical and mineralogical analyses by the writer (table 48) revealed the terra rossa there to contain free aluminium minerals only sporadically and in minute quantities. There is a significant amount of quartz besides kaolinite.

Summarising the results of the studies so far performed, the following conclusions may be drawn.

(i) Terra rossa cannot, either in the chemical or in the mineralogical respect, be identified with the clayey bauxites, bauxitic and bauxite-like clays of the Mediterranean bauxite deposits. The principal chemical differences are a higher content of silica and lower contents of aluminium and titanium; terra rossa tends to contain more sodium, calcium and magnesium by factors of 4 to 10 than the bauxitic rocks.

In the mineralogical respect, the difference is even more pronounced. Whereas bauxitic and bauxite-like clays contain quartz and feldspars only to the tune of a tenth or a hundredth of a percent or so, terra rossa usually contains 10 to 30% quartz and 3 to 10% feldspar. The predominant clay mineral of bauxitic and bauxite-like clays is kaolinite. In terra rossa, however, significant quantities of illite and chlorite have been determined besides kaolinite: in fact, some clay minerals have been identified occasionally which are totally unknown in any other karst bauxites (atrapulgite, vermiculite).

(ii) The composition of terra rossa is highly varied from one region to the next, and only the principal features of its composition remain the same.

(iii) The composition of terra rossa differs not only from that of the bauxites, but also from that of all other pelitic sedimentary rocks. It is therefore justified to regard it as a separate entity in both chemical and mineralogical respect. One of its principal features is that it contains a few percent of gibbsite and nordstrandite as a rule.

6.2. THE ORIGIN OF TERRA ROSSA

The first to devote a profound study to the formation of terra rossa was Neumayr (1875), who stated it to be the weathering product, the insoluble residue, of limestone. A similar view was held by Fuchs (1875), according to whom the terre rosse of the Mediterranean region had formed in the Tertiary, under a climate hotter than our present one.

An essay, very modern and pioneering for its time, was written by Krenner (1876) on the origin of the Hungarian terre rosse. He pointed out that terra rossa is almost ubiquitous on the surfaces of carbonate rocks around the Mediterranean, whereas it is missing from the regions farther to the north—e.g. in Northern France and Belgium—, although there are exposures of carbonate rocks of identical composition there. From this fact he drew the correct conclusion that the distribution of terra rossa is a function of climate and vegetation: it will form under a comparatively sparse vegetation and a warm climate. No terra rossa will form under temperate, moist climes, where the

limestones are forest-covered. In order to prove that a red or reddish-brown terra rossa can in fact be the dissolution residue of a white limestone, he dissolved a white limestone from Dalmatia in dilute hydrochloric acid. He found the dissolution residue to be red and to contain some 20% Fe_2O_3. The difference in colour, then, is no valid argument against deriving terra rossa from white limestones.

Kramer (1889) held the opinion that terra rossa was not a limestone residue. According to him, the Dalmatian karst was covered by a thick mantle of clays and marls in the Tertiary, the bulk of which has been eroded since. It is the residue of this process which is the terra rossa of today.

Tučan (1912) on the strength of his chemical and micrsocopic examinations referred to the above reached conclusion that terra rossa is the insoluble residue of the surface weathering of limestones and dolostones. He regarded as a proof for his assumption that he could identify the same minerals in the terra rossa as in the dissolution residues of the underlying limestones and dolostones. Tučan was the first to raise the idea of a connection between terra rossa and bauxite. According to him, the two formations are essentially identical: bauxite is an old (fossil) terra rossa, and terra rossa is recent bauxite. Kišpatić (1912) in an essay on the origin of the bauxites of Croatia reached the same conclusions.

De Weisse (1948) also regarded terra rossa as the weathering residue of limestone and dolostone. In addition to the carbonate parent rock, he attributed the principal importance to a Mediterranean–subtropic climate, in which a wet winter season is followed by a hot dry summer. This does not permit the evolution of a lush vegetation, so that humus can form in no great quantities, either. Such conditions entail the enrichment of iron oxides and hydroxides which in the presence of abundant humus would have been removed from the soil in solution. According to de Weisse, it is uncertain whether any terra rossa is being formed today in the Mediterranean region. De Weisse derives the karst bauxites from the desilication of terra rossa.

The idea of terra rossa being the weathering residue of limestones and dolostones was in the subsequent years accepted fairly widely, e.g. by Bonte (1958), Hose (1960) and Jung (1958). On the other hand, Haberfelner (1951) went so far as to reverse the theory developed by Tučan stating as he did that terra rossa is the weathering product of bauxite. The same conclusion was reached also by Vadász (1956) who in support of his idea adduced the observation that, in the Mediterranean region, the geographic distribution of terra rossa coincides by and large with that of the bauxites: wherever there is no bauxite, there is no terra rossa, either. He regarded as another argument in favour of his assumption that many terre rosse contain pebbles of bauxite. This would mean that terra rossa was a surface deposit attached to the bauxites rather than a climatically controlled formation.

Zans (1959) likewise denied the origin of terra rossa as a weathering product of limestone. He regarded it as an allochthonous deposit, the product of the weathering and partial desilication of alumosilicate rocks. He thus placed it in an intermediate position between an alumosilicate parent rock and a bauxite.

According to Miholić (1960), terra rossa, in bulk is the weathering product of marls and clays, transitional between those and bauxite. According to him, terra rossa can be turned into bauxite by desilification, and bauxite can be turned into terra rossa by resilification.

Wall *et al* (1962), state the terra rossa on Kalimantan (Borneo) Island to be a product of limestones weathering, with contributions from the weathering residues of nearby volcanics and shales. According to Weinmann (1964), terra rossa was produced *in situ* out of the underlying limestones on Kephallinia Island in the Ionian. He assumed the process to have started in the Tertiary and to go on still at altitudes below 800 m. According to Marić (1964) too, the formation of terra rossa is still going on in the Yugoslav karst region of today. He derives the bulk of terra rossa from the underlying carbonate rocks, but admits minor contributions by the weathering products of nearby outcrops of Palaeozoic alumosilicate rocks. He found 0·1 to 0·3 m of terra rossa even on the highest summits of the Karst and also on the Dalmatian islands, where no other alien matter than wind-blown dust could have been added to the *in situ* weathering products.

On the basis of the information available in literature and of his own field observations, the present writer has formed the following opinion concerning the origin of terra rossa.

(i) The formation of terra rossa as an *in situ* weathering product in isolated places—on mountain summits and islands—can be regarded as an established fact. These may contain no alien material other than wind-borne dust.

(ii) In all other places, terra rossa is likely to be composed in different proportions of the weathering products of the underlying limestones, dolostones and of nearby alumosilicate rocks, and of wind-blown material. Wherever there were bauxite outcrops near by, the fine weathering products of those and pebbles of smaller or larger size derived from them may also have contributed to the terra rossa. The proportions of these parent materials vary widely from region to region. This diversity of parent materials is the reason for the varied chemical and mineralogical composition of terra rossa.

(iii) The prime conditions for terra rossa formation are a suitable climate and the peculiar physico-chemical environment and plant cover determined by the carbonate rocks.

(iv) Wherever these conditions are satisfield, terra rossa did form in the past: wherever they are satisfied in the present, terra rossa goes on forming. There is no fundamental difference between the terre rosse of today and those of the Tertiary. No evidence of satisfactory reliability is available to decide whether any formation of terra rossa also took place in the Mesozoic and Paleozoic.

We shall discuss the genetic relationship between bauxite and terra rossa in the chapter on bauxite genesis.

7. STRUCTURAL POSITION

7.1. GEOTECTONIC POSITION OF THE BAUXITE REGIONS

7.1.1. SPATIAL POSITION

Tectonic literature subdivides the Earth's crust into first-, second-, third- and fourth-order geotectonic units. Structural geologists' opinions differ rather widely as to the details of this subdivision. The classification used by the writer is based on the principles of plate tectonics, taking into account all the geophysical evidence as far as possible. The geotectonic units to which the individual bauxite deposit groups belong according to the writer are indicated in Appendix 1.

The *first-order geotectonic units* are the continental platforms, the oceanic basins and the orogenic belts. The writer has calculated the tonnage distribution of bauxites over these three types of geotectonic unit (table 49).

Table 49. Distribution of bauxites on first order structural units (in percent).

First order structural units	Lateritic bauxites	Karst bauxites	Tikhvin-type bauxites
Continental plates	96	8	100
Orogenic belts	3	92	–
Oceanic basins	<1	–	–
	100	100	100

The geotectonic control is quite obvious: the bulk of karst bauxites is found in orogenic belts, whereas the bulk of laterite bauxites and all the Tikhvin-type bauxites are on continental platforms. Bauxite in the oceanic basin areas is confined to a few islands, e.g. Hawaii. In the writer's opinion, geotectonic control is one of the principal reasons for the differences in the composition, constitution and origin of karst and laterite bauxites.

First-order geotectonic units subdivide into second-order ones. On the *continental platforms*, one distinguishes ancient shields, younger platforms covered with sediments and volcanics, and *mobile paraplatforms*. The latter are much less extensive than the two former. The tonnage distribution of bauxite over these second-order units, as calculated by the writer, is presented in table 50.

Laterite bauxites are distributed about evenly between ancient shields and platforms; they are less frequent on paraplatforms. The bulk of the karst bauxites deposited in continental areas, on the other hand, is situated on paraplatforms. By

Table 50. Distribution of bauxites on continental plates (in percent).

Second order structural units	Lateritic bauxites	Karst bauxites	Tikhvin-type bauxites
Ancient shields	45	–	–
Platforms	45	2	20
Paraplatforms	10	98	80
	100	100	100

virtue of their enhanced mobility, paraplatforms constitute, as it is well known, a sort of transition between the orogens and the true extensive continental platforms made up of the ancient shields and of the platforms. We have seen the bulk of karst bauxites to be connected with the orogenic belts: it now becomes apparent that the karst bauxites found on the continental platforms in their turn concentrate on the paraplatforms, those second-order geotectonic units closest to the orogens. This underscores even more the connection between karst bauxites and the mobile zones of the Earth's crust. Remarkably, the bulk of Tikhvin-type bauxites is also found in paraplatform regions (e.g. on the Northern China and Southern China paraplatforms).

Within the *orogenic belts*, the following second-order units are distinguished: intracontinental, Andean, Alpine and island-arc orogenic belt. The percentage distribution of bauxite tonnage over these second-order units is presented in table 51.

Table 51. Distribution of bauxites on orogenic belts (in percent).

Second order structural units	Lateritic bauxites	Karst bauxites	Tikhvin-type bauxites
Intracontinental belts	–	9	–
Andean-type belts	46	–	–
Alpine-type belts	17	47	–
Island-arc belts	37	44	–
	100	100	–

Intracontinental orogenic belts lie within continental plates, or between older orogens of prior consolidation. This type of orogen is of comparatively small areal extent: this is probably why its bauxite tonnage is so low comparatively. All the bauxites known from this type of geotectonic unit are karst bauxites (in the orogenic zone of the Pyrenees and the Provence).

On the margins of the continental plates, there are the so-called pericontinental orogens: they are subdivided into Andean and Alpine ones by Aubouin (1972). In the Andean or liminar orogenic belts, both the eugeosynclinal and the miogeosynclinal zone rests on a continental type crust. They do not include any flysch or ophiolites, but abound in continental molasse sediments and andesitic volcanics. The most typical examples include the Andes and the Cordilleras. No karst bauxites have been found in these orogenic belts so far, although there are minor lateritic bauxite deposits in several places (e.g. Oregon, US; Honduras, Costa Rica, Panama, Columbia, Chile; the latter near Santiago de Chile).

In the Alpine type orogenic belts, the eugeosynclinal zone rests on an oceanic crust, and the miogeosynclinal zone on a continental crust. Ophiolites, flysch and carbonate rocks abound. Coarse clastic sediments are more subordinate. Almost half of all karst bauxites was formed in Alpine type orogenic zones, in the Mediterranean region above all. Lateritic bauxites are few and far between (e.g. Central Timan Mountains, USSR; Taurus Range, Turkey; Southeast Australia and Tasmania).

Island-arc orogens differ from the above ones in that both their eugeosynclinal and miogeosynclinal zones rest on an oceanic crust. The first to point out the difference between orogens on the continental margins and in island-arc regions were Wilson and Burke (1972). The main features of island-arc orogens were summarised by Coleman (1975). Island-arc orogens are made up largely of volcanics, but carbonate rocks are also frequent. The tonnage of karst bauxites formed there is almost as great as in the Alpine-type orogens, especially in the Caribbean and in the island arcs bordering on the Pacific from the southwest. There are minor occurrences of lateritic bauxite in Malaysia (on both the Malaysian Peninsula and in Sarawak on Kalimantan Island, Borneo), in Indonesia, in the Philippines, on the Palau Islands and on New Zealand.

To summarise, karst bauxites are most abundant in the Alpine type and island-arc orogens. Much smaller tonnages of lateritic bauxite are found especially in the Andean and island-arc orogens. The reasons for this distribution are to be discussed in the section on bauxite origin.

In the *oceanic basin regions* bauxites are so scarce that their description by second- and third-order geotectonic units is deemed superfluous.

The *third-order geotectonic units* are not as a rule sharply distinct: there are zones of transition between them. The writer has accordingly refrained from an evaluation of these with respect to tonnage, confining the discussion here to a description of their principal features.

In the *ancient shield regions*, the writer distinguishes, after Moorbath (1977), ancient granito–gneiss intrusions, folded metamorphic zones, and younger intrusions. The large lateritic bauxite deposits tend to concentrate in the ancient metamorphic zones, e.g. in the Eastern Ghats of India, on the West African Shield, and on the Guyana Shield. Some younger intrusions may consist of rocks especially rich in aluminium, e.g. nepheline syenites. Lateritic bauxites unusually rich in aluminium and low in iron were formed on these: e.g. Poços de Caldas and Serra do Mantiqueira in Brazil; Los Islands of Guinea.

On the *platforms*, Bogdanov (1968), Khain (1972) and Muratov (1972) distinguish anteclines, synclines, aulacogens and rift valleys. There are extensive lateritic bauxite deposits on the anticlines and on the gentle flexures inserted between the anticlines and the synclines. It is here that one finds those few deposits of karst bauxites which had formed in platform areas (e.g. Maktesh Ramon, Israel). No bauxite is known at all from the synclines, aulacogens and rift valleys. In places, extensive synclines are filled with basalt flows from rift volcanoes. There are also extensive lateritic bauxite deposits on the surfaces of these plateau basalts, e.g. on the Deccan Plateau in India or in Cameroun.

In the *paraplatform areas*, outcrops of crystalline Precambrian and sediment-covered regions can be distinguished: they are the less extensive and more mobile counterparts of the ancient shields and of the platforms, respectively. All the karst

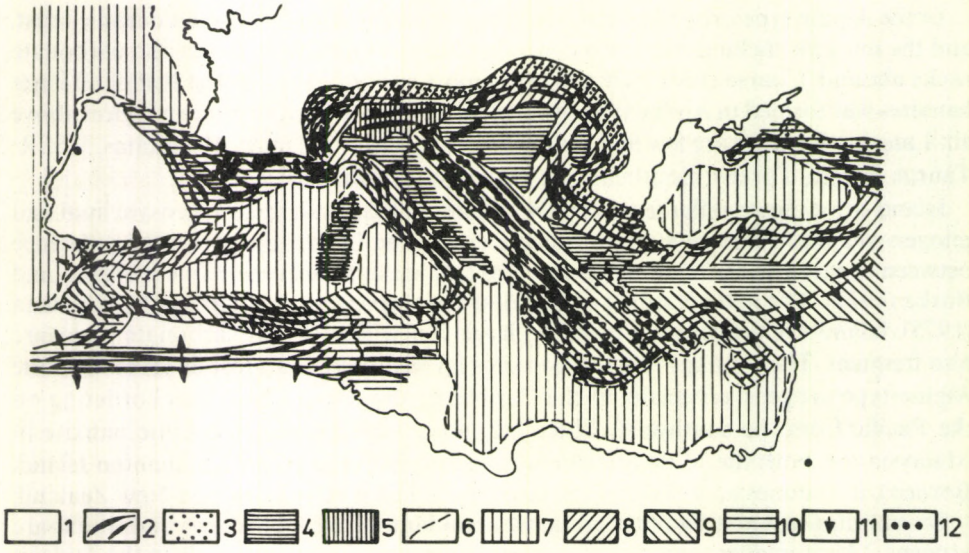


Figure 166. Geotectonic map sketch of the Mediterranean region. After Aubouin and Durand-Delga (1971)
(1) continental platforms and pre-Alpidic orogens; (2) strikes of orogens; (3) foredeeps; (4) interarc basins and uplifted median masses; (5) eastern alps; (6) 1000-m isobath in the seas; (7) sea deeps beneath the 1000-m isobath; (8) northern arc system of alpine orogen; (9) southern arc system of Alpine orogen; (10) intracontinental orogens; (11) vergency; (12) bauxite deposit groups

bauxite deposits of the paraplatforms are in the sediment-covered regions. On the margins of the Precambrian regions and in their proximity there are Tikhvin-type bauxite deposits.

In the *intracontinental orogenic belts*, one distinguishes an axial zone and two symmetrically disposed marginal zones. There are karst bauxites in these latter, e.g. on the French and Spanish slopes of the Pyrenees. No lateritic bauxites have been identified in any intracontinental orogen.

In the *Andean-type orogenic belts* eu- and miogeosynclinal zones, intermountain basins and foredeeps are distinguished. All lateritic bauxites known in these belts are found in the eugeosynclinal zones. Karst bauxites are not known at all.

In the *Alpine-type orogenic belts*, eugeosynclinal and miogeosynclinal zones, median basins, median masses and foredeeps are usually distinguished. The median basins have recently been renamed interarc basins of continental type. Bauxite had formed in about equal tonnages in the eugeosynclinal and miogeosynclinal zones, and in much smaller quantities in the interarc basins. No bauxite has so far been identified in any of the foredeeps. The bulk of lateritic bauxites in the Alpine orogens is found in the eugeosynclinal zones.

The bulk of the karst bauxites of the world is concentrated in the Alpine orogenic belts of the Mediterranean region. It is usual to subdivide these orogens after Aubouin and Duran-Delga (1971) into a double system of ranges. There is a southern system comprising the Taurids, Hellenids, Dinarids, East Alps, Appennines and the Maghrebids, and a northern one comprising the Betic Cordillera, the West Alps, the Carpathians, the Balkan Ranges and the Pontids (fig. 166). In both zones, vergencies

are directed outward, towards the continental plates; the miogeosynclinal zones are on the outward side, the eugeosynclinal zones on the inward side. The writer has placed the Eastern Alps into the southern system of arcs because, according to the most recent evaluations, these have overridden the original Alpine range (the West Alps) from the south to the north (Trümpy 1960).

The bulk of karst bauxites is found in the southern system of arcs, in its central sector extending from the Appennines to the Western Taurids. In the northern arc system, there are just a few scattered minor bauxite deposit groups. The writer attributes this arrangement to causes connected with plate tectonics: at the time of bauxite formation, the two arc systems were still far apart, on the northern and southern shores, respectively, of the Tethys, and their climate must have been as different as is implied by this situation. Of the interarc basins and median masses, the only one to have major bauxite deposit groups is the Pannonian mass.

In the *island-arc orogenic belts*, eu- and miogeosynclinal zones and incterarc basins can also be distinguished. The bulk of karst bauxites in this type of zone is in the eugeosynclinal zones (e.g. Jamaica and Hispaniola). There are also some lateritic bauxite deposits in the eugeosynclinal zone, and fewer ones in the miogeosynclinal zone. No bauxites are known from the interarc basin regions.

7.1.2. TEMPORAL POSITION

On the *ancient shields*, bauxite is invariably on the surface, as the product of a post-Upper Cretaceous lateritic weathering.

In the *platform regions*, it is usual to distinguish a temporal sequence made up of a Precambrian crystalline basement and of a sedimentary–volcanic epeirogenic cover succession. Most lateritic bauxite deposits are on the surface also here, and are also post-Upper Cretaceous. The only lateritic bauxite deposit groups to lie at the base of an epeirogenic succession are those of the Ukraine (Bielgorod, Vysokopolsk). They were formed in the Lower Carboniferous. The Tikhvin-type deposits on the Russian Platform and on the rim of the Baltic shield are of similar age; they are likewise covered by an undisturbed epeirogenic succession. All the karst bauxites found in platform regions are intercalated in the epeirogenic successions, attached in time and space to emergences of limited extent (e.g. Missouri State and Western Coastal Plains, US; Maktesh Ramon, Israel; Chadobetsk, USSR).

In the *paraplatform regions*, the bulk of bauxite is found in sedimentary successions overlying the crystalline Precambrian, typically in several horizons one above another. In the Northern and Southern Chinese paraplatforms, there are seven successive bauxite horizons, from the Lower Carboniferous to the Middle Permian. The lowermost horizon is a karst bauxite; all the others are of Tikhvin type. Each was formed during a brief episode of emergence. It is the greater mobility, already referred to, of the paraplatforms that made possible this repeated bauxite formation.

In the *orogenic belts*, structural geologists distinguish several cycles of orogeny. After Muratov (1972), the present writer regards the Baikalian as the first cycle which gave rise to an independent orogen. There were older cycles also, but the role of these was in making up the continental plates, and that is where the structural geologists place

them. In the following, then, let us evaluate the Baikalian, Caledonian, Hercynian and Alpine cycles of orogeny. The writer has lumped together the orogenic cycles distinguished by some Soviet authors under the names 'Mesozoic' and 'Kamchatkan' with the Alpine orogeny.

The term *orogenic succession* is used to denote the beds formed during a single orogenic cycle, regardless of the fact that any orogenic cycle is an alternation of long epeirogenic periods with brief spasms of orogeny. The succession deposited on a consolidated part on the terrestrial crust subsequent to the folding and erosion of an orogen is called a *postorogenic succession*. It is called more specifically an epi-Baikalian, epi-Caledonian or epi-Hercynian succession, depending on the orogenic cycle that it follows. The evolution of the Alpine orogenic cycle has not as yet come to an end, so that there is, properly speaking, no epi-Alpine post-orogenic succession.

As we have seen it above 92% of all karst bauxites is found in orogenic belts. Equating the totality of these bauxites to 100%, we find that 85% of that total is in orogenic successions and only 15% in post-orogenic ones. Of the latter group, 4% is found in epi-Baikalian, 3% in epi-Caledonian, and 8% in epi-Hercynian successions.

The distribution of karst bauxites formed within orogenic successions is as follows: 0% in Baikalian successions, 1% in Caledonian, 14% in Hercynian, and 70% in Alpine successions. That is, bauxite formation is the more widespread, the younger is the succession. This increase of abundance is a significant feature even with the greater likelihood of erosion of the older bauxite deposits taken into account.

There is as yet little consensus among structural geologists as to the delimitation of the phases of evolution within the orogenic cycles. Aubouin (1972) for one distinguishes geosynclinal and tardigeosynclinal phases. Others subdivide the evolution of a geosyncline into a pre-inversion and a post-inversion phase. Still others regard as an orogenic phase in the strict sense of the term only the time span between the beginning and the end of the phase of principal foldings and overthrustings. In Soviet literature, a subdivision into so-called structural stages is resorted to with preference (e.g. palaeo-Alpine, meso-Alpine and neo-Alpine structural stages). Evaluation is hampered by the comparatively large number of orogenic phases and their tendency to shift to and for in time and space within a given orogen. In view of these incertitudes, the writer has undertaken no detailed analysis of units smaller than the orogenic cycles, although he regards such an evaluation as one of the most important tasks of the near future.

It goes without saying that bauxite deposits were formed during the more quieter phases of epeirogenic nature preceding or succeeding a phase of orogeny rather than in the phases of orogeny proper.

As it was shown in table 49, no more than 3% of all lateritic bauxites is found in orogens. The bulk of these is associated with the Alpine orogeny. They are Plio–Pleistocene or Neogene in their majority, whereas the deposits of Southeastern Australia and Tasmania are Eocene (Patterson 1967), and the laterites of Santiago de Chile are Senonian (Romero 1972). The iron-rich lateritic bauxites in the Taurus Range probably came to exist in the Dogger and in the Lower Malm.

Lateritic bauxites associated with the Hercynian orogen are even more sporadic. Only their eroded remnants have been identified in the Urals (Bushinsky 1971, 2nd ed. 1975). In recent years, partly reworked lateritic bauxite deposits of Middle and Upper

Devonian age have been found in the Central Timan Mountains (Kirpal' 1977). No laterite bauxites are known from any of the Caledonian or Baikalian orogenic successions.

7.1.3. RELATIONSHIP BETWEEN TYPE OF DEPOSIT AND GEOTECTONIC POSITION IN KARST BAUXITES

Table 52 presents the geotectonic position of the six deposit types of karst bauxites grouped into units of time and space. There is a conspicuous relationship between the type of deposit and geotectonic position.

Table 52. Tectonic position of the main types of karst bauxite deposits (in per cent).

Type of deposit	Tectonic unit	Percent	Time unit (cycle)	Percent
Mediterranean	orogenic belts	100	orogenic	100
Ariège	orogenic belts	100	orogenic	100
Salento	orogenic belts	100	post-orogenic	74
			orogenic	26
Kazakhstan	platforms	5	epeirogenic	5
	orogenic belts	95	post-orogenic	95
Timan	paraplatforms	80	epeirogenic	80
	orogenic belts	20	post-orogenic	20
Tula	platforms	100	epeirogenic	100

Almost all the *Mediterranean-type deposits* are in orogenic belts. The only exception is the Maktesh Ramon deposit group, Israel, but its tonnage does not attain 0·1% of the total Mediterranean-type bauxite tonnage. The rather lapidary description of Slatkine and Heller (1961) makes it questionable whether this deposit group is in fact of Mediterranean type. In the time dimension, almost all the Mediterranean-type deposit groups are found within orogenic successions. The exceptions to this rule comprise just four small deposit groups, together making up less than 0·2% of the total Mediterranean-type bauxite tonnage. The exceptions are Andorinha, Portugal; Northern and Southern Ferghana and Kansay, Soviet Central Asia, USSR; all are situated in post-orogenic successions.

All the *Ariège-type deposits* are in orogenic belts, and were formed during cycles of orogeny.

The *Salento-type deposits* are also all situated in orogenic belts, but owing to their secondary reworking they are rather obviously intercalated in post-orogenic successions in their majority.

It is a common feature of the *Kazakhstan-*, *Timan-* and *Tula-type deposits* that they are invariably found in epeirogenic or post-orogenic successions. If we admit the post-orogenic successions to be also of an essentially epeirogenic nature, an even closer relationship results. The principal differences between the three types are in spatial

position: most of the Kazakhstan-type deposit groups are associated with orogenic belts, and most of the Timan ones with paraplatforms, whereas the Tula-type ones are found on platform areas.

Within the Mediterranean deposit type, there are differences in deposit features, texture and composition, according to whether or not the deposit group falls within this or that third-order geotectonic unit. Let us adduce only the most important examples.

The deposits of the highest average grade are those in the geosynclinal zones. The deposits of Dalmatia and Crna Gora (Montenegro) contain a mere 2 to 15% bauxitic clay: high grade bauxite locally attains 70 to 75% of total bauxite tonnage. In the bauxite deposits of the median masses and interarc basins, on the other hand, bauxitic clay is more abundant, and even some bauxite-like clay may occur. The deposits of the Transdanubian Hills e.g. contain 10 to 30% bauxitic clay and 2 to 10% bauxite-like clay on average, and the abundance of clayey bauxite ranges from 15 to 30%.

Ooidic–pisoidic textures are most frequent in the deposits of the geosynclinal zones, whereas in the deposits of the median masses, microclastic and arenitic textural types are most abundant.

Bauxitic formations are also controlled by their geotectonic position. The pure carbonate subtype of the carbonatic formation is invariably found in geosynclinal zones. The clayey–sandy subtype of the carbonatic formation is found largely in the median masses of the Alpine orogenic belts and in the intracontinental orogens. The coal–clay–carbonate formation is found in its bulk in the epeirogenic sedimentary successions of the paraplatforms and platforms, whereas a minor part constitutes postorogenic successions in consolidated orogens. Finally, almost all the sandy–clayey bauxite formations are in the post-orogenic successions of consolidated orogens. In the final reckoning, then, a close control of bauxite formation and deposit type by geotectonics emerges.

7.2. RELATIONSHIP BETWEEN BAUXITE DEPOSITS AND LOCAL TECTONIC STRUCTURE

It is best to analyse the influence of local tectonic structure in terms of the movements that took place prior to, simultaneously with and subsequent to the formation of bauxite.

(i) *Pre-bauxitic tectonic movements* influence the localisation and form of the deposits. In many deposit groups, the deposits are located along faults and the largest and thickest deposits often coincide with the crossing of two faults. This in the writer's opinion is due to the fact that the carbonate rock was shattered at the crossing and the better drainage resulted in a stronger karstification. It was the karst depressions thus formed that were filled with bauxite later.

Many bauxite deposits fill pre-existing structural grabens: the writer has distinguished these as a separate deposit form within the Mediterranean deposit type (fig. 16). In these deposits the faults played a pre-forming role; the bauxite then filled out the grabens formed by the faults. The role of pre-forming faults in the Transdanubian Hills, Hungary, was assessed by Szantner and Szabó (1970).

(ii) *Syn-bauxitic structural movements* influenced deposit formation and bauxite grade even more. This can be best illustrated by the example of Jamaican bauxites: Hill and Davis (1971) recognised that the bauxite deposits of Jamaica are associated with structural blocks delimited by the principal faults; they distinguished on this basis plateau-type and graben-type deposit groups. The former are of higher and more uniform grade, and are predominantly gibbsitic. The latter are of more varied grade; they locally exhibit traces of reworking. Boehmite is enriched in them as well as gibbsite.

The writer after his own field observations prefers to further refine this basically correct subdivision into deposits of high-plateau, low-plateau and graben-valley type. The high-plateau type includes e.g. the Cockpit Country and the northern part of the Manchester Plateau down to the fault line traversing it, called the Sixteen Mile Gully. To the south of this fault scarp there is a low-plateau type deposit group extending down to the structural scarp along the coast. For examples of the graben-valley type, let us cite the Porus Valley and the Essex Valley. Deposits are doline-type and lenticular on the high plateaux, lenticular and blanket-type on the low plateaux, and stratiform and blanket-type in the graben-valleys.

There is evidence that the movements of these structural blocks took place partly during the period of bauxitisation, and partly after it (Lee 1964).

(iii) The influence of *post-bauxitic structural movements* can be studied best where the bauxite deposit is buried under some cover sediments. The older—Mesozoic and Paleozoic—deposits frequently exhibit traces of repeated structural deformations, each of which may be of a different nature. The original deposit form, the lithology of the bauxite and its chemical and mineralogical composition were modified by these successive structural dislocations the more, the greater was the deformation. In order to characterise the deposits also in this respect, the writer has distinguished the following five degrees of tectonic deformation.

(1) *Undisturbed deposits.* The deposits are horizontal, and exhibit traces neither of faulting, nor of folding.

(2) *Weakly deformed deposits:* the deposit top dips at 10 to 20° at most; the deposit is dissected by few faults: no folding is observed.

(3) *Deposits of medium deformation.* The bauxite top dips at 45° at most; the deposit is cut up into structural blocks by faults, or it may be gently folded.

(4) *Strongly deformed deposits.* Dip greater than 45°, traversed by numerous faults and overthrusts; strongly folded.

(5) *Very strongly deformed deposits:* steeply dipping, sometimes vertical or overturned, with overthrusts, imbrications, or strong folding, e.g. overturned folds.

The degree of structural deformation is indicated in terms of the above five-degree scale in Appendix 1 for each deposit group. If structural deformation is not uniform within a deposit group, then the most widespread degree is indicated.

It is interesting in this context to compare the bauxite deposit groups of the Alpine orogenic belt within the Mediterranean region (table 53).

The table reveals that the overwhelming majority of all the deposit groups in the eu- and miogeosynclinal zones to be strongly or very strongly deformed. Most of the deposit groups in the median masses and basins, and in the forelands attached to them,

Table 53. Structural dislocation of bauxite deposits in the Mediterranean zone.

Third order structural units	Quiet	Dislocated				Total
		Weakly	Medium	Strongly	Very strongly	
Eugeosynclinal belts	–	1	3	9	8	21
Miogeosynclinal belts	–	1	7	28	11	47
Median masses and basins	1	18	1	–	–	20
Altogether	1	20	11	37	19	88

on the other hand, are weakly deformed. The degree of deformation, then, is controlled basically by geotectonic position.

The deposits of strong deformation, and *a fortiori* those of very strong deformation, have undergone substantial changes of form. Imbrication has cut up the deposits and stacked the slices one above another. In strongly folded areas, a structural swelling of the deposits in the axial zones of the anticlines is observed (Kiskyras 1960). Folding combined with overthrusts may have given rise to some incredibly contorted deposit forms of structural sculpture (e.g. in the Eleusis–Mandra bauxite region in Greece).

Structural stress resulted in the compaction of bauxite, in a reduction of its porosity, initiating the recrystallisation of the bauxite minerals from gibbsite through boehmite and diaspore towards corundum. These secondary processes were described in the mineralogical chapter. In the deposits of weak and medium deformation, mineralogical changes are confined to the fault zones. In these zones, segregations of calcite are plentiful. Elsewhere, ground water deposited silica, argillising the bauxite along the fault zone. In the Northern Urals, USSR, segregations of pyrite have been observed along faults.

8. GENETIC ASSESSMENT

8.1. A REVIEW OF THEORIES ON BAUXITE ORIGIN

The basic factors of karst bauxite formation are

— the parent material,
— transport, if any, and its mode,
— the place and environment of deposition,
— the mode of bauxitisation; its climatic, geomorphologic, physico-chemical, biological conditions etc.

Theories on the origin of bauxite can be grouped according to the role, importance and characteristics that they attribute to each of the above factors.

8.1.1. LATERITOGENIC THEORIES

These theories derive karst bauxite from pre-existing laterites, the erosion products of which they assume to have been deposited in their present lies by various modes of transport. The first to formulate such a theory was Berthier (1821), with a view to explaining the origin of the deposits in the French Alpilles. His ideas were subsequently extended by Dieulafait (1881) and Lacroix (1913b) to all the bauxite deposits of Southern France. In recent years, the theory was revived by Nicolas (1970) who found that the bauxites of the Provence were forming by the lateritic weathering products of a variety of local rocks, Aptian and Albian marls in particular. Valeton (1965, 1972) and Erhart (1956, 2nd ed. 1967) held the primary laterites involved to have formed exclusively on the crystalline of the Maures–Esterel Mountains.

In the USSR, N. I. Arkhangelsky (1937) found that the deposits of Kamensk–Uralsk in the Central Urals Mountains were formed from pre-existing laterites. His opinion found no response in a period dominated in the USSR by the chemogenic theory. Since the works of Bushinsky (1958, 1970, 1971), the bauxites of the Urals, of Siberia and Kazakhstan in the USSR are also derived by many Soviet geologists to have been formed from pre-existing laterites.

The bauxites of Central Greece are the products of laterites originally situated in the Locris region according to Aronis (1955), Papastamatiou (1964) and Bárdossy and Mack (1967). Hungarian bauxites were regarded as the lateritic weathering products of igneous and metamorphic rocks by Telegdi-Roth (1927) and Vadász (1951). Zans (1954) derived the Jamaican bauxites from the lateritic weathering products of

andesites and andesite tuffs. Similar opinions have been put forward by Burns (1961) and Lee (1964).

The deposits in the Ariège, France were attributed to the *in situ* lateritization of clays and marls by Combes (1969) who also admitted that a certain role was played by the carbonate bedrock in bauxitisation. This reveals the Ariège type of deposits to be transitional between the laterite bauxites and the karst bauxites in the strict sense.

8.1.2. THE TERRA ROSSA THEORY

According to this theory bauxite is a product of terra rossa, the weathering residue of carbonate rocks older than the bauxite. It was formed *in situ* or underwent a very limited transport.

Dolfuss (1904) was the first to propose this theory to explain the origin of French bauxites. It was subsequently applied to the bauxites of Yugoslavia by Kišpatić (1912) and Tučan (1912), and to the bauxites of the Bihar Mountains by Pauls (1913). In the wake of the impressive oeuvre of Lapparent (1930) the theory has been generally accepted in France. The origin of the Caribbean bauxites was explained in terms of this theory by Butterlin (1958) and Hose (1963). The bulk of the Transdanubian bauxites was attributed to the chemical weathering of carbonate rocks by Komlóssy (1967) and by Vendel *et al* (1971). According to Weisse (1964), all the karst bauxites of the Earth had their origin in terra rossa.

The terra rossa theory in its application to the Yugoslav Karst was refined by Marić (1964a, 1966, 1969). His assumptions are illustrated by the following diagram.

Parent rock	Karstification under continental conditions	End product
Limestone	H_2O	Desilication: saldame, pH > 5·5
Dolostone (authigenic and allothigenic silicate minerals Al-, Fe-hydroxides + X-ray amorphous matter)	+ CO_2 terra rossa + organic matter pH = 4·5 to 8·5	Bauxitisation: bauxite, pH = 8·3

Bauxite is accordingly perceived as a product of the surface weathering of limestones and dolostones, with possible contributions from the weathering products of alumosilicate rocks. Terra rossa got bauxitized in an alkaline medium: the silica thus liberated was deposited in the form of 'psammitic quartz' ('saldame') in cavitites of the limestone, in the proximity of the bauxite region. Such accumulations of quartz are known from the vicinity of Pula, Istria, and from the islands of Susak and Viš.

The hypothetical relationship between terra rossa, bauxite and saldame is illustrated in one of Marić's geological profiles (fig. 167). According to Marić, bauxite and saldame represent the end products of the Earth-surface migration of Al, Fe and Si. He assumes that, even today, substantial quantities of silica are being dissolved on the Yugoslav karst and carried into the Adriatic. A karst spring called Pantan near Divulje, e.g., dumps according to the analyses some 100 tons per year of dissolved silica into the sea,

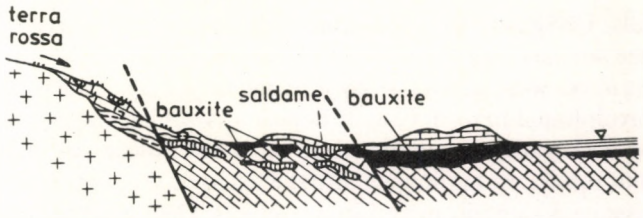


Figure 167. Genetic connection between terra rossa, bauxite, and saldame, according to Marić (1964)

at a discharge of 1400 litres per second. The 7500 cubic kilometres of water making up the Adriatic contain some 1 600 000 tons of dissolved silica and only some 140 000 tons of dissolved aluminium.

8.1.3. CHEMOGENIC THEORIES

Their basic idea is that aluminium, iron and titanium, dissolved from the rocks in the course of weathering, were entrained in solution into the lakes and seas, where they could precipitate and settle. A. D. Arkhangelsky (1937) attributed the role of principal leaching agent to organic matter, which according to him forms highly soluble complex salts with Al, Fe and Ti. Arkhangelsky denied all forms of lateritisation: according to him, there are no residual laterites anywhere in the world. His theory has few adherents today: of those, Krivtsov (1968, 1969) deserves to be mentioned.

One of the variants of the chemogenic theory is the so-called *sulphuric-acid theory*, which attributes to sulphuric acid the role of prime dissolving agent. The acid is assumed to be produced by the surface weathering of rocks containing pyrite. Ansheles (1927) was the first to propound this idea, which was then developed in detail by Goretsky (1960). Trubina (1958) interpreted the origin of the Tulsk-type deposits also in terms of pyrite decomposition.

8.1.4. HYDROTHERMAL THEORIES

The idea that bauxite could precipitate out of solutions discharged by thermal springs was first raised by Daubrée (1869) and Coquand (1871) to explain the origin of French bauxites. Hayes (1895) found that the deposits of the Southeastern US were formed from thermal springs assuming those to have dissolved the aluminium content of argillaceous rocks at considerable depth and lifted it to the surface in the form of sulphate solutions which got precipitated in geyser ponds. Szádeczky (1905, 1908) found that the Bihar Mountains bauxites were formed in hot springs associated with nearby dacitic eruptions. Lachmann (1908) explained the origin of those bauxites in terms of a unique hydrothermal–metasomatic theory, assuming that the surging thermal waters dissolved the aluminium out of the primary clay content of limestones. The aluminium sulphate solutions thus formed displaced the limestone in places, and replaced it metasomatically with bauxite. The bauxite did not, according to this theory form on the surface at all!

Kormos (1930) and Lotti (1941) associated the bauxites of the Istrian Peninsula, Yugoslavia with thermal springs. According to Yanshin (1968), the bauxites of the

Northern Urals, USSR, are indirectly due to hydrothermal activity: the aluminium content of some volcanics was dissolved by a system of thermal springs. The solutions, risen to the surface, were delivered by streams to the sea, where their aluminium content was precipitated to form bauxite deposits. A similar theory was elaborated by Zelenov (1964, 1972) who in certain areas of recent volcanism in Kamchatka and Indonesia found some streams flowing into the sea to be very strongly acidic, with high amounts of dissolved aluminium and iron contents. Once dumped into the sea, these solutions are neutralised and their Al and Fe contents are precipitated. According to Zelenov, all the karst bauxites were formed in this way.

Peive (1947) ascribed the bauxites of the Northern Urals, USSR, to submarine magmatism. The solutions rich in Al, Fe and Ti, rising from the igneous body, precipitated their metal contents once they reached the sea bottom. This is how the bauxite deposits came to exist; Peive admits no contribution at all to have occurred from any material of continental origin!

The only one to postulate a hydrothermal origin for Hungarian bauxites was Pávai-Vajna (1948).

8.1.5. VOLCANOGENIC THEORIES

These theories suggest bauxite to have formed from the lateritic *in situ* weathering of volcanic ash fallen onto carbonate rocks. According to Gedeon (1952), the ash fell into, and got bauxitised in, a shallow inland sea. He posited the formation all karst bauxites to have taken place in this manner. According to Waterman (1962) and Comer (1967), Jamaican bauxite was formed out of volcanic ash weathering into bauxite on dry land. Goldich and Bergquist (1947, 1948) found that the bauxites of Hispaniola island formed from a volcanic ash of andesitic–basaltic composition, fallen onto a limestone relief. According to Kalugin (1967), all the karst bauxites were formed in this way. Volcanic ash fell onto limestone islands rising above the sea and was weathered to a bauxite on those. The bauxite on Rennell Island in the Pacific is attributed to a similar process by Taylor and Hughes (1975).

According to Tercinier (1971), the karst bauxites on the islands of the Southwestern Pacific formed out of volcanic pumice floating in the sea, deposited on the beaches of those islands. He extended this theory to all karst bauxites by implication, although he did not expressly state it so.

8.1.6. THE PHYTOGENIC THEORY

The only proponent of this theory was Berg (1949), who stated that the Devonian bauxites of the Northern Urals, USSR, must have formed as a result of the life processes of a swamp vegetation.

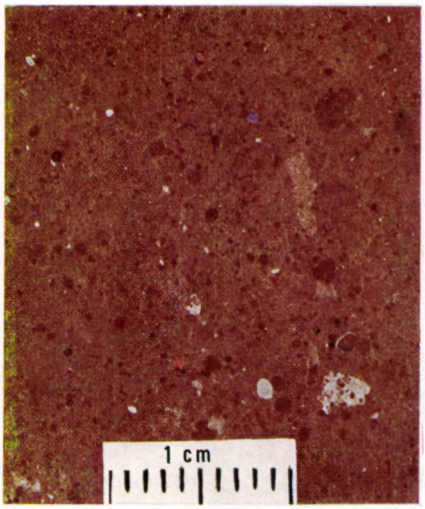
8.1.7. AN ASSESSMENT OF THE THEORIES OF BAUXITE ORIGIN

From the brief survey above, the following two general conclusions emerge.

(1) Each individual theory was originally proposed to explain the geological features of a single bauxite region, and to account for the salient characteristics of the deposits in



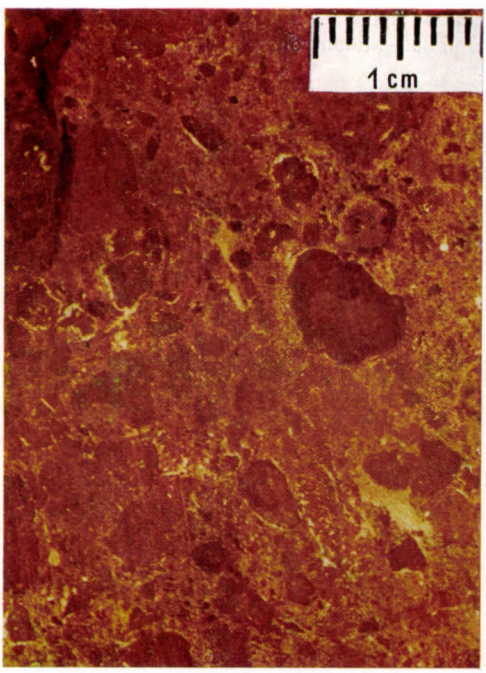
Colour figure 3. Pelitomorphic matrix. Malm bauxite, Itea, Kephali Hill, Greece. Polished rock surface



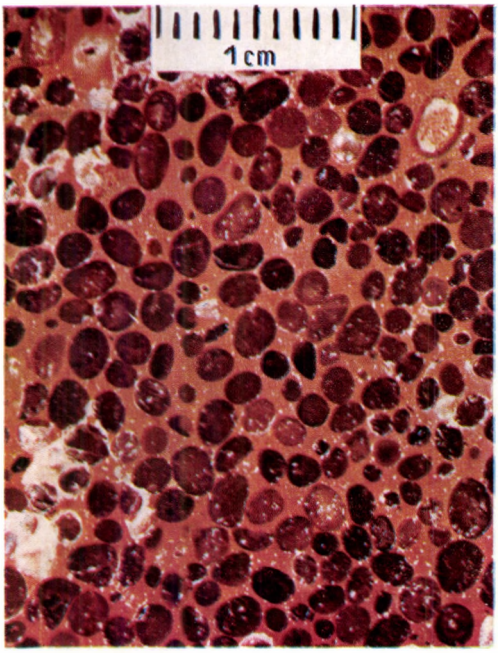
Colour figure 5. Bauxite of microlastic texture. Southern Urals, USSR. Polished surface



Colour figure 4. Macro-pisoids (pisoliths). Fuentespalda, Maestrazgo bauxite region, Northeastern Spain. Hand specimen



Colour figure 6. Bauxite of arenitic texture. Gánt, Hosszúharasztos, Hungary. Polished surface



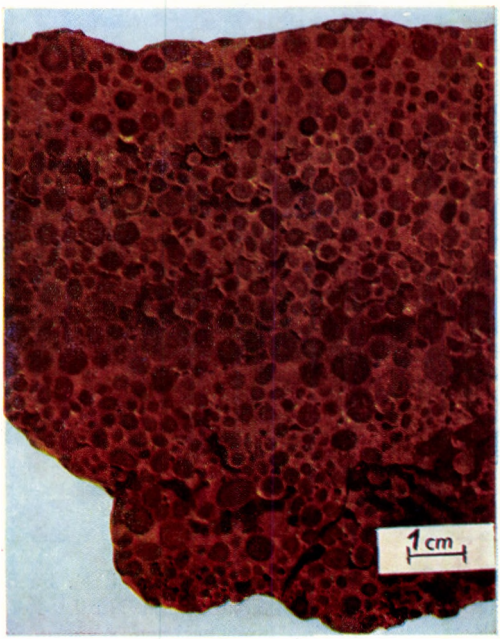
Colour figure 7. Bauxite of orbiculate texture. Halimba–Malomvölgy, Lens X, Hungary. Polished surface



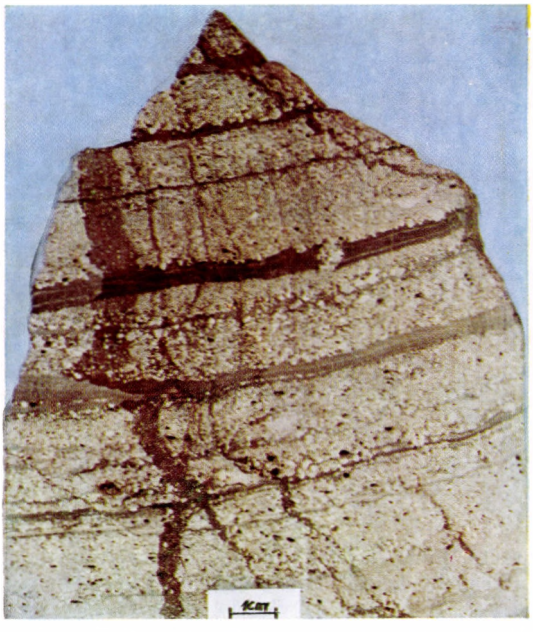
Colour figure 8. Bauxite of conglomerate texture. Nézsa open pit, Hungary. Fracture surface



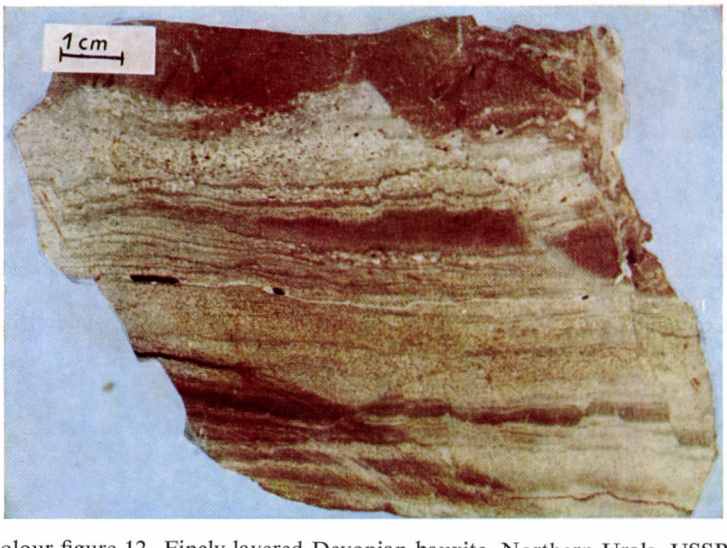
Colour figure 9. Bauxite of vague stratification. Gánt, northeastern part of Hosszúharasztos deposit, Hungary. Bauxite face



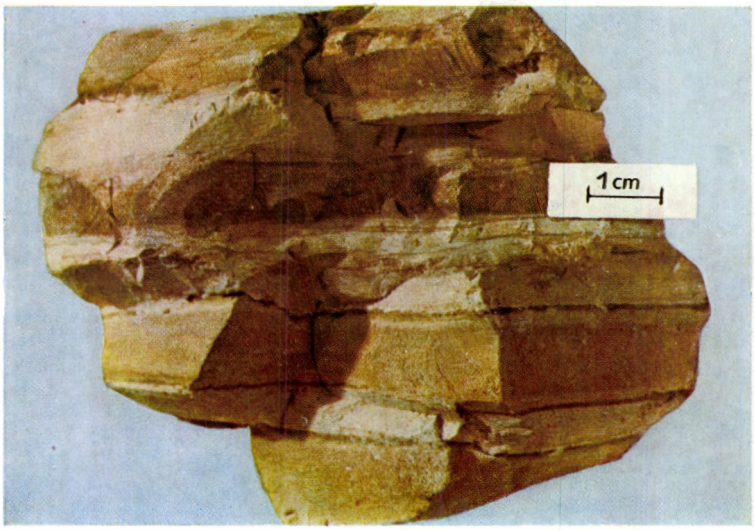
Colour figure 10. Graded bedding due to the arrangement of transported pisoids and spheroids. Verchne-Tobolsk, Turgay Plateau, USSR



Colour figure 11. Banded stratification. Devonian bauxite. Northern Urals, USSR



Colour figure 12. Finely layered Devonian bauxite. Northern Urals, USSR



Colour figure 13. Finely layered bauxite-like clay from the middle of a bauxite deposit. Hrast, Slovenia, Yugoslavia



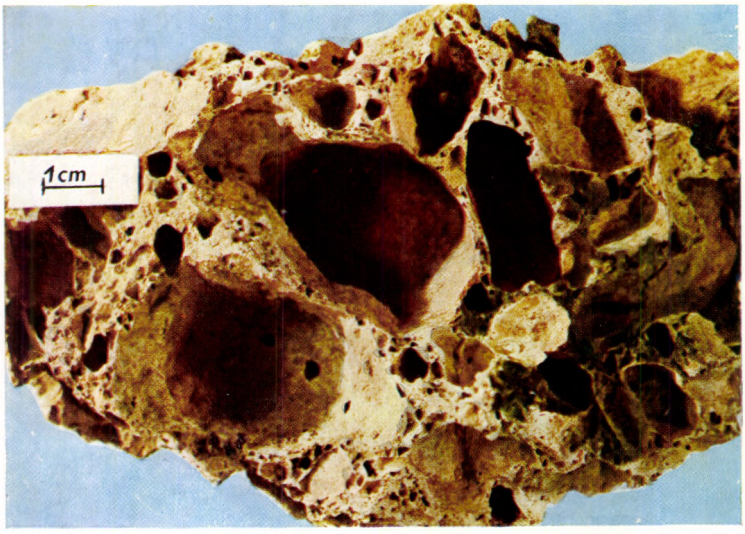
Colour figure 14. A ferruginous crust at the boundary of the underlying dolostone. Gánt–Hosszúharasztos, Hungary



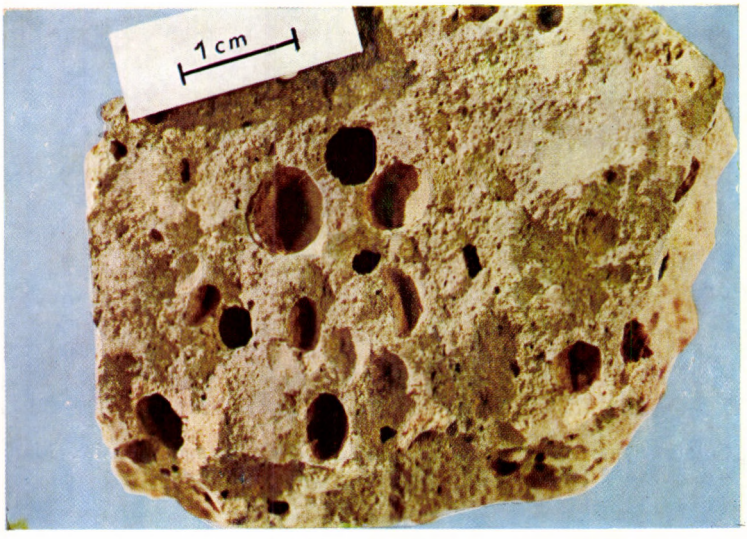
Colour figure 15. Pelitomorphic bauxite of conchoidal fracture. Zlabor Hran, Savinja Valley, Slovenia, Yugoslavia



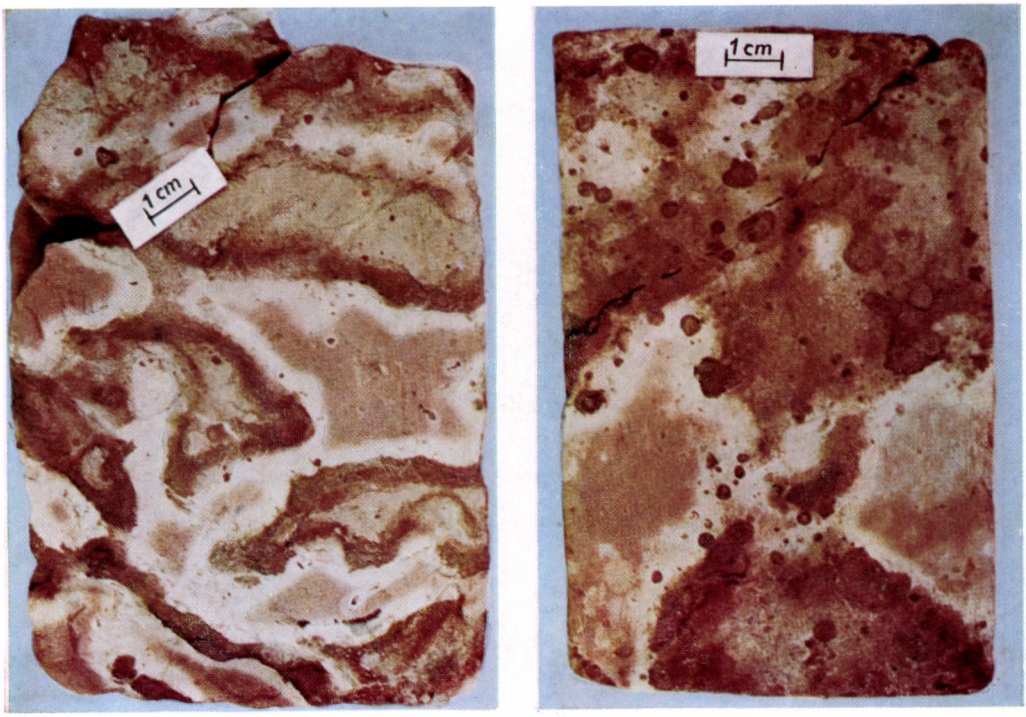
Colour figure 16. Wholly unconsolidated Quaternary bauxite. Maré Island, Loyalty Group, Western Pacific



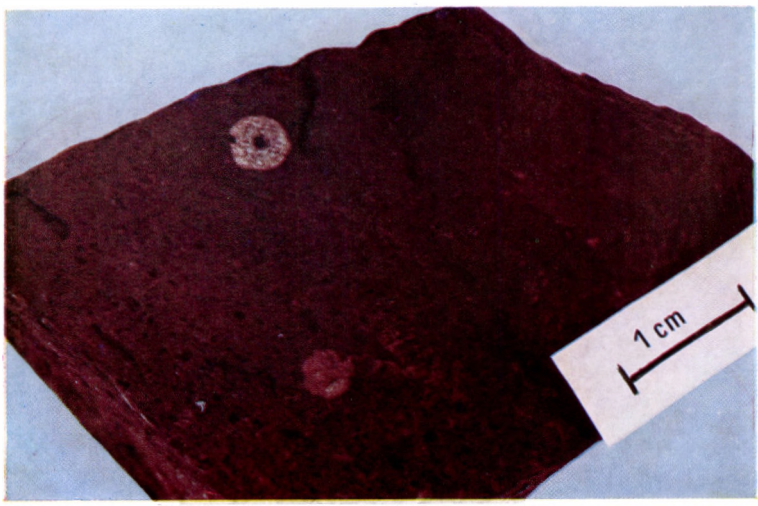
Colour figure 17. Spongy Devonian bauxite. Northern Urals, USSR



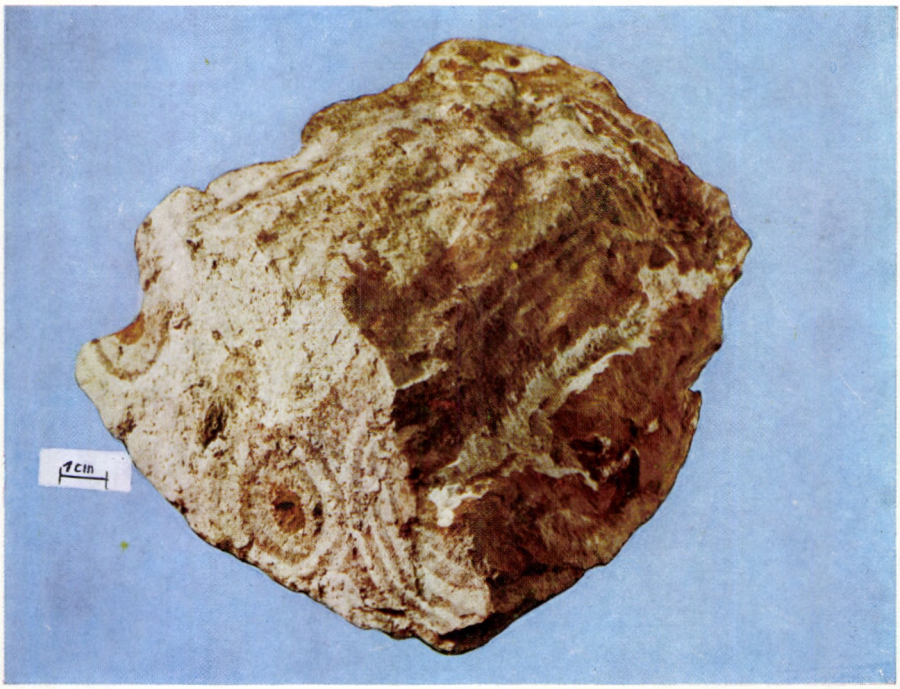
Colour figure 18. Vugular Paleocene–Lower Eocene bauxite. Colbert County, Alabama, USA



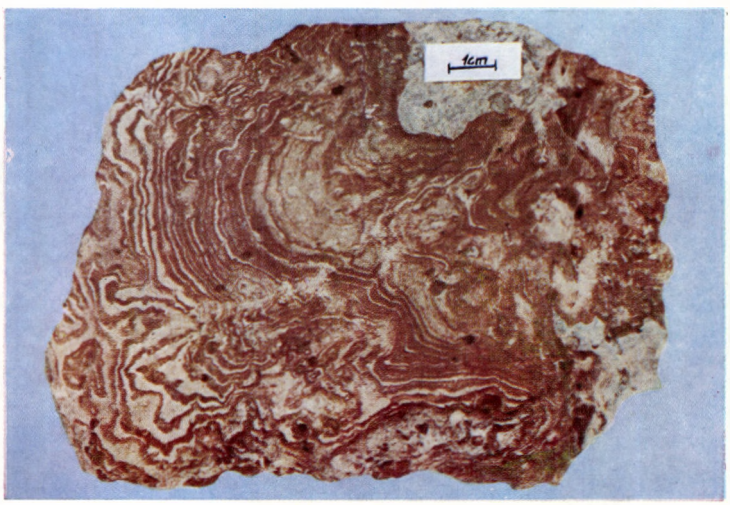
Colour figure 19. Multicoloured bauxite types. Gánt–Hosszúharasztos, Hungary



Colour figure 20. A round, pale pink spot with a dark grey nucleus in dark red bauxite. Middle Triassic. Podlipa, Slovenia, Yugoslavia



Colour figure 21. Concentric Liesegang rings about traces of roots. Nyirád–Dűlt–Nyires, Lens V, Hungary



Colour figure 22. Bauxite with Liesegang strips next to a grey pyritic bauxite. Nyirád–Izamajor, Lens I, Hungary

that region. It was only later and by degrees that their authors built them up into world-wide theories.

(2) The activity of a strong scientific personality could for years and even decades impose this or that theory in a given country (or indeed internationally). In France e.g. from the discovery of bauxite to the 1930s, most geologists were in favour of the lateritogenic theory. Subsequently, in the wake of the prominent personality and impressive works of J. de Lapparent, the terra rossa theory became sovereign, and held that position up to the late 1950s, not only in France but all over Western Europe.

In the USSR, a majority of geologists accepted the lateritogenic theory up to the 1930s. Under the influence of A. D. Arkhangelsky, however, the chemogenic theory became sole ruler between 1933 and 1955: it was only subsequently that diverse varieties of the lateritogenic and volcanogenic theories came to the fore. In Hungary, the scientific activity of Vadász established the lateritogenic theory and relegated the terra rossa theory into the background for several decades.

It would not be justified to speak of the sovereign rule of this or that theory now. Greatly intensified international scientific exchanges turned the last decade into a period of fruitful discussions laying the foundations for novel syntheses.

In the following, let us review the theories of bauxite formation in the sequence of the basic factors enumerated at the beginning of this section. Let us examine the arguments for and against the principal theories; for and against their more or less confined, local or general acceptability.

8.2. ISSUES OF PARENT ROCK

The writer uses the term parent rock to (or parent material) to denote the rock (material) from which the material of bauxite was directly derived. As far as the potential parent rocks are concerned, the theories of karst bauxite formation can be grouped as follows.

(1) Laterites formed on alumosilicate rocks (lateritogenic theory).

(2) *In situ* lateritisation of sedimentary claystones and marls (application of the lateritogenic theory to the Ariège-type deposits).

(3) Weathering residues of older limestones and dolostones (the terra rossa theory).

(4) Weathering residues of marly, clayey, tuffaceous beds intercalated in carbonate successions (a variety of the terra rossa theory).

(5) Alumosilicate rocks, with the Al, Fe and Ti dissolved out of them in the course of weathering (A. D. Arkhangelsky's chemogenic theory).

(6) Pyritised zones of alumosilicate rocks, with their Al, Fe and Ti contents dissolved by weathering (the sulphuric-acid theory).

(7) Alumosilicate rocks, their Al, Fe and Ti contents dissolved by hydrothermal processes (hydrothermal theories).

(8) Volcanic ash fallen onto a carbonate-rock relief (volcanogenic theories).

(9) Pumice floating in the sea (a variety of the volcanogenic theory).

(10) A diversity of parent rocks (multifactor theories).

8.2.1. A LATERITIC PARENT ROCK

According to the most frequently invoked argument against this theory there are no laterites or their remnants of erosion in the proximity of most karst bauxite deposits (Denizot 1961, 1962; Bonte 1970). In the context of the bauxites in the Transdanubian Hills, Hungary, this was regarded as the main counterargument by Vendel *et al* (1971) who felt that, at the time of origin of the bauxites, the crystalline rocks amenable to lateritisation were not even exposed. They were probably covered by Permian and Triassic strata, which were only later eroded.

The absence of pre-existing laterites is insufficient for the refutation of a theory of lateritogenic origin, as the laterites invoked may in fact have existed, to be only fully eroded afterwards. Another point to be kept in mind is that the laterite in question must of necessity have lain higher than the karst bauxites derived from them, which renders their complete erosion even more likely. More painstaking research in recent years has discovered laterites or remnants of such in areas previously regarded as unlikely, e.g. in the USSR, in the Northern Urals (Bushinsky 1971) and in Soviet Central Asia (Abdullayev 1970; Konnov 1972).

According to de Weisse (1964), the outcrop area of volcanic rocks in Jamaica is too small to have yielded the parent material of the more than one thousand million tons of bauxite accumulated on the limestone plateaux of that island. This quite correct argument does not, however, imply that no lateritic weathering products could at all contribute to the accumulation of these bauxites as is stipulated by de Weisse.

The credibility of Zans's theory in its turn was reduced by Robinson's (1971) palaeogeographic studies. Robinson states that the entire central and western part of Jamaica was covered by an uninterrupted limestone mantle. After its emersion, this had to be eroded first: it was only eight million years ago that the alumosilicate rocks of the Central Inliers became exposed. Their detritus in fact first appears in the surrounding latest Miocene and Pliocene sediments. This greatly reduces the time available for the lateritisation of the central Inliers. A further difficulty is that most of the exposures of these rocks lie deeper than the bauxite-covered limestone plateaux: that is, their weathering products could not be delivered onto those by stream transport.

In the present writer's opinion, it would be wrong either to reject this mode of origin out of hand or to declare it to be of general validity. Only by a thorough palaeogeographic study of each separate bauxite deposit group can it be found out whether alumosilicate rocks suitable as a parent material were sufficiently exposed to be available for lateritization.

As another argument against a lateritic derivation of the parent material, the alleged absence of clastic lithologic features and of clastic mineral grains in karst bauxites has been invoked (de Weisse 1964). More recent research has shown, however, clastic texture types to be very widespread. Furthermore, allothigenic mineral grains deriving from igneous and metamorphic rocks have also been found in many bauxite deposits, albeit in small quantities. Demangeon (1965, 1970) e.g. proved by micromineralogical means in the Languedoc and in the Alpilles regions, France, that the sizes and abundances of allothigenic heavy minerals decrease in step with increasing distance from the Massif Central. The heavy mineral composition is in good agreement with the associations of the Massif Central crystalline.

In the Transdanubian Hills bauxites allothigenic minerals formed in igneous and metamorphic rocks were found by Kiss (1955) and Vörös (1958). The monazite and xenotime grains identified in Gánt bauxite by means of the electronmicroprobe (Bárdossy and Pantó 1973) very probably came from the Velence Hills granite.

The writer regards as a direct proof of a genetic association of karst bauxites with laterites those bauxite pebbles which can be identified by their textures and mineral composition with the ferricrete crusts ('cuirasses') of pre-existing laterites. The pebbles in question were described in the chapter on petrography. Nicolas (1968, 1970) went so far as to perceive the bauxite deposits of the Provence as inverted laterite profiles, with the pebbles from the ferricrete at the bottom, with the pisoidic–concretionary zone in the middle, and with the more clayey 'lithomarge' on top. Moreover, he surmised that the sandy strata in the immediate cover of the Haut-Var deposit to represented the less weathered basal parts of these ancient laterites. The present writer regards this as an exaggeration of an otherwise correct basic idea, as it is difficult to imagine how erosion could take off the laterite slice by horizontal slice, and deposit it in exact reverse order.

The last main counterargument is that, on a karsted limestone and/or dolostone relief, no alien material can be brought in by stream transport: that is no parent material could be brought in from the outside (de Weisse 1964). This counterargument does not hold water, either. In recent years post-bauxitic karstification has been recognised in an increasing number of bauxite regions. This means that, at the time of bauxite formation, the relief was much less dissected than it is today, and the introduction of foreign material by stream transport was quite possible. The sandy, and in places even gravelly beds found between and under the bauxite lenses of the Kazakhstan-type deposits are a direct proof of this statement. Such clastics exist, albeit not so widespread, also in the Mediterranean-type deposits (cf fig. 33).

In the immediate vicinity of the huge Weipa bauxite deposit in Australia, some reworked laterite bauxite was deposited on the limestone relief of the seashore. Karstification is currently going on under the bauxite; indeed, the bauxite itself goes on being desilificied (Grubb 1973). This, then, is direct evidence for a lateritic parent rock.

In summary, a lateritic parent rock is to be rejected out of hand only where there are no igneous or metamorphic alumosilicate rocks amenable to lateritisation in the vicinity of the bauxite region. It is not necessary to stipulate that lateritisation should have allowed the development of a fully developed bauxite; it is sufficient for it to have produced a clayey type of weathering product, which could get bauxitised during one or several redepositions and also in its definitive lie. Such a process was posited e.g. by Demangeon (1970) and Roch (1973) for the French bauxites.

8.2.2. CLAYEY, MARLY PARENT ROCKS

There is direct evidence for the *in situ* lateritisation of such rocks in the Ariège-type deposits, where clays and marls grade into bauxite upward (cf the section on deposits). This comparatively rare combination of parent rock and mode of origin, confined to a single deposit type, should not be extended and generalised to all the other karst bauxites.

8.2.3. A CARBONATE PARENT ROCK

The main argument against deriving bauxite from the weathering residues of pre-existing dolostones and limestones is that these rocks tend to be very pure, containing only some tenths or hundredths of a percent of aluminium, iron and titanium. The accumulation of the more sizeable bauxite deposits, then, would have required the decomposition of several thousand metres of carbonate rock (cf e.g. Vadász 1951; Goldich and Bergquist 1947; Tercinier 1971).

A fact worthy of mention in this context is that Brugger (1940) in Hungary was one the first workers world-wide to perform calculations based on thorough chemical analyses to find out about the quantity of bauxite potentially produced by the chemical weathering of dolostone. He found that one cubic metre of dolostone from the Buda Hills gave a mere 1·9 kg of indissoluble residue on average, which according to him is much too little for deriving the Hungarian bauxites from dolostones. Although his calculations contain several errors (e.g. failure to account for the Al dissolved in the hydrochloric acid used to dissolve the rock), his early statement of the problem was nevertheless a noteworthy achievement.

This quite real difficulty was removed by the adherents of the terra rossa theory by letting drop the principle of strict *in situ* origin and by admitting that the weathering residues could be washed together from sizeable areas into the local depressions of the relief (de Weisse 1964). In this way, the deposits could be derived from much more extensive areas, which greatly reduced the depth of weathering required. In recent times, calculations based on this idea were carried out for many bauxite regions with a view to establishing the depth of weathering of the carbonate formations required to produce a given tonnage of bauxite. Lack of space forbids the presentation of all these calculations here: for an example, let us consider those referring to the bauxite region of Jamaica.

According to Hill (1955), bauxites cover 15% of the limestone surface: their avarage thickness is 7·6 m. The weathering residue of the limestone complex is 0·5% on average. On this basis, the weathering of 230 m of limestone would be sufficient to account for the deposits. In Hill's opinion, however, about half of the weathering residue was removed by erosion: he accordingly posits the weathering of 460 m of limestone.

Hartman (1955) states that Jamaican limestones contain 0·075% Al_2O_3 on average. All his other starting data are the same as Hill's. He calculated on this basis that the removal by weathering of 244 m of limestone, without admitting the highly probable partial erosion and removal of the weathering residues.

According to Hose (1961), the limestones in question contain 0·31% Al_2O_3 on average. His calculations, not presented in detail, gave him a thickness of 122 m for the limestone whose weathering down was required to yield today's bauxites. He probably did not reckon with any erosion loss, either.

Burns (1961) dissolved his samples of bedrock limestone in hot 30% hydrochloric acid. Working back from his analyses of the dissolution residue, he obtained a content of 0·009% Al_2O_3 for the original limestone. Admitting a loss on erosion, he obtained a thickness of 915 m limestone to be weathered.

Kelly (1961), in calculating the average Al_2O_3 content of Jamaican limestone made use of chemical analyses of both surface samples and borehole cores, finding an average

content of 0·036% Al_2O_3 which, with a certain loss on erosion taken into account, gave him a thickness of weathering of 409 m.

According to undetailed calculations by Sinclair (1966), the weathering away of 240 m of limestone must be reckoned with.

Comer (1974) assumed most of the Miocene limestones to be contaminated in their near-surface parts by fine particles of bauxite and clay. He therefore took into account the composition only of the whitest, purest-looking limestone lumps. The average Al_2O_3 content of these was 0·006% for a SiO_2 content of 0·0275%, giving an Al_2O_3/SiO_2 module of 0·22 only. Starting from these figures, Comer found that the thickness of limestone to be removed by weathering was 1500 m, provided no aluminium at all was lost in the process. Reckoning with a certain loss on erosion, he arrived at 9000 m as the most likely figure.

The figures arrived at by these calculations scatter rather widely, between a minimum of 122 m and a maximum of 9000 m, although the region envisaged is an island whose areal extent is quite well-defined. Bauxite tonnage is fairly accurately known: the principal point of divergence between the results is in the average content of weathering residue, and in the assessment of the loss on erosion. A methodological error was also committed: it was first pointed out by Komlóssy (1967). Notably on obtaining a dissolution residue from a limestone or dolostone by treating it with hydrochloric acid, part of the Al_2O_3 content is also dissolved. This is why the Al_2O_3 contents found by Burns are so low, and the depth of erosion of limestone stipulated by him so great.

Another source of error is that a limestone or dolostone complex looking uniform to the eye can in fact vary rather widely in chemical composition with depth. According to Clarke, e.g. (1966), in Manchester Parish, Jamaica, the composition of the bedrock limestone varies with depth as follows.

Depth (ft)	Al_2O_3	Al_2O_3/SiO_2
0–100	0·25 %	2·08%
100–190	0·052%	1·57%
200–310	0·021%	1·05%
450–500	0·011%	0·44%

The decrease in average aluminium content is seen to be accompanied by a substantial decrease in the Al_2O_3/SiO_2 module. This emphasises the impossibility of obtaining the desired accuracy using samples taken from the surface only. Finally, whenever the bauxite region in question is not an island, the surface area to be taken is itself ill-defined.

Robinson (1971) deduced from the findings of his stratigraphic studies a removal of 70 to 270 m of limestone on the north side and of 70 to 100 m on the south side of Jamaica. It is only above the Central Inliers that dissolution and erosion could remove 500 to 700 m of limestone.

These thicknesses are insufficient to make the carbonate rocks the only source of material, especially as regards the south side of the island.

A given bauxite deposit may be derived exclusively from pre-existing carbonate rocks only if its clastic minerals can also be detected in those rocks: these particles may not be finer than those found in the bauxite. Demangeon (1970) proved by

micromineralogical analysis that the allothigenic grains of zircon, rutile and tourmaline found in the bauxites of the Alpilles and the Languedoc, France, are larger by one order of magnitude than the grains of these same minerals found in the bedrock limestones. Moreover, the bauxite contains clastic minerals which could not be identified in the underlying carbonate rocks (e.g. disthene, staurolite, corundum). In this case, then, the underlying carbonate rock could not have been the only source of bauxite.

Elsewhere, the fit may be quite good. According to Hartman (1955), e.g. Jamaican bauxite contains, the same allothigenic heavy minerals as the limestone underlying it. It thus becomes obvious that this counterargument must also be proved or disproved separately for each bauxite region.

According to several authors, the weathering of limestone and dolostone is a very slow process, incompatible with the short hiatuses in which many deposit groups are found. De Weisse (1964) pointed out in this connection that the underlying carbonate rocks on Jamaica and Hispaniola are very loose and porous: they are prone to extremely rapid weathering and dissolution on the surface. He deduced that the carbonate bedrock strata must have been just as loose and prone to weathering at the time of formation of the Mesozoic deposits in the Mediterranean region; those were turned into today's hard, massive, hard-to-dissolve rocks only by the millions of years which have elapsed since. De Weisse's well-founded argument thus dispenses with this particular counterargument.

In the proximitiy of some bauxite regions there are no outcrops of alumosilicate rocks amenable to lateritisation, or such outcrops as exist are too small to have produced enough weathering products to account for the bauxites of the region. Jamaica itself is a case in point for the latter contingency. Also, the entire surface, 55 km long and 15 km wide, of the Ile de Gonave, belonging to Haiti, is covered with limestones ranging from the Eocene to the Pleistocene. According to Butterlin (1958), these limestones are covered by a mantle of red bauxitic clay, which may attain a thickness of several metres in local depressions of the karst relief. The bauxitic clay contains no minerals indicative of volcanic ash falls: it is therefore regarded by Butterlin as the product exclusively of limestone weathering. In such situations—provided no ash fall took place, or no traces of such can be demonstrated—the pre-existing carbonate rocks may in fact be the sole or principal parent rocks of the bauxite.

8.2.4. CLAYS, MARLS OR TUFFS INTERCALATED IN CARBONATE SUCCESSIONS

If such beds are to be regarded as producing the bulk of parent material, with the weathering residues of pure limestones and dolostones merely contributing the balance, the required depth of weathering can be substantially reduced.

Miholić (1956), for one, derived the bauxites of Yugoslavia from the weathering of marls and clays intercalated into the succession of Mesozoic limestones and dolostones. According to Luković (1962), the bauxite of the Gornje Polje deposit group in Crna Gora (Montenegro), Yugoslavia, was formed, in addition to pre-existing limestones and dolostones, primarily out of tuffaceous layers intercalated in those. According to Jurković and Sakać (1964), there were, during the Carnian episode of

bauxite formation in Yugoslavia, outcrops of marls and clays in the proximity of the bauxite deposits, in topographically higher positions, in addition to the outcrops of limestones and dolostones. Under a climate favourable to weathering, the marls and clay also were weathered, and their weathering products, washed together into the depressions of the relief, served as the parent material of the bauxite. The same process was also repeated during the more recent episodes of bauxite formation. Differences in composition between the bauxite horizons may, according to those authors, be due to the fact that they represent mixtures in different proportions of the weathering products of carbonate rocks on the one hand and clayey–marly ones on the other.

According to Denizot (1961, 1962), the parent rocks of bauxite in the Provence, France, were largely lower Aptian marls. These marls are found today immediately to the south of the bauxite region; according to Denizot, however, they once covered the entire surface from the Alpilles to the Brignoles area. In the Upper Aptian, the area was laid dry, forming what is called the Durancian Isthmus, and the marl weathered into a bauxite in its position then. The bauxites of this region, then, are of a stricly *in situ*, eluvial origin. The Lower Aptian sedimentation probably did not extend to the Haut-Var and the Languedoc. Denizot accordingly regards some marls in the Jurassic succession as the parent rocks of the bauxites there. He calculated some chemical analyses of the Lower Aptian and Jurassic marls in such a way as to remove first the carbonate components and then most of the silica. He regarded it as a proof of his hypothesis that the residue thus obtained strikingly resembled the composition of Brignoles bauxite.

Denizot's theory is contradicted by the fact that, so far, it has not been possible to detect any traces of Lower Aptian marls in the bauxite profiles of the regions. It is rather unlikely for the total bulk of those to get bauxitized without leaving any recognizable residue anywhere. Roch (1956) states the Lower Aptian marls to contain numerous chert nodes. It is difficult to imagine all of these dissolving everywhere without a trace. If the bauxite is in fact eluvial, it is hard to conceive why it should be confined to comparatively narrow strips today provided the marl—if it did in fact extend that far—covered the area in an uninterrupted mantle. Rousset (1968) held views similar to those of Denizot, stating as he did that the parent materials of bauxite were Aptian and Albian marls in the southwestern and Neocomian marls in the northeastern part of the Provence.

Szabó and Ravasz (1970) regarded the trachyte tuffs and tuffites found in the Upper Anisian and Ladinian carbonate succession of the Transdanubian Hills, Hungary as the parent material of the bauxites there. In support of their hypothesis, they performed micromineralogical analyses, finding in the bauxites among others grains of idiomorphic quartz of volcanic origin, weathered feldspar and volcanic glass.

The parent rocks thus specified being in the nature of intercalations, this hypothesis will operate only in combination with the hypothesis of the weathering of the entire carbonate succession. There are many bauxite regions where the pre-existing carbonate succession contains no such intercalations. Hence, this mode of origin also may be postulated for some bauxite regions only, and not for the karst bauxites in general.

8.2.5. A. D. ARKHANGELSKY'S CHEMOGENIC THEORY

The principal argument against this hypothesis is that lateritization is characterised precisely by Al, Fe and Ti being left in place to concentrate rather than by the dissolution and removal of these elements. That is, the very foundations of the theory are wrong. We shall see it in the next chapter that the removal of bauxitic components in true solution is not likely, either. The present writer accordingly is of the opinion that no bauxite deposits could form in the manner stipulated by A. D. Arkhangelsky.

8.2.6. THE SULPHURIC ACID THEORY

In most bauxite regions, the fundamental postulate of the theory remains unfulfilled; there are no pyritised alumosilicate rocks near by. Even where there are such rocks and alteration could in fact produce sulphuric acid, it is inconceivable for the sulphuric acid solutions to have reached where this hypothesis requires them, because they would have been neutralised immediately on contact with a carbonate rock. All the bauxite deposits should accordingly be confined to the outskirts of a carbonate region. In fact, however, the bauxites may lie all over a carbonate region, even at distances of 20 to 50 km from its periphery. Hence, this hypothesis is likewise unsuited as an explanation of the origin of bauxite and its parent rock. An origin involving sulphuric acid may only be envisaged for the Tulsk-type deposits (cf Section 8.5).

8.2.7. THE HYDROTHERMAL HYPOTHESIS

The deposit features, the chemical and mineral composition of bauxite and the fossils found in it are so profoundly incompatible with all sorts of hydrothermal theory that no detailed refutation seems to be required. In the writer's opinion, no bauxite deposits came to exist in this way.

The only variant worth mentioning of the hypothesis is Zelenov's, outlined in Section 8.1.4; its author maintaining to this day that all the karst bauxites in the world were formed in this manner. Such a generalisation of his hypothesis is contradicted by the fact that, in numerous bauxite regions, there are not even traces of any contemporary volcanism. The acidic waters of Kamchatkan creeks of hydrothermal origin, adduced as an example by Zelenov, are high in aluminium and iron only; their dissolved titanium content is insignificant. Hence, the bauxites hypothetically forming out of those waters should also contain very little titanium. However, all karst bauxites except those of the Tulsk type contain quantities of TiO_2, in the 2 to 5% range.

Finally, as pointed out by Strakhov and Bushinsky (1971), around Kamchatka and the Kuriles, where acidic creeks even today, dump vast quantities of dissolved aluminium and iron into sea, no bauxite-like sediment even faintly resembling bauxite can be observed either close inshore or farther offshore, because the surf and the marine currents spread out the dissolved Al and Fe which soon unites with the silica content of sea water to form clay minerals. Hence, Zelenov's hypothesis is also unacceptable.

8.2.8. VOLCANIC ASH AS A PARENT MATERIAL

In view of their high Al, Fe and Ti contents, volcanic ash is ideally suited as a parent rock of bauxite. It is. however, rightly argued against the generalization of this theory that in many a bauxite regions there was no contemporary volcanism, so that no ash falls could take place. Another counterargument is that in no bauxite deposit could any weathering residues of volcanic ash be identified so far, although it is unlikely for those to have disappeared without any trace. To counter this attack, the adherents of the theory refer to the extraordinarily easy weathering of volcanic ash, which leaves only micromineralogical observations as apt to furnish the only evidence for testing the hypothesis.

Goldich and Bergquist (1947, 1948) found grains of magnetite, ilmenite, zircon and quartz, presumed by them of volcanic origin, in the bauxites of Hispaniola island. In the region, there was uninterrupted volcanism from the Cretaceous to this day, so that the possibility of ash falls did in fact exist.

Waterman (1962) stated the parent material of Jamaican bauxite to be most probably wind-borne volcanic ash which, falling onto a karst surface, was bauxitised on dry land. The idea was further elaborated by Comer (1974) who regards his finding of bentonite interbeddings 5 to 15 cm thick and finely scattered bentonite particles of volcanic origin in the Middle and Upper Miocene Montpelier limestone on the north side of the island as supporting evidence. According to Comer, the ash fell also onto those parts of the island which had emerged by that time. Ash fallen there could undergo rapid bauxitization on the limestone relief. He calculated four metres of ash to have been sufficient to produce the entire tonnage of bauxite known today. Ash fallen into the shallow sea surrounding the island was turned into bentonite while the part fallen onto the emerged limestone relief has being bauxitised.

In the present writer's opinion, a derivation from volcanic ash is supported by the peculiar convex reliefs of certain Jamaican bauxite deposits (cf figs 13 and 14), which cannot be explained in terms either of limestone weathering or of erosion. He does not, however, agree with Comer's view that the weathering residues of the limestones must be totally excluded from among the parent materials of these bauxites.

Taylor and Hughes (1975) attribute the bauxite on Rennell Island also to scatterings of volcanic ash. Two hundred kilometres to the north of Rennell, on Guadalcanal Island, there are volcanoes that have been active with interruptions, from the Pliocene to the present. According to those authors, the winds could easily transport the volcanic ash to Rennell Island. They regard the ilmenite grains found in the bauxite there as a confirmation of their hypothesis.

In 1971, in one of the deposits of the Campo Felice group, Abruzzi Mts, Italy, a vivid green clay resembling a volcanic ash was found at the boundary of bauxite and bedrock by Bárdossy *et al* (1977). The clay, 2 to 30 cm thick, grades upward with no sharp limit into a clayey bauxite. Apart from some local, secondary clacite impregnations, its mineral composition comprises 76% glauconite, 18% kaolinite, 6% montmorillonite. No montmorillonite had been found before in rock-forming abundance in any bauxite deposit; the mineral was known only as a very sporadic component in a few deposits. No glauconite had been identified either in any bauxite deposit. The two minerals cannot be weathering products of the underlying limestone, which neither of them

contains in detectable quantities; anyhow, both would have changed into kaolinite in the course of alteration. They could not form in the course of bauxitisation, either.

D'Argenio's (1970) palaeogeographic studies proved the region to have been an archipelago at the time of bauxite formation. The deposits had formed on small islands made up exclusively of limestone. Even in a palaeogeographic context, then, the only option available is to regard this green clay as the strongly weathered derivative of a volcanic ash, as confirmed by the fact that montmorillonite and glauconite are frequently found as the weathering products of tuffs. Investigating the green clay by means of the electron microprobe the writer has found a very few sporadic columnar grains of 40×5 μm in size resembling an amphibole, as another point in favour of a tuffaceous origin. In a bauxite deposit near Cusano Mutri of this same deposit group, the writer has found a pseudomorph of boehmite after a pyroxene (fig. 73), which implies the same sort of tuffaceous parent rock. It may thus be stated that an ash fall is highly likely to have contributed to the parent material of the Central Italian bauxites.

In summary, then, in all the regions where volcanism took place contemporaneous with bauxite formation, contributions of volcanic ash to the parent material of bauxite are possible in principle. Traces of ash falls will presumably be found by more painstaking search also in other deposits.

8.2.9. FLOATING PUMICE AS A PARENT MATERIAL

Pumice, stipulated as a parent rock by Tercinier (1971), can in fact be found in the Loyalty group of islands in the Pacific, but only in the tidal zone or slightly above it. These lumps of pumice of dacitic composition were in fact delivered into their present lies by floating on the waves. Bauxite, on the other hand, covers the entire surface of each island, in places at altitudes higher than 100 m. No lumps of pumice have so far been found in the bauxite. The soil developed on the pumice also differs greatly from a bauxite: it is almost black and very rich in organic matter (up to 30%), and contains some allophane-like weathering products. Only traces of boehmite could be detected in it. In the light of all this, Tercinier's explanation seems rather unlikely.

According to Taylor and Hughes (1975), pumices in the Pacific contain 60 to 75% SiO_2 on average. They too think it unlikely for the bauxites of the Loyalty group to have formed out of a material as rich in silica as this.

An extension of this hypothesis to the other karst bauxite regions of the world is even less acceptable. In many places, there was no contemporary volcanism, and there were accordingly no pumice falls either. In any case, it is hard to fathom how several hundred million tons of bauxite could form in comparatively small areas, uniquely out of lumps of pumice floating in the sea.

8.2.10. COMBINATIONS OF SEVERAL PARENT MATERIALS

This idea is most likely where the hiatus is long and a variety of rocks are exposed on the surface in or around the bauxite region. In such cases, provided the climate was a suitable one, lateritic weathering affected all exposed rocks, the weathering products of which contributed in different proportions to bauxite accumulation, depending on the relevant features of the relief.

These conditions fit the Transdanubian Hills region in Hungary extremely well, and the writer holds this to be the most likely derivation of parent material for that region. Barnabás (1966) stated the parent material of the bauxite to have been made up by weathering products of all the rocks exposed on the surface in the neighbourhood. According to Balkay (1966), the limestones of the Upper Triassic, Jurassic and Lower Cretaceous all contributed to the parent material of bauxite, and so did the Paleozoic sedimentary and crystalline rocks of the southern foreland.

Micromineralogy also implies a variety of parent rocks. It is indeed hard to imagine that, during such a long period of weathering, on a relief so varied as to lithology, the bauxite should have been made up of the weathering products of one rock type only. Such an assumption would in any case leave open the question as to what happened to the weathering products of all the other rock types. The proportions of parent materials were different from one deposit group to the next, depending on geological structure, palaeorelief and transport conditions. The clarification of the details will require careful micromineralogical, electron microprobe and trace elements studies. The quantitative micromineralogical studies performed by Vörös and Gecse (1976) did in fact permit an estimation of the shares of the principal parent rocks for the Nagyegyháza and Iszkaszentgyörgy deposit groups.

According to Bonte (1970), French bauxites were formed, in addition to the weathering residues of carbonate rocks, also from the lateritic weathering products of the aluminium silicate rocks in the surroundings. Nicolas (1970) stated that the parent materials of the bauxites in the Provence could be all the rocks that, exposed on the elevated parts of the relief—on the anticlines—at the time of bauxite formation, underwent lateritisation there. This includes Triassic, Jurassic and Lower Cretaceous rocks: of those, Nicolas assigned major roles to Aptian–Albian marls, to Permo–Triassic and Rhaetic clayey, clastic rocks, and to Triassic basaltic volcanics; the latter were most important in the Haut-Var area. Combes (1973) stated that the parent rocks of the Bédarieux deposit group were the Precambrian, Ordovician and Visean shales, Permian pelitic rocks and Triassic and Liassic clay marls and marls exposed in the Monts Faugères immediately to the south of the deposits.

According to Marić (1966), Yugoslav bauxites do not have carbonate rocks for their sole parent materials, either; the weathering products of volcanics also contributed to their accumulation. Clarke (1966) stated that the Jamaican bauxites formed out of the weathering residues of the pre-existing limestones, plus the lateritic alluvia proposed by Zans, plus the material of volcanic ash falls. He holds the weathering products of the limestones to have made the greatest contribution.

The theory of parent-rock combinations should not be generalised either, however. It should not be applied where the area involved is uniform geologically and the hiatus was a short one.

In summary, none of the theories concerning the parent materials of karst bauxite is of general validity. Parent materials may differ from one deposit group to the next, depending on the geological evolution of the bauxite region under scrutiny; in fact, in several bauxite regions, contributions by a combination of parent rocks seem to be the most likely.

8.3. ISSUES OF TRANSPORT

Different authors suggest karst bauxite to be autochthonous, allochthonous and parautochthonous, respectively. A bauxite deposit is autochthonous if it was bauxitised in place; it is allochthonous if the material was deposited in its present lie as a fully developed bauxite; and parautochthonous if the fully developed bauxite underwent a rather limited, local redeposition only.

8.3.1. THEORIES OF AUTOCHTHONY

The most typical representative of these hypotheses is the classical terra rossa theory, which derives bauxite from the *in situ* weathering of a carbonate bedrock. This view is called '*autochthonie absolue*' in French literature. There are many arguments against it, including textural and lithostructural features implying transport of the bauxite, the existence of pebbles and lumps of alien rock between the bauxite and its bedrock, etc. This is why this strict version of autochthony is retained today by very few authors (e.g. Denizot 1961, 1962).

Volcanogenic theories also stipulate the *in situ* bauxitisation of volcanic ash fallen onto a carbonate relief.

Bauxitic and bauxite-like clay found at the base of a deposit are held to be autochthonous by de Weisse (1964) and Bonte (1970). In the course of karstification under a bauxite, the decomposition products of the underlying limestone or dolostone are trapped in any case on the bedrock top, and must gradually thicken there. The thickness of this formation, however, is very limited of necessity, as the observations so far available imply decomposition under the bauxite to have removed 10 to 15 m of bedrock at most, and the resulting decomposition residue could not be too copious in Mediterranean-type deposits.

Ariège-type bauxites are autochthonous beyond doubt, since the parent rocks underlying the deposits pass gradually into the bauxite upward; also in the profiles, the imprints of *in situ* bauxitisation can be observed.

8.3.2. THEORIES OF PARAUTOCHTHONY

These theories posit the washing together of weathering products into local depressions of a karst relief, over distances ranging from a few hundred metres to a kilometre at most, the routes of transport being confined strictly to the karsted area. In other words, no contributions are admitted from the neighbouring areas where alumosilicate rocks were exposed. It is the modern adherents of the terra rossa theory who stipulate this, with a view to resolving the contradictions of absolute autochthony (de Weisse 1964; Périnet 1964; Bonte 1970). Vendel *et al* (1971) also conjectured transport routes not longer than a few hundred metres during the formation of the Transdanubian Hills' bauxites of Hungary.

In the writer's opinion, parautochthony should not be generalised to all karst bauxites. It is acceptable without any qualifications wherever a bauxite formed on a smaller or larger island entirely made up of limestones, e.g. in the Western Pacific. Paleogeographic investigations imply such islands also to have existed at the time of

bauxite formation on Hispaniola, in Central Italy, and in certain bauxite regions of Dalmatia. In Jamaica, too, most of the bauxite is presumably parautochthonous, but some admixtures of allochthonous origin cannot be excluded, in view of the proximity and higher elevation of some exposures of alumosilicate rocks. In all the other bauxite regions, the joint accumulation of parautochthonous and allochthonous materials seems most likely.

8.3.3. THEORIES OF ALLOCHTHONY

These theories posit transports of material over longer distances, from outside the karsted region. Length of transport is shortest in the case of the Kazakhstan-type deposits (0·5 to 5 km). The actual laterite-type weathering profiles providing the parent materials of these deposits could be identified more often than not.

In the case of the Mediterranean-type deposits, it is usual to stipulate a length of transport between 2 and 20 km, depending on how far the exposures of the conjectured parent rock are from the deposits. According to Bushinsky (1966), transport routes of 30 and even 40 km were possible in some exceptional cases. No routes longer than this are admitted as a rule because, the longer the transport, the more opportunity there was for alien contaminants (sand, gravel), to get admixed to the bauxite. The lack of quartz sand and gravel in the bauxites therefore tends to exclude such lengths of transport. The only author to envisage transport routes longer than 100 km was Arkhangelsky in his chemogenic theory which, however, is rejected today by an overwhelming majority of specialists.

The argument against allochthony raised by de Weisse (1964, cf Section 8.2.1) is not valid.

In reality, autochthony and allochthony cannot be separated at all sharply. There are many deposits in which the introduction from the outside of bauxite, clayey bauxite and bauxitic clay constitutes an element of allochthony. Bauxitisation, however, could continue also after deposition, and the bauxite so formed is autochthonous already. It was Combes (1972) who first called attention to the importance of this feature. In the present writer's opinion, a detailed study of the textures and lithostructures of a bauxite permits to an approximate decision to be made as to the shares of autochthonous and allochthonous (parautochthonous) components in its makeup. Detailed investigations in this direction could greatly promote our understanding of bauxite genetics. Until then, the term *allo–autochthonous* is recommended for these deposits of combined material.

8.3.4. MODE OF TRANSPORT

Most authors envisage some form of water-borne transport. Within that category, those theories positing a longer transport usually have permanent streams in mind, whereas the adherents of the theories involving short transport routes, especially of the parautochthonous theories, prefer to assign a leading role to intermittent streams.

As regards the form of transport in water, the partisans of the chemogenic theories posit transport in true solutions. In addition to the chemogenic theories of A. D. Arkhangelsky and Goretsky, true solutions were envisaged by Haberfelner (1951) and

Kozlowsky (1956) to explain the formation of the Austrian bauxite deposits. According to them, these solutions, weakly alkaline to start with, were neutralised by carbonic acid. It was this influence that triggered the precipitation of bauxite. The trigger effect was contributed according to them by the carbonate relief.

Ruttner (1970) also assumed true solutions but, invoking a Brazilian analogy, he held a weakly acidic medium to be more probable. In Brazil, rainwater, accumulated in shallow pools in the catchment area of the Rio Negro and its tributaries, is turned acidic by the humic acids produced by the tropical vegetation. These 'black waters' can as a rule keep much more aluminium and iron in solution than the usual surface waters. According to Ruttner, the parent material of the Austrian bauxite deposits was delivered in acidic solutions such as these. The alkaline waters of the 'Gosau Sea' transgressing over a carbonate relief precipitated aluminium and iron out of the acidic solutions. A mud was settled out in places of least turbulence, and this is how the bauxite deposits came to be formed. Ruttner's theory is contradicted by the fact that Al and Fe removed in solution by the waters of the Rio Negro does not settle out to form deposits anywhere: in fact, it gets admixed with the muddy–sandy waters of the Amazonas. No bauxite deposits could therefore form in this way.

Caillère and Pobeguin (1964) hold the French bauxites to have deposited out of acidic true solutions. Aluminium, iron and titanium were the first to precipitate; silica was not added until later, out of the ground water. Their theory is not confirmed by any of the geological, petrographic or geochemical features of the deposits.

In the present writer's opinion, the formation of bauxite deposits out of true solutions could possibly take place under exceptionally favourable conditions only (under those stipulated by Ruttner). In any case, no formation out of a true solution could so far be proved for any deposit.

Others envisage transport in the form of *colloidal solutions* or of very fine-grained *colloidal suspensions* in sluggish permanent streams (Nemecz and Varju 1967; Boldizsár 1948; Jurković and Sakać 1964; Kiss and Vörös 1965; Valeton 1965). This mode of transport, however, is acceptable only for those deposits which contain no clastic textural elements. Transport is more likely anyway in colloidal than in true solutions, as present-day streams carry negligible amounts of aluminium, iron and titanium in true solution. Organic complementary colloids can substantially increase the transportability in colloid form of these elements.

It is not well understood how and when these colloidal solutions, assumed to be in a stabilised state, were precipitated. For the bauxitic clay deposits of Pilisvörösvár, Hungary, Nemecz and Varju (1967) solved the problem by assuming a simple drying out of the colloid. This mechanism, however, does not work everywhere, either. It is to be assumed that in certain deposits it was an abrupt change in electrolyte concentration and in pH that triggered the precipitation. These little-understood issues require further investigation, with the help especially of the information provided by colloid chemistry.

For the deposits of clastic texture, transport in the form of a *mud-flows* or pulp seems the most probable (Nicolas 1970; Bushinsky 1971), presumably by means of intermittent streams. Such streams are formed readily by the great downpours of a tropical–subtropical climate. Sheetwash flushing down the entire relief could also contribute. Each downpour could move the material by a couple of dozen to a hundred

metres, but these redepositions were repeated fairly often. A fact confirming this assumption is that the pebbles and fragments of bauxite found as textural elements in other bauxites are often themselves of a clastic texture, containing smaller bauxite pebbles, broken ooids and pisoids. In certain deposit groups, the writer has observed bauxite pebbles carrying traces of 3 to 5 phases of redeposition. Rapid changes of clastic textures from one place to the next in certain deposits and the presence of bauxite agglomerates made up of boulders of 0·5 to 1 m^3 size can be explained in terms of transport in torrents. Such bauxite was accumulated in its present lie after several successive phases of redeposition, as proved also by surfaces of erosion within those deposits (e.g. at Bédarieux, France).

This assumption is confirmed by an observation, made by Hill and Davis (1971), that Jamaican bauxite despite its unconsolidated nature is fairly resistant to erosion, being stable for years even in vertical bluffs. This is made possible by its very great porosity and permeability. Most of the rain falling onto the bauxite is immediately absorbed; there is practically no run-off. Only the most violent rainstorms can erode some bauxite in the form of larger or smaller lumps.

According to Zans (1959), the parent material of bauxite was transported not only in surface streams, but predominantly by underground karst rivers. It was through a special type of karst spring called 'boiling spring' that the material could return to the surface, spread on the relief and get bauxitised in place. In extensive karst regions, this mode of transport cannot be excluded, but the present writer finds it exaggerated to try to derive in this fashion e.g. the bauxite deposits of vast extent of the St Ann or Manchester plateaux. The parent material of bauxite surging through the karst springs and settling out of the spring discharge could fill only the depressions of the relief, and the deposit should have a plane horizontal surface. In the area under consideration, however, bauxite often covers the bedrock blanket-like, hugging the relief as it were. It would be even more difficult to explain by this mechanism the evolution of the bauxite deposits shown in figures 13 and 14.

The fact that no alien material, quartz etc, was added to the bauxite in the course of redeposition is explained by the fact that during bauxite formation the relief was a peneplain, precluding the introduction of alien detritus from the surrounding areas.

Under a favourable climate, bauxitisation could go on even during redeposition. The first to point this out was Vadász (1951). This is why it is so important to distinguish this *primary form of redeposition*, an element of bauxite formation, from subsequent, *secondary redeposition*. The latter took place well after the end of the phase of bauxite formation (in fact, after the erosion of the cover succession). Secondary redeposition involved no bauxitisation; on the contrary, it involved a degradation of the original bauxite by admixtures of alien material. In the course of secondary redeposition, the bauxite underwent a mechanical comminution, and its softer, more clayey parts were completely elutriated, whereas the parts richer in aluminium and iron were carried away in the form of pebbles. This is why the writer has seen it fit to distinguish the deposits due to secondary reworking as a separate deposit type (Salento-type deposits).

The *aeolian, wind-borne transport* of bauxite was stipulated by György (1923) and Telegdi-Roth (1927) to explain the formation respectively of Halimba and Gánt bauxite in Hungary. Roch (1956, 1959) derived the French bauxites from wind-borne aeolian dust. The prime reason for his assumption was the paleogeographic dif-

ficulties raised by the stream transport hypothesis. Most French specialists agree in that, at the time of bauxite formation, there was an isthmus about 250 km long and 80 km wide, called the Durancian Isthmus, linking the crystalline of the Massif Central with that of the Maures–Esterel massif. The bauxite deposits are arranged along the axis of this isthmus. According to Roch, it is hard to imagine a drainage system which, running parallel to the axis of this elongate isthmus, could have deposited the bauxite in its present lie. And if stream transport and *in situ* formation are both improbable on geological grounds, then according to Roch the only possibility left over is aeolian transport. More recently, Roch (1972) admitted the possibility that the material delivered by aeolian transport could be carried on by stream transport.

There is actually little likelihood for aeolian transport of bauxite, a rock formed under a humid climate, where a lush vegetation does not permit the weathering products to be picked up by the wind. According to Demangeon (1965, 1970), the aeolian transport of the parent material of the French bauxites is contradicted, among other reasons, also by the too large size of the clastic grains found in them. A volcanic ash fall, where the ash has been hurtled into the air by a volcano, is quite another matter. It cannot be prevented either by the climate or by the vegetation.

In summary, bauxite formation may involve a variety of transport mechanisms. Neither the presence or absence of transport, nor one or another transport mechanism can be regarded as exclusive.

8.4. SITE AND FACIES OF DEPOSITION

In the 1930s to 1950s, the question of the sites (environment) of deposition was a controversial one which seemed to offer no scope for compromise. The adherents of the terra rossa theory considered every karst bauxite to be formed on dry land, whereas the adherents of the chemogenic theory asserted most of them to have deposited in ancient seas. Opinions are rather more wide-ranging today; some deposits are assumed to have accumulated in the depressions of a dry-land relief, whereas others are regarded as the products of accumulation on valley floors, in swamps, and in permanent or ephemeral bodies of water. Yet others are held to have deposited in lagoons or close inshore in a sea. No deposit is assumed to be a delta sediment, or a pelagic sediment formed far offshore in an oceanic basin.

Origin on dry land can be proved beyond any doubt wherever a bauxite keeps on forming on dry land to this day (on the Pacific islands) or where the region has not been inundated since the bauxite formation (e.g. in Jamaica and Hispaniola). According to Zans (1954), Jamaican bauxite was accumulated in shallow depressions periodically inundated by the discharge of karst springs. On the evidence now available, however, bauxitisation extended also to the entire limestone relief (high plateaux) including its un-inundated parts.

In all the other bauxite regions, the facies of deposition must be determined by paleogeographic and paleontological considerations. Most authors interpret the fact that the bauxite overlies the more or less karsted surface of its bedrock with more or less of a hiatus by invoking an emergence. Nagorsky (1958), however, interpreted the uneven bedrock surfaces in the Devonian bauxite region of the Salair, USSR, to be

submarine coral reefs. The bauxite should thus, according to Nagorsky, have been deposited in sea water at a depth of 20 to 80 m without any emergence.

Under some bauxite deposits, the bedrock relief is hardly karsted, smooth or gently undulating. It is in these cases that the assumption of no emergence makes sense, as according to the present state of our knowledge, a karst relief can develop on dry land only.

The only known instance where the deposition of bauxite in a brackish lagoon is incontrovertibly proved is Kephanedes in the Helicon Range, Greece, where the deposit develops in continuity, without a hiatus, from a lagoon-facies bedrock limestone (fig. 27).

The most reliable way to determine the facies of deposition would be by the study of *fossils*. Unfortunately, no fossils have been found in a majority of bauxite deposits, and those that were found in the rest are also pretty sporadic. Many authors regarded this circumstance as one of the arguments in favour of a dry-land origin (e.g. Vadász 1951). The present writer feels that the preservation of the fossils originally deposited with the bauxite was prevented by the intense chemical activity of a bauxitic environment. This is confirmed by the fact that most fossils found in bauxite are poorly preserved (casts or impressions).

The interpretation of the fossils found in bauxite requires a great deal of circumspection. For example, the nannoplankton described by Báldi-Beke (1974) from the bauxite lens Tüskésmajor II, Csabrendek, Hungary indicates no marine facies, as the fossils in question were deposited together with the bauxite in their present lie. This is indicated by the inhomogeneity as to age of the fauna; in addition to the Cretaceous species making up its bulk, it invariably also contains Eocene ones, albeit in less abundance. The author cited found nannoplankton specimens of both ages in each of the samples studied by her.

In most of the Kazakhstan-type deposits, a swampy facies is indicated by the carbonaceous clay and lignite beds intercalated in the bauxite (figs 42, 44 and 48). In one of the deposits in Missouri State, US, a trunk remnant of a Stigmaria tree was found, its interior filled with fair grade bauxite (Keller 1938). This is proof for the existence of a typical swamp forest at the time of bauxite accumulation. Plant impressions indicative of a swampy environment are frequent also in the Chinese bauxite deposits. These well-preserved plant remains exhibit no traces of transport. Species of Lepidodendron and Linopteris are the most abundant (Beneslavsky 1961).

In the Mediterranean-type deposits, it is in the reduced (grey, black, white) bauxite types that fossils indicative of a swampy environment are found. In Hungary these traces include a lignitic intercalation in the Bagolyhegy deposit at Gánt (Kiss and Vörös, 1965) and root remnants in grey pyritic bauxite at Nyirád, Halimba, Szőc and Iszkaszentgyörgy. According to Pantić (1970), plant impressions are also very frequent in the grey and white bauxites of the Bijele Poljane deposit group of Crna Gora (Montenegro) Yugoslavia. They include Gleichenia, Pteridophyte and Conifer species above all. Nicolas (1970) found lumps of lignite, fragments of tree trunks and a stratified lignitic bauxite in the bottom part of the mine La Baume Nord of the Brignoles deposit group.

Oxidised bauxites have yielded a few fossils indicative of a freshwater–swampy environment. In Hungary, these included the sporangium of an Osmundacea species

found in Gánt bauxite (Kiss 1953) and the tooth fragment of a crocodile in a bauxitic clay near Olaszfalu (Noszky in Vadász 1951).

The Paludina casts found in the Senonian bauxite of the Akseki deposit group of the Western Taurus Ranges, Turkey, indicate a *freshwater–lacustrine facies* (Göksu 1954). Poorly preserved casts of gastropods *(Pyrgulifera, Melania, Strophostomella, Bulimus)* were found in the top part of the Halimba Cseres ore body by Barnabás (1961) who ascribed them to a freshwater, possibly brackwater facies. A similar facies is indicated by some bivalve and gastropod shell fragments found in calcareous bauxite near the cover of the Unterlaussa deposit group, Austria, among which the forms Glauconia renauxiara d'Orb, and Pyrgulifera pichleri Hoernes could be identified (Haberfelner 1951). Bauxite near Lesh in Albania yielded imprints of the gastropods Charydrobia and Kallomastoma species indicative of the freshwater Cosina facies (Barnabás 1961).

Nicolas and Esterle (1965) found Paludines indicative of a freshwater facies in the middle part of the Mazaugues deposit, Provence, France, whereas the top part of this same deposit yielded Rhynchonella and Terebratula species of marine facies. The delicacy of the shells precludes any transport. In this case, then, the facies changed gradually from freshwater to marine in the course of bauxite accumulation.

A *brackwater facies* is indicated by a lumachella bed 0·3 to 0·5 m thick intercalated in the bauxite of the Dréveneuse deposit in Switzerland, which yielded some well-preserved specimens of Astarte rayensis (Badoux and de Weisse 1959).

Fossils implying a *nearshore marine facies* were found in the USSR largely in the Palaeozoic bauxites of the Urals and of Siberia. In the upper, grey zone of the Devonian deposits in the Northern Urals, there are frequent Amphipores, corals, calcareous algae, brachiopods, bivalves and Psilophyta (Bushinsky 1971). There are also in places thin intercalations of organic limestone. These parts of the deposits were obviously deposited in a marine medium. The lower, red parts of the deposits, on the other hand, are regarded by Bushinsky as the sediments of small ephemeral freshwater lakes. This is proved according to him by the absence of any marine fauna, the frequent occurrence of graded bedding, and desiccation cracks observed in a few places. These latter are undeniable proofs of the temporary drying-out of some body of water.

Dementev and Rodchenko (1966), who also found in places some intercalations of an Amphipora limestone in the lower and middle parts of those deposits, consider the entire bauxite to be of a lagoon and nearshore facies on this evidence. According to them, the grey bauxite indicates a somewhat deeper facies, and the Amphipara limestone a still deeper one, the reef zone of an ancient seashore. In the present writer's opinion, the rate of bauxite delivery varied from place to place along this intricate shoreline. Some deposits may indeed have accumulated in the sea in their entirety, whereas the adjacent deposits may be of freshwater origin in their lower parts and of marine origin in their upper parts.

In the Devonian deposits of Berdsk–Obukhovsk, USSR, Nagorsky (1958) identified algae indicative of a marine facies (Rugosa, Pachypora and Stenophyllia). These deposits grade with no hiatus into the marine limestone directly overlying them. The transitional term is a bauxitic limestone of 11 m thickness at most, containing a rich coral fauna with Brachiopods and Trilobites.

In the Mediterranean region, intercalations and fossils indicative of accumulation in the sea are much rarer. The Grepča deposit of Crna Gora (Montenegro), Yugoslavia,

contains an intercalation 20 cm thick of a marine limestone. According to Radoičić (1959), the bauxite there was deposited in the sea. Likewise in Crna Gora, in the Kosijerevo and Petrovići areas, some deposits contain intercalations 20 to 50 cm thick of marine limestone (Burić 1966). Sakać (1966b) found marine gastropods, bivalves and bryozoa in the upper parts of the deposits in the Imotski and Drniš regions, Dalmatia, Yugoslavia. The fossils, typical of the Upper Eocene Promina formation, indicate a nearshore neritic facies. In the present writer's opinion, the marine facies is also confined here to the upper, fossil-bearing parts of the deposits: the lower parts were accumulated on dry land.

In summary, part of the Paleozoic deposits of Mediterranean type was deposited in a shallow sea close inshore, or in a brackish lagoon. A minor part is of freshwater origin. In the Mesozoic and Paleogene, deposits of marine and lagoon facies are less frequent, and those formed on dry land are more numerous; finally, the Neogene to recent deposits are all of dry-land facies. This gradual change implies the necessity of some qualifications concerning the principle of uniformitarianism as applied to bauxites.

Most of the bauxite deposits of Mediterranean type are found at the base of a transgressive succession. A slow ingression of the sea arrived in some of those in the midst of bauxite accumulation. In such cases, the deposits tend to be freshwater near the base, and lagoon or neritic near the top. Swamp facies are usually found only where the ingression was a very slow one indeed, at first manifesting itself only by a rise of the ground water level, so that there was time for a swamp to develop. In such cases, the actual marine ingression took place higher up in the cover succession. This is observed e.g. in the Transdanubian Hills bauxite region, Hungary.

The Kazakhstan-type deposits are swampy in the majority, less often lacustrine, with a marine ingression arriving only in the cover succession. Finally, the Ariège-type deposits are of dry-land or swamp facies, but there may be brief episodes of transgression intercalated between the episodes of bauxitisation.

On the lithological evidence available, the deposits of secondary reworking seem to have deposited on dry land without exception.

8.5. ISSUES OF BAUXITISATION

After having reviewed the fundamental factors of the origin and accumulation of bauxitic material, let us now concentrate upon bauxitisation proper. As pointed out in the introduction, bauxitisation is a special case of chemical weathering. A fairly rare bird among sedimentary processes, its formation requires the coincidence of a largish number of conditions. The process of bauxitisation can be best understood by reviewing those conditions one by one. For their assessment, let us adduce the observations made in regions of contemporary lateritisation, on the one hand, and those referring to the karst bauxites which, according to some, are still being formed today on the islands of the Western Pacific, on the other.

8.5.1. SPATIAL CONDITIONS

Bauxitisation is a process of weathering which may be regarded as a special form of pedogenesis; it is accordingly confined to surface or quite shallow near-surface conditions. It does not continue at depth after burial. This is true of both the karst and the laterite bauxite.

8.5.2. CLIMATIC CONDITIONS

It is a familiar fact that lateritisation today is restricted to a fairly narrow range of wet tropical climatic conditions. According to Sinitsyn (1976), the optimum climatic conditions of lateritization are the following.

Mean annual temperature		~26 °C
Mean temperature of hottest month		30 °C
Mean temperature of coolest month		20 °C
Annual sum of mean daily temperatures		9000 to 10 000 °C
Number of rainy months		10 to 11
Number of dry months		1 to 2
Near-surface soil temperature	in rainy season	20 to 30 °C
	in dry season	35 to 45 °C
Annual rainfall		1200 to 1500 mm
Aggregate annual insolation		160 to 180 kcal/cm^2
Aggregate annual reradiation		40 to 60 kcal/cm^2
Radiation balance		+70 to +80 kcal/cm^2
Climate type		tropical monsoon.

The intensity of lateritisation varies as the annual rainfall. It is zero below an annual sum of daily temperatures of about 4000. If rainfall exceeds 1500 mm, and especially if it is more or less evenly distributed the whole year round, then a clayey (kaolinitic) laterite is formed instead of a laterite bauxite. Sinitsyn attributes this to the fact that the regions where it rains the year round are covered by a dense rain forest. The uninterrupted accumulation and decay of vegetal debris lowers the oxidation–reduction potential in the ground, and the humic acids give rise to an acidic pH. Under such conditions, even iron and aluminium remain in solution: they cannot therefore be enriched within the weathering zone.

In the areas most favourable to lateritisation, the ground water level subsides during the month or two of the dry season following the rainy season: the weathering zone is dried out and warmed up, and processes of oxidation are initiated in it. Soil colloids coagulated and crystallise largely in the form of gibbsite, haematite and goethite. The minerals of fair crystallinity thus precipitated are not readily redissolved during the next rainy season. This fact results in their relative concentration increasing year by year. If the dry season is longer, say 3 to 4 months, then the share of iron minerals in the precipitate increases; ferricretes and iron-rich macropisoids are formed. This is how a ferrallitic laterite profile can grade into a ferruginous soil and a ferricrete.

A similar opinion is held by Maignien (1966), according to whom the optimum mean annual temperature of lateritisation is 26 °C, and a minimum rainfall of at least 1200

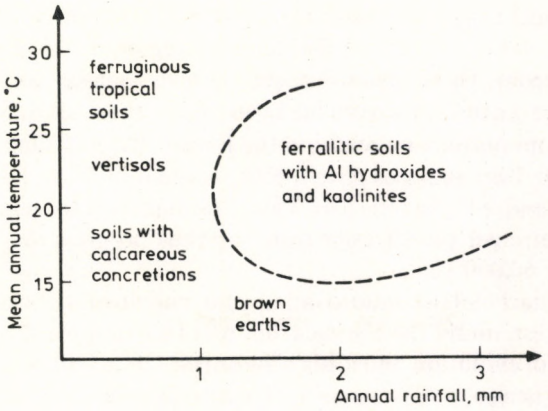


Figure 168. Connection between annual average temperature, rainfall and pedogenesis according to Segalen (1965)

mm is required. According to that author, a laterite may form even under 4000 to 6000 mm of annual rainfall, provided the relief and the drainage conditions are favourable enough, although, in the regions where rains fall the year round, clayey laterites gain the upper hand over lateritic bauxites. Segalen (1965) assesses the role of rainfall and temperature in surface weathering and pedogenesis in the manner shown in figure 168.

There is no general consensus as to the climatic conditions of lateritisation. According to Akayemov *et al* (1975), laterites will form only in those parts of the tropical zones where rains fall uniformly the year round and annual rainfall attains 9000 to 10 000 mm, under a mean annual temperature of 28 to 29 °C. Such regions are very limited on today's globe; those authors accordingly regard the intensity of present-day lateritisation to be much less than in the past. On the evidence produced by the current spate of intensive pedological research in the tropical regions, the present writer regards this opinion as exaggerated, and holds the data published by Maignien and Sinitsyn to be closer to reality.

As regards karst bauxite formation, that process must also satisfy the same conditions, by and large. In the Recent karst bauxite deposit groups of the Western Pacific islands, mean annual temperature is 23 °C, and annual rainfall is 1600 mm, confirming this assumption. In the writer's opinion, similar climatic conditions also prevailed during karst bauxite formation in the past.

8.5.3. ROLE OF THE PLANT COVER

Most of the lateritic bauxite regions regarded as recent are situated where the tropical forest passes into the savannah zone, but some extend down into the forest zone. On the Pacific islands, karst bauxite is covered by a dense but low forest. Bauxite is penetrated by a dense skein of roots and rootlets; the longer roots even reach down into the bedrock (Tercinier 1971).

According to Lovering (1959), some tropical forests can extract about 2000 tons of silica from a one-acre area (0·4 hectare) in 5000 years. This is a very rapid desilication

indeed on a geological time scale. Mikhailov (1964) made similar observations in the Guinean Republic, where most of the bauxite region is covered by tall grass, Pennisetum purpureum. This is usually burnt in the dry season, its ashes being blown away by the wind or washed down by the rains of the wet season. Mikhailov cut and burned the grass from one square metre in the Fenaria region. One kilogram of grass gave 6 to 7 g of ash. Two analyses gave a SiO_2 content of 69 and 72% but an Al_2O_3 content of only 1·9 and 2·1% for the ash. This is tantamount to the removal of 25 g of silica per square metre per year. Desilication can thus progress one centimetre in 600 years, one metre in 60,000 years.

According to Erhart (1956), lateritisation and bauxitisation can only take place under a tropical forest, under the chemical action of a living plant cover. He calls this hypothetical phase of evolution 'biostasis' (*'biostasie'*). It is followed according to him by a phase of 'rhexistasis' (*'rhexistasie'*), characterised by the decay of the tropical forest as a result of a climatic change or of an uplifting; it entails the destruction of the laterite by erosion. Erhart states all karst bauxites to be weathering products, reworked in a phase of rhexistasis and deposited onto a carbonate relief. They did not, in his opinion, undergo any subsequent bauxitisation. He thus attributes to the tropical forest a unique bauxitising role.

The present writer regards Erhart's views as exaggerated, since bauxitisation may be brought about also by purely chemical, inorganic factors. Bauxite deposits were formed as early as the Late Precambrian and Cambrian, when the land was not yet covered by a vegetation worthy of the term. More likely, the role of the forest is an indirect one, the micro-organisms living in the forest soil promoting the decomposition of the alumosilicate minerals and the dissolution of silica. Unfortunately, these microbiological processes are not understood in any detail as yet. The micro-organisms may have decomposed the organic acids formed in the course of vegetal decay, thus forestalling too acidic a pH in the soil. They may further have played a catalytic role in certain phases of bauxitisation. According to Miholić (1956), it is the micro-organisms that break up the kaolinite lattice, making desilication possible in the first place.

Maignien (1966) attributes an important role to the protection afforded by the plant cover in preventing the erosion of the soft weathering products by the tropical rainstorms.

8.5.4. GEOMORPHOLOGICAL FACTORS

Karst bauxite formation requires a relief where, in the course of primary redeposition, no alien material (e.g. quartz sand) can get admixed to the bauxite. This is why most authors stipulate a peneplained relief. A flat cone karst near the seashore could also satisfy these conditions. It was presumably separated from the hills in the interior of the continent by a plain or a gently rolling landscape which trapped the detritus descending from the hills.

Even more often, bauxite formation could take place on smaller or larger islands emerging from a shallow sea, where there were no high mountains at all, so that no coarse detritus could form to start with.

8.5.5. HYDROGEOLOGICAL FACTORS

One of the requirements of bauxitisation is that the rock should be percolated by as much water as possible, in order to promote chemical alteration by hydrolysis, and in order to remove in solution some of the alteration products (silica, alkalies, alkali earths, etc). This is why we have insisted so much on annual average rainfall among the climatic conditions. Provided there is sufficient rainfall, percolation is ensured by the extraordinary porosity and permeability of bauxite (cf Section 4.5.4).

Balkay (1973), analyzing the hydrogeological factors of bauxitisation, pointed out the importance of examining the processes of *dissolution* separately, which is basically a chemical process, and *drainage*, which is a decidedly physical one. According to him, lateritic bauxite deposits are characterised by a centrifugal (lateral outward) drainage, whereas in karst bauxites drainage is predominantly downward, a phenomenon made possible by a karsted and fissured, high-permeability carbonate bedrock. Inward (centripetal) drainage is a rarer phenomenon of limited extent. It characterises those deeper, doline-like karst bauxite deposits where percolating waters leave through a central sinkhole at the bottom.

According to Balkay, centripetal drainage seldom acts in its pure form, as the karst fissures of the bedrock usually permit drainage in other directions also. In the present writer's opinion, centripetal drainage may have played an important role in the sinkhole-type and contact deposits of Kazakhstan type, as indicated by the concentric arrangement of bauxite thickness and grade in those (figs 44 and 46).

For leaching and drainage to take place, a deposit must lie above the karst water table. In the deposits under the ground water table, where the rock was saturated with stagnant or very slowly moving ground water, the progress of bauxitisation was slowed down greatly or came to a dead halt.

One requirement of intense leaching is that the bedrock must permit a satisfactory drainage. This condition is fully satisfied by the pure limestone and/or dolostone bedrock of the karst bauxites. The writer has observed in several bauxite regions the relationship that bauxite grade is best, silica content is lowest above the places where the limestone or dolostone bedrock is strongly karsted, is traversed by dense networks of open fissures and karst passages, or has been shattered by structural stresses. It is in such deposits that subsequent karstification is strongest. A fair case in point is the contrast between the Nyirád and Halimba deposit groups, Hungary. In the Nyirád region, the bedrock is a shattered dolostone of excellent permeability, whereas at Halimba it is a such less fractured dolostone and limestone, clogged up with clay, of much poorer permeability. In the former region, the majority of the deposits is made up of high-grade bauxite, whereas in the latter the average silica content of the bauxite is much higher. The two deposit groups are not far apart; their geological age is near-identical, and their accumulation took place in similar palaeogeographic settings.

Requiring an intense percolation and the removal of the leached-out components, bauxitisation is a process inconceivable in the sea or in a lagoon. The material of the deposits accumulated in the sea or in a lagoon would be washed in as fully developed bauxite from the nearby land, and its present site is merely a site of accumulation.

8.5.6. GEOCHEMICAL AND PHYSICO-CHEMICAL FACTORS

According to a thermodynamic analysis performed by Fritz and Tardy (1973), the most important geochemical factors of bauxitisation are the pH, the temperature, the initial composition of the percolating rain water (ground water), and the partial pressure of the carbon dioxide gas dissolved in it.

Szádeczky-Kardoss (1955) specified the essence of bauxitisation with respect to its geochemical nature as (1) an increase in oxygen number, (2) water intake, and (3) the development of bonds of greater ionicity.

In regard of the Hungarian bauxites, similar opinions have been expressed by Barnabás (1966) and by Vendel *et al* (1971). Széky-Fux (1961) and Széky-Fux and Szepesi (1959) proved that $Ca(OH)_2$ will form on the dissolution of finely disperse $CaCO_3$ in water containing no CO_2. This process lends a strong alkalinity to ground water. The calcium hydroxide thus formed exerts a strong digestive influence on the clay minerals, as a result of which amorphous silica and aluminium hydroxides are produced. There is a considerable likelihood for such conditions to have in fact prevailed in the karst bauxite deposits.

Weathering in a stabilised (buffered) acidic medium, on the other hand, results in the congruent dissolution of kaolinite. The Al/Si ratio in the rock is not affected thereby, and no bauxitisation takes place. This is the geochemical explanation for the fact that, in those tropical regions where rains fall the year round, a kaolinitic zone of weathering is formed instead of a lateritic bauxite. Under a temperate climate, the ground water is maintained at a pH much more acidic than what is needed for bauxitisation by the slower decay of organic matter and by the greater influence of humic acids. Nor is the temperature high enough. Under such conditions, aluminium is leached more readily than silica. The Al/Si ratio in the rock is displaced in favour of Si; in fact, silica may precipitate out of the ground water. This process leads to the *podzolisation* of the weathering zone (Fritz and Tardy 1973).

The dissolved CO_2 content of the percolating ground water increases the acidity of the solution and hence also the solubility of aluminium, promoting thereby the congruent dissolution of kaolinite.

The importance of an oxidising environment was already emphasised in the chapter on mineralogy. Bauxitisation can take place also under reducing conditions, but desilication is less intense there. This is why the syngenetically reduced bauxites invariably contain more silica than the oxidised bauxites of the same deposit.

Knyazeva (1966) postulates a peculiar form of diagenetic bauxitisation for the Ivdel deposit group, Northern Urals, USSR. According to her, strong organic acids were formed by the interaction of groundwater with some bituminous organic matter in the immediate cover and in the upper part of the deposit. These acids decomposed the alumosilicate minerals and removed the silica in solution. There was a concentration of diaspore, pyrite, marcasite, siderite and calcite. Bauxite was recrystallised; its original clastic texture was effaced, Knyazeva regards as evidence the fact that these alterations were much weaker wherever the cover of the bauxite is a limestone of lower organics content. In the present writer's opinion, the process involved there must be a peculiar form of action by a localised geochemical environment, as this alteration is absent from other deposits whose immediate cover includes limestones rich in organics.

In addition to the pH and Eh, water temperature is another important factor of bauxitisation. It was found experimentally by Okamoto *et al* (1957) that the solubility of amorphous silica gel increases more than threefold in the temperature range from 0 to 58 °C. According to Krauskopf (1967), not only is solubility increased, but so is also the rate of dissolution.

According to a thermodynamic analysis by Fritz and Tardy (1973), the solubility of gibbsite increases with temperature only in an alkaline medium; in a weakly acidic one it actually decreases!

Another important condition is that the water percolating the rock must be unsaturated or free of silica, as its dissolution potential is a function of its silica content. If the water percolating the bauxite is largely rainwater, then this condition is satisfied, and the full dissolution potential of the rainwater will promote bauxitisation.

8.5.7. COLLOID CHEMICAL FACTORS

Bauxite being a colloidal system, the laws of colloid chemistry also apply to the bauxites. It is these that control the peptization or coagulation of the individual components in the course of weathering. Dissolution also takes place primarily in the form of colloidal solutions.

Kiss and Vörös (1965) attribute a special importance to colloidal segregation. According to them, colloidal suspensions of allitic and siallitic composition were delivered simultaneously to the Gánt bauxite area in Hungary: once there, however, they coagulated and settled separately, as a function partly of their respective specific gravities, and partly of the pH of the environment. Kiss and Vörös assume that no significant desilication took place during and after settling. The possibility of such colloidal segregation does exist, but it is impossible to interpret the formation of the Gánt deposits in terms of this process alone. The frequent coarse clastic textures found in this bauxite, including conglomerates and agglomerates, and the graded bedding observed in parts of some of the deposits all imply that, in addition to colloidal segregation, an inflow of bauxite in clastic form also took place. Instead of a fine suspension, a coarse, dense slurry should be visualised, delivered by tropical torrents into the depressions of the relief.

8.5.8. COMPOSITION OF THE ROCK EXPOSED TO BAUXITISATION

In lateritic bauxites, processes of direct and indirect bauxitisation can be distinguished (Balkay and Bárdossy 1967). In the former process, the alumisilicate minerals (feldspars, melanic silicates) of the parent rock are decomposed directly into Al and Fe hydroxides. In the latter, bauxite minerals are formed by the intermediary of clay minerals. The former is predominant where leaching is very strong and drainage very good, and the latter where both are less intense.

In karst bauxites, direct bauxitisation may be assumed to have been operative in the bauxitisation of volcanic ash fallen onto a carbonate relief above all. In all other cases, bauxitisation passed through a clay mineral stage. The same opinion is held in regard of the Hungarian bauxites by Szádeczky-Kardoss (1955), Fülöp (1964) and Barnabás

(1966). In many instances, the entire aluminium content of the parent rock was in the form of clay minerals.

In bauxitisation, the clay minerals are decomposed into amorphous weathering products. From these, silica is removed in solution by the percolating waters, but it can find opportunities to reprecipitate as kaolinite. In this respect, the findings of de Kimpe *et al* (1964) are of considerable importance: their model experiments have showed that, under Earth-surface temperature and pressure, kaolinite is readily synthesised in a weakly acidic medium containing no alien ions.

In the recent karst bauxites of the Pacific islands, on the other hand, alkali-earth ions (Ca^{2+}, Mg^{2+}) are invariably present; this hinders the crystallisation into kaolinite of the amorphous weathering products (Tercinier 1971). According to Siefermann (1969), the high organics content of these bauxites acts in the same direction. A carbonate bedrock, then, promotes bauxitisation also by hindering the crystallisation of kaolinite and hence also the slowing down or stoppage of desilication.

There is today a general consensus that clay minerals do in fact get decomposed in the course of bauxitisation. Kaolinite is the only mineral that was held by some—e.g. Erhart (1956)—to be so stable a product of weathering as not to be further decomposed even by the bauxitisation process. In refutation of this idea, it was experimentally proved by Pedro and Berrier (1966) that kaolinite is in fact decomposed into amorphous silica and bauxite minerals in the course of bauxitisation. Silica is removed in solution. The assumption of Pedro and Berrier was confirmed by Fritz's and Tardy's thermodynamic calculations (1973).

Of great importance for our subject are the experiments carried out by Pedro (1970) on a marl consisting of 37% calcite, 59% illite and a few percent of quartz and feldspar. He percolated his sample with water of some 65 °C temperature, saturated with carbon dioxide, at a rate of four litres per day for three years in a Soxleth device. All the calcite was decomposed in the first half-year. At the end of the first year, a little gibbsite appeared in the rock; its abundance gradually increased to the end of the experiment, at which time it attained 15%. The principal teaching of the experiment is that carbonate decomposition is a much more rapid process than bauxitisation proper. Consequently, the decomposition of carbonate rocks produces clay minerals as a first stage; it is these that are further altered into bauxite. (This succession was stipulated by Vadász on geological grounds as early as 1951.) The experiment has further proved that marls can also get bauxitised.

The adherents of the terra rossa theory regard as an intermediary term a red clay from which the carbonate minerals have already been removed but which is still in the early stages of desilication. Pedro's above-cited experiments have confirmed the viability of this process.

8.5.9. THE TIME FACTOR

According to numerous authors (e.g. de Weisse 1964), bauxitization is a very slow process even on a geological scale; geological evolution must be fairly tranquil during all that time. According to Bushinsky (1967), an emersion of 2 to 5 million years as a minimum is required for bauxite deposits worthy of the term to form. Observations on recent bauxites, on the other hand, prove that, under suitable conditions, bauxitisation

can be very rapid; very seldom, its progress can be observed even within a few years (Balkay 1972, oral communication). The youngest lavas on Hawaii which already carry a bauxitic profile are about 10 000 years old (Valeton 1972).

Thermodynamical calculations by Fritz and Tardy (1973) showed that, under favourable conditions of climate and drainage, a gibbsitised laterite zone can grow down one metre in 350 000 years on quartz-free basic igneous rocks. On granites, the gibbsitic laterite zone can thicken by 32 cm in 100 000 years: in the process, a kaolinitic zone 12 cm thick is formed under it. The full dissolution of quartz from a weathering zone one metre thick takes 210 000 years in theory. These figures are, of course, just the most likely rates of advance in the most general case. There may be substantial deviations depending on the fracturing and permeability of the rock and on the grain size and crystallinity of its minerals. The order of magnitude of the advance of lateritisation is illustrated quite well by these calculations, however.

Some of the karst bauxite deposits are found in very brief hiatuses of 1 to 5 million years: these deposits may nevertheless contain large tonnages of fair grade bauxite. A long time on a geohistoric scale then, is no absolute precondition for the formation of a karst bauxite.

8.5.10. MODE OF PROGRESS OF BAUXITISATION

In most lateritic deposits, bauxitisation progresses from the top downward and laterally outward, along the directions of drainage; it may be intensified or retarded in places, depending on the composition of the rocks encountered. As a result, a zone of advanced bauxitisation can underlie a zone of rudimentary bauxitisation (Balkay and Bárdossy 1967).

In numerous karst bauxite deposits, bauxitisation must have progressed from the top downward, as the upper parts of these deposits are lower in silica than their lower parts.

Numerous deposits, on the other hand, are of a concentric structure: the highest-grade bauxite at the middle is surrounded by bauxites that become more clayey in every direction, the outermost zone being made up of bauxitic or bauxite-like clays. Vendel *et al* (1971) interpret this phenomenon by assuming a resilification of the bauxite from above, from the sides and from below by a humus-rich ground water percolating the carbonaceous clays of the immediate cover. The deposit was argillised in its entirety in those places where it was thinnest: this is why the lateral, tapering parts of the deposits are the most clayey.

The above arrangement, however, may also exist where there is no trace even of carbonaceous clays (of a swampy environment) in the immediate cover. The present writer accordingly prefers the explanation that, in the period of bauxite formation, the bauxite deposit was entirely saturated by ground water after each heavier rainfall. According to Tercinier (1971), the recent karst bauxites of the Loyalty Islands are characterised by extreme capacities of absorption and water retention, and the same is true of the Neogene bauxites of the Caribbean. The older karst bauxites were probably much more similar prior to their lithification. After each rainfall, then, the limestone bedrock and its environment dried out sooner than the bauxite deposit. Within the deposit proper, the near-surface parts dried out first, of course. The ground water

solutions saturating the deposits then started to seep outward, down the 'humidity gradient'. In the course of this diffusion, they carried along the silica dissolved in them. Depending on the climatic conditions, this process was repeated more or less often: as a result, it was the deposits central part that got bauxitised most strongly.

This is not to deny the importance of secondary resilication; in the present writer's experience, however, that process tends to act from the top downward: it can be recognised by both the textural and the mineralogical changes it entails, and thus be distinguished from the above-outlined process.

In addition to bauxitisation progressing downward and outward radially, some less intense upward-acting bauxitisation can also be conjectured according to Vendel *et al* (1971), its vector being the colloidal solutions enriched in Al and Fe, rising by capillary action in the dry seasons.

8.5.11. DIFFERENCES IN BAUXITISATION BETWEEN DEPOSIT TYPES

The processes of bauxitisation in the lateritic and karst bauxite deposits are very similar but not entirely identical. The climate, the plant cover, dissolution and the time factor play identical roles. The principal differences are as follows.

In lateritic bauxites, bauxitisation is a predominantly *in situ* process; that is, it is basically eluvial. In karst bauxites, it goes on also during the repeated primary reworkings, and in places even after the final deposition. Bauxitisation takes place in a neutral or weakly acidic medium in lateritic bauxites but in a weakly alkaline one in karst bauxites. In the lateritic bauxites, ground water is totally 'un-ionised', whereas in the karst bauxites it invariably contains alkali-earth ions. As we have seen above, these divergences may entail substantial mineralogical differences.

It finally seems that syn- and diagenetic and epigene internal chemical rearrangements (mobilisation) were much stronger in the lateritic bauxites than in the karst bauxites. Their conspicuous products include a ferricrete, a zone of concretions and some collomorphic textural types: all these are much rarer in karst bauxites (cf the chapter on Petrography).

The writer perceives the principal cause of these differences in the fact that lateritic bauxites get bauxitised above an alumosilicate bedrock, and karst bauxites above a more or less pure carbonate bedrock. All this affects the hydrogeological, geochemical, physico-chemical and mineralogical conditions of bauxitisation.

The bauxitisation process, however, is not quite uniform even within the karst bauxite category. Bauxitisation in the *Ariège-type deposits* is an exclusively eluvial process: this is the process that stands closest to lateritic bauxite evolution. The influence of the carbonate bedrock, however, can be demonstrated also in this instance: this is why the writer has placed these deposits among the karst bauxites.

The parent material of the *Kazakhstan-type deposits* consists in its bulk of the erosion products of laterites. Redeposition involved such a short distance of transport that no significant bauxitisation could take place during that process. Bauxitisation after the deposition of the material was quite intense, however. Post-depositional bauxitisation was incontrovertibly proved the microscopic examinations of Lisitsyna and Pastukhova (1963). This process much resembles the formation of lateritic

bauxites; in places, ferricretes came to exist on the top of the bauxite lenses. Textures typical of lateritic bauxites are also the most frequent in these deposits. Each episode of bauxitisation was followed by intensified erosion and the accumulation of material of a more detritic nature. This is how the peculiar composite ('storied') structure of the Kazakhstan-type deposits came to be developed. The influence of the carbonate bedrock, however, is also felt here, especially in the sinkhole and canyon-type deposits. Bauxitisation was strongest in the central parts of the dolines or canyons, where drainage was best.

Karst bauxite formation is at its most typical in the *Mediterranean-type deposits*. This is where the influence of the carbonate bedrock is strongest, in the hydrogeological and geochemical as well as in the depositological respect.

It was in the *Timan-type deposits* that the role of *in situ* bauxitisation was probably least, as implied by the weak karstification of the bedrock relief and the stratified nature of the deposits. Bauxitisation presumably took place in nearby laterite areas, and it was the fully developed bauxite that was redeposited in the present lie. Bauxitisation after redeposition must have been of a very limited extent.

The *Tulsk-type deposits* represent a very rare and peculiar form of bauxitisation. Sulphuric acid produced by the surface- and near-surface decomposition of pyrite contained in carbonaceous clay and coal formed solutions that dissolved the aluminium from the rock and, seeping downwards, reached the limestone bedrock after a travel of a few metres. The sulphuric acid was neutralised there by the limestone, and the aluminium was precipitated in the form of gibbsite. This hypothesis is fully borne out by the deposit features of the bauxite (fig. 56). There is pyritic carbonaceous clay everywhere in the cover: the largest uninterrupted bauxite bodies occupy depressions in the limestone bedrock, where its physico-chemical influence could act on the largest possible surface. Further up, in the ochroid clay, there are sporadic nests of gibbsite identical in appearance and composition with the bauxite in the deposits.

The above-outlined process of formation is also confirmed by the absence of titanium from these deposits: the solutions did not attain the very strong acidity (a pH from 1 to 2·5) required for the dissolution of titanium. That this process could take place only in the proximity of the ground surface is proved by the fact that bauxite gradually disappears as the depth of the limestone bedrock increases (Trubina 1958).

8.5.12. THE ROLE OF THE KARST IN BAUXITE FORMATION

(i) Karst depressions served as traps collecting the parent material in the accumulation of bauxite. In the process of its reworking also, the bauxitic parent material could most readily accumulate in the karst depressions of the relief. According to Bonte (1965), the parent material 'slid' by degrees from above a non-karsting carbonate bedrock (marl) into depressions in the surface of rocks amenable to karstification. This 'sliding' is in the present writer's opinion a process equivalent to primary redeposition.

(ii) Karst depressions provide a fair drainage. The intense drainage and percolation by rainwater required for bauxitisation are realised there to the greatest degree.

(iii) The weakly alkaline medium so favourable to bauxitisation was provided by the carbonate bedrock. The more rugged was the surface of the limestone, the more intense was its karstification: the greater was the surface exposed to the rains, the more readily could a weakly alkaline medium be generated.

(iv) The depressions of the karst relief could protect the bauxite deposits from subsequent erosion. In numerous areas of intense erosion, bauxite was preserved exclusively in the karst dolines where the protection from erosion was sufficient; it was eroded from the rises separating the sinkholes.

8.5.13. RELATIONSHIP BETWEEN BAUXITE FORMATION AND PEDOGENESIS

Many scientists regard bauxite as a sort of soil (or fossil soil, as far as the older bauxites are concerned), and bauxitisation as a form of pedogenesis. This, however, is only partly correct. Bauxite formation, notably, cannot simply be equated with chemical weathering; nor can the weathered zones of the tropical regions, 100 to 200 m deep in places, be regarded as soils. According to the pedologists, notably, the *soil profile* zone extends only to a depth of a few metres under the ground surface, as deep as the influence of the plant cover, of the roots and soil bacteria can be demonstrated to extend.

Bauxitisation proper does involve a peculiar form of tropical–subtropical pedogenesis, often termed *ferralitisation* in soil science. Karst bauxite formation, however, involves also reworking and redeposition in karst depressions, and these processes taken together should be regarded as a process of sedimentary rock formation.

8.6. THE EVOLUTION OF BAUXITE FORMATION AND HISTORICAL GEOLOGY

This section is to be concerned with the evolution of bauxitisation in the course of geohistory; at the same time, it is to be an attempt at clarifying the factors controlling the evolution over geohistorical time of the process of bauxite formation.

8.6.1. PLATE TECTONICS AS AN EXPLANATION FOR THE DISTRIBUTION OF BAUXITE FORMATION

The geographic distribution of bauxites raises a number of still unsolved geohistorical issues. Bauxitisation is confined to humid tropical and subtropical climate. The climatic zone having these features is situated today between about the 30th parallels, south and north. It is in this zone that almost all the recent Pleistocene and Neogene laterites, lateritic bauxites and karst bauxites are found. The majority of Paleogene and Mesozoic deposits lie in today's temperate zone, between 30 and 50° of northern latitude. The Paleocene and Cretaceous deposits of Siberia are even farther north, between the 50th and 60th parallels (fig. 3). The Palaeozoic bauxites of the Urals

and of the Timan Mountains are between the 50th and 70th parallels: in fact, the recently discovered deposits of the Pripolyarny (= Praepolar) Urals reach up to the Artic Circle. The northernmost bauxites known today are the Timan-type deposits of the Taimyr Peninsula: they are between the 74th and 77th parallels (Mikhailov *et al* 1973; Budnikov 1971). Their formation can be explained in one of the following ways:

(i) By the assumption that, in the Mesozoic and Paleozoic, bauxites could form under climates colder than today (e.g. by invoking Erhart's (1973) theory of bauxite formation by podzolization). This is contradicted by the geochemical mechanism of bauxitisation, a process requiring warmth and humidity. Furthermore, the fauna and flora found in the immediate bedrock and cover of these deposits also indicate a warm climate. This option must therefore be rejected.

(ii) By the assumption that, in certain geological periods, the climate was warmer all over the Earth than it is today. There are workers who stipulate globally warm and globally cool periods to have alternated in the course of the Earth's history. In the former, the climatic zone suitable for bauxite formation could extend very far from the equator on either side of it. According to Mikhailov (1973, in a letter to the writer), the Lower Carboniferous was such a period, and so was the transition between the Upper and Lower Cretaceous.

Global climatic fluctuations did repeatedly take place in the course of the Earth's history—it is enough to recall in this context the great glaciations of the Pleistocene and the interglacials separating them. There exist today means for assessing the climatic fluctuations of the earlier geologic periods. Oxygen isotope analyses performed on the shells of various marine organisms, e.g. permit inferences of approximate average temperatures for the water in which these creatures lived. A curve of marine temperatures for the West European Mesozoic was constructed by Bowen (1961, 1966) and others from measurements performed on the rostra of Belemnites. A palaeotemperature curve was constructed by Góczán (1963) for the Central European region on the basis of palynological findings. The two curves exhibit a fair fit apart from minor deviations (fig. 169).

It is thus conceivable that in the warmer periods the climatic zone suitable for bauxite formation extended north of the 30th parallel, perhaps even up to the 40th or 50th parallel. These temperature fluctuations of a few degrees, however, cannot by any stretch of the imagination explain the existence of bauxite deposits beyond the 70th parallel. The palaeoclimatological, meteorological and geophysical evidence does not admit any hypotheses by which in certain periods, the tropical zone should have extended beyond the polar circle. The causes of the climatic fluctuations (changes in albedo, shifts in marine currents, astronomical changes, changes in the CO_2 content of the atmosphere, the volcanic dust content of the stratosphere etc) were insufficient to provoke such deep-reaching changes (Brooks 1949; Schwarzbach 1961).

(iii) Assuming a shift of the terrestrial crust as a whole vis-à-vis the poles, which would have resulted in gradual displacements of the climatic zones while preserving the continental configuration of today was Goretsky's idea (1960), according to whom the equator passed through the Northern Urals in the Devonian, parallel by and large to today's equator. In the successive geohistorical periods, the equator was gradually displaced southward, into its present position. A similar conjecture is implicit in

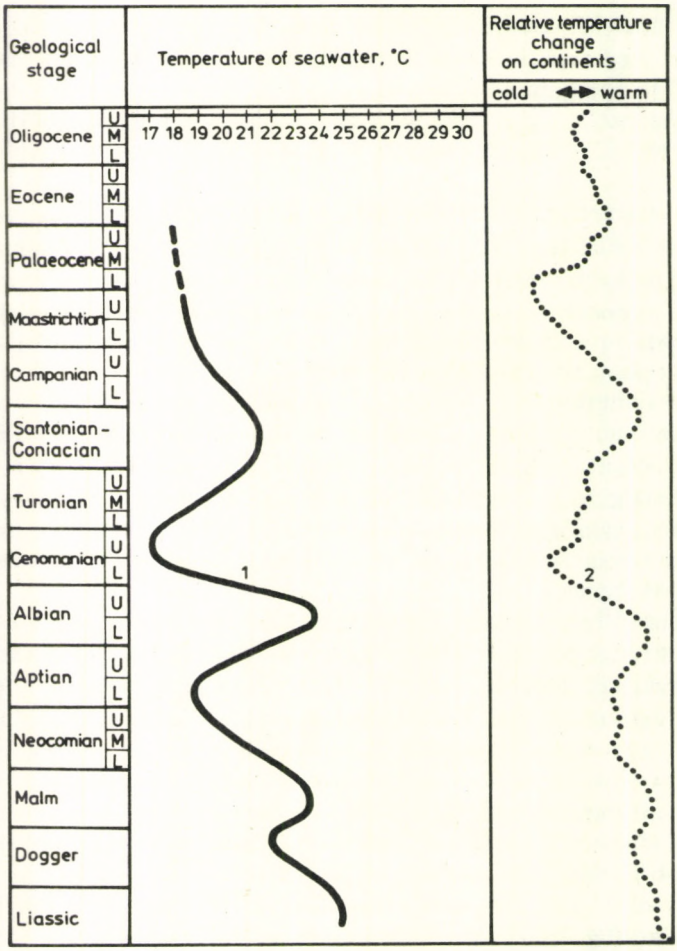


Figure 169. Paleotemperature curves for Western and Central Europe, according to Góczán (1973)
(1) marine temperatures in Western Europe, derived from oxygen isotope analyses on Belemnoid shells; (2) temperature curve for Central Europe, derived from palynological evidence

Strakhov's (1960) paleoclimatological and paleogeographic map of the Carboniferous (fig. 170) which, however, contains a number of paleoclimatological contradictions (in regard e.g. of the Siberian coal and Chinese bauxite deposits). This, however, is not the author's fault: it is impossible to get all the geological data to fit wherever one places the pole, as long as one insists on today's continental configuration.

(iv) Hence, it is only the changes in the relative positions of the continents, that is, Wegener's idea of continental drift, that can provide a satisfactory explanation for the origin of all the bauxite regions of the world. The familiar, modern interpretation of this hypothesis was provided by what is called the '*theory of plate tectonics*' (Dietz and Holden 1970; Dewey and Bird 1970; Isaaks *et al* 1968; Le Pichon 1968; Le Pichon *et al* 1973; Uyeda 1972; and Tarling and Runcorn 1973).

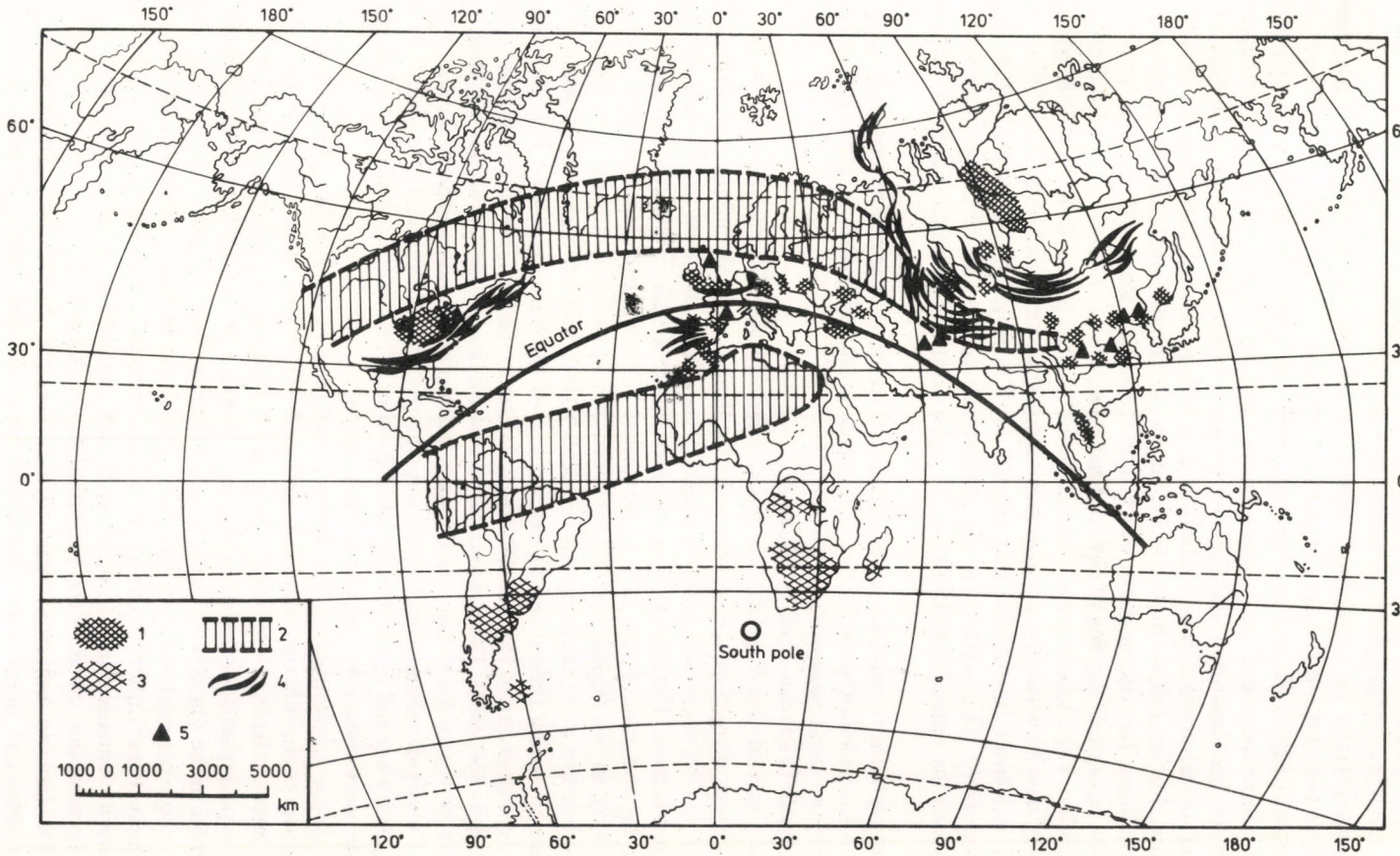


Figure 170. Paleoclimatological map of the Middle and Upper Carboniferous. After Strakhov (1960)

(1) coal measures; (2) arid zone; (3) glacial deposits; (4) orogens; (5) bauxite deposits

In a paper published in 1973, the present writer made an attempt at a global explanation of the geographic distribution of bauxite deposits on a plate tectonic basis. No reconstructions of continental positions for the Palaeozoic, however, were available at the time, and so the writer was obliged to derive palaeoequators for the principal continental plates from their respective palaeomagnetic poles. In this way, climatic zones favourable to bauxite formation could be outlined. The agreement obtained for the Cambrian and Devonian bauxite regions was quite good even so.

A series of reconstructions by Smith *et al* (1973) permitted a much more accurate evaluation. Let us follow now the evolution of bauxite formation by using these reconstructions. Bauxite deposits older than Cambrian were not included into this evaluation because of the lack of any continental reconstructions from this era. Early in the Cambrian, karst bauxites were formed in the Eastern Sayan Ranges, USSR. Minor deposits of lateritic bauxite were also formed in the Lower and Middle Cambrian, out of basic volcanics to the south-southeast of Krasnoyarsk (Bronevoy 1974, oral communication). The corresponding continental reconstruction places both sets of deposits into the immediate vicinity of the palaeoequator, between it and 10° latitude south.

According to the reconstruction of Smith *et al*, the minor karst bauxite deposit group formed in Kazakhstan in the Upper Ordovician presumably also lay in the immediate vicinity of the palaeoequator.

No bauxite is known from the Silurian.

The first sizeable karst and laterite bauxite deposits were formed in the *Devonian*, in the Hercynian orogen of the Urals, in the Central Timan Mountains. According to Hamilton (1970), there was, along the axis of the Urals an island arc formed along a zone of subduction. The arc had been attained by the Russian platform in its clockwise rotation by the end of the Lower Devonian. The bauxite deposits were formed subsequently, in the Middle and Upper Devonian. It is clearly apparent in figure 171 that this bauxite zone, striking north-south today, had an east-west strike in the Devonian, at which time it lay in the vicinity of the palaeoequator. The Devonian bauxite deposit group of Northern Pakistan lay far to the East, but likewise close to the equator (Stauffer 1969). The continental reconstruction allows the conjecture that the formation of minor karst bauxite deposits elsewhere on the southern shores of the Tethys. Finally, the Salair bauxite region was situated on the northern shore of the Tethys, hardly beyond the 30th parallel of the time. The climatic conditions of bauxite formation were thus also satisfied in this region.

In the *Lower Carboniferous*, the clockwise rotation of the Russian platform and its edging closer to the Siberian platform continued. It is in the Lower Carboniferous that sizeable lateritic bauxite deposits appear (Bielgorod, Severoonezhsk), together with Tikhvin-type deposits formed at the expense of their erosion (Tikhvin). Karst bauxites also formed, in the Timan and Turkestan Mountains. A weaker lateritisation produced bauxitic clays beneath the coal measures (Ayrshire, Scotland; Decazeville, France, Slask [Silesia] and Lublin, Poland). In the US, karst bauxites formed in Missouri and minor lateritic bauxite deposits in Pennsylvania. The map in figure 172 shows all these regions to have lain close to the Carboniferous palaeoequator.

Sizeable bauxite deposits were formed on the North and South Chinese paraplatforms and minor karst bauxite deposits on the Taimyr Peninsula. These,

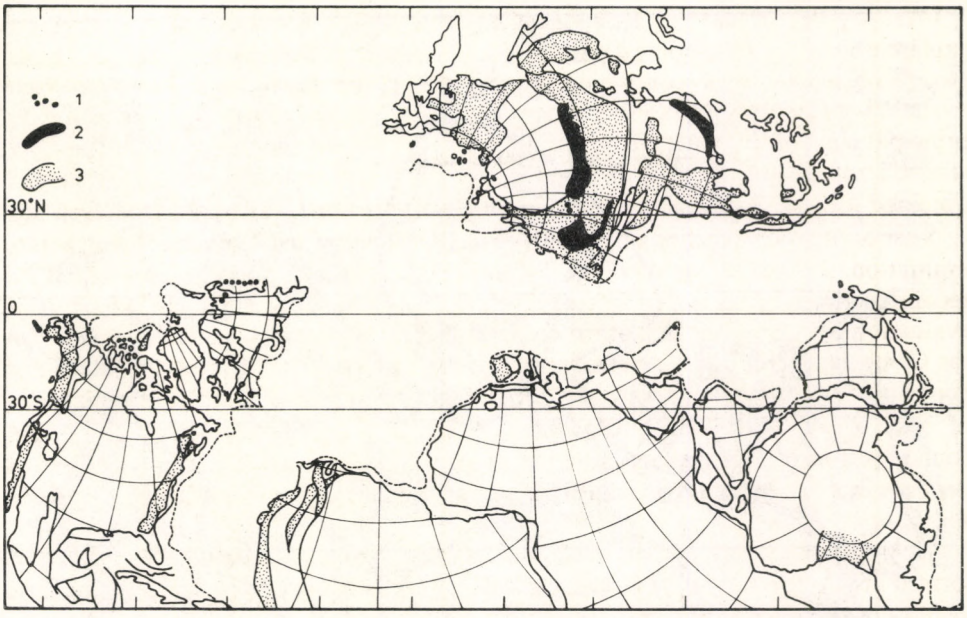


Figure 171. Lower Devonian continent reconstruction (about 380 m.y.). After Smith *et al* (1973) completed by bauxite deposits

(1) bauxite deposits; (2) Caledonian orogenic belts; (3) Hercynian orogenic belts

however, would have lain far to the north of the paleoequator according to the original continental reconstruction (between the 40th and 60th parallels, in the temperate zone where the formation of bauxite is hardly conceivable). The contradiction is explained by the modus operandi followed in the reconstruction. In the words of its authors; 'In the Paleozoic, some large continental areas for which paleomagnetic data are lacking were separated from the principal fragments. For example, China is now joined to Siberia via a Hercynian orogenic belt. It was presumably separated from Siberia in pre-Hercynian time by an ocean, but we have not discovered any paleomagnetic data that would allow us to reposition China in pre-Hercynian time. Without such data China can be repositioned only by using geological data, which we have deliberately not used.'

In view of these considerations, the present writer has rearranged on the map the Siberian platform and the paraplatforms of Tibet, Tarim and North and South China in their most probable positions prior to the folding up of the Hercynian orogens (cf fig. 170). In the writer's opinion, the East Asian paraplatforms—microcontinents in the terminology of plate tectonics—drifted to the north similarly to the Indian microcontinent, to coalesce one by one with the great Siberian continental plate. This assumption is borne out among others by the 'Geotectonic Map of Eurasia', published in Moscow in 1966, according to which, in the course of a trip due south from the Siberian Shield, a traveller would successively traverse younger and younger orogens, Baikalian, Caledonian and Hercynian. The folding up of the orogens took place the

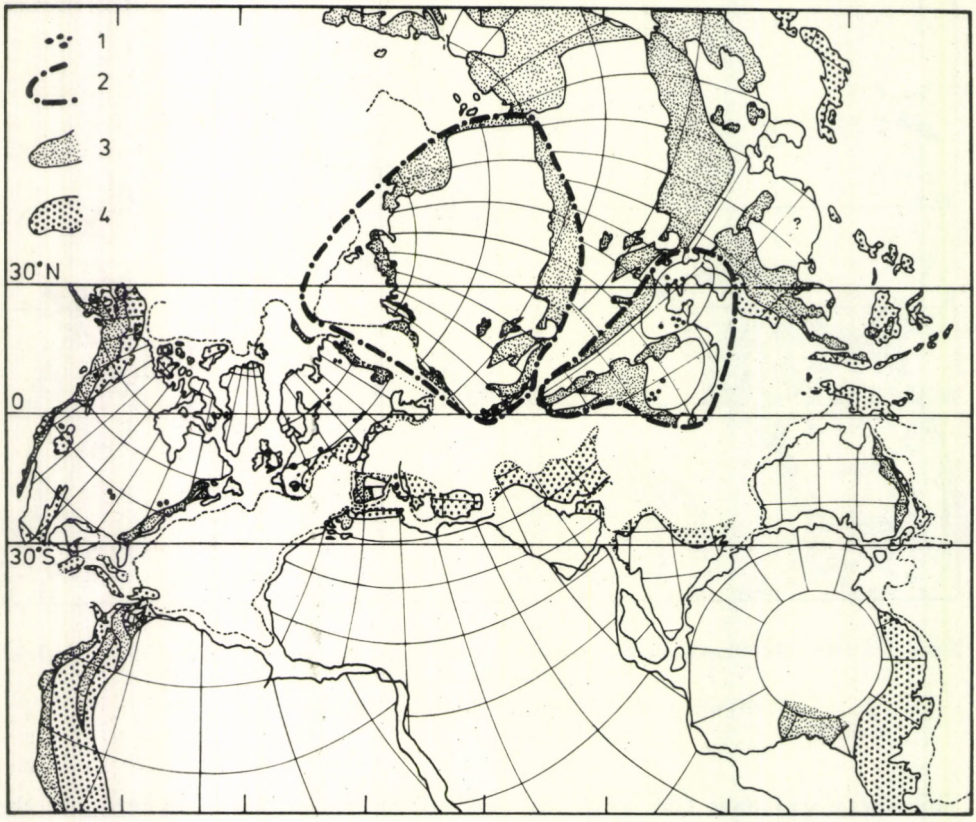


Figure 172. Lower Carboniferous continent reconstruction (about 340 m.y.). After Smith *et al* (1973) completed by bauxite deposits

(1) bauxite deposits; (2) positions of plates suggested by the present writer; (3) Hercynian orogenic belts; (4) Mesozoic and Tertiary orogenic belts

later, the farther south they lie. This is why there is a Hercynian orogen inserted between the North and South China platforms. The orogen between the South Chinese and the Indochinese platform was folded up in the Mesozoic; in fact, still farther south one finds mountain ranges folded up in the Neogene.

The consequential application of this mechanism of motion places all the bauxite regions of the Carboniferous into a suitable tropical climatic zone.

In the *Permian*, the Russian and Siberian continental plates coalesced, and the northward migration of the East Asian paraplatforms continued. The present writer has modified the corresponding continental reconstruction in this sense (fig. 173). In the Permian, bauxite formation ceased in North America, Europe and on the Siberian continental plate. It continued at reduced intensity on the North and South Chinese paraplatforms, and thence transgressed into Vietnam and Campuchea. According to the figure, the only bauxite deposits to lie slightly to the north of the tropical climatic zone, between the 30th and 40th parallels, are those of Northern China.

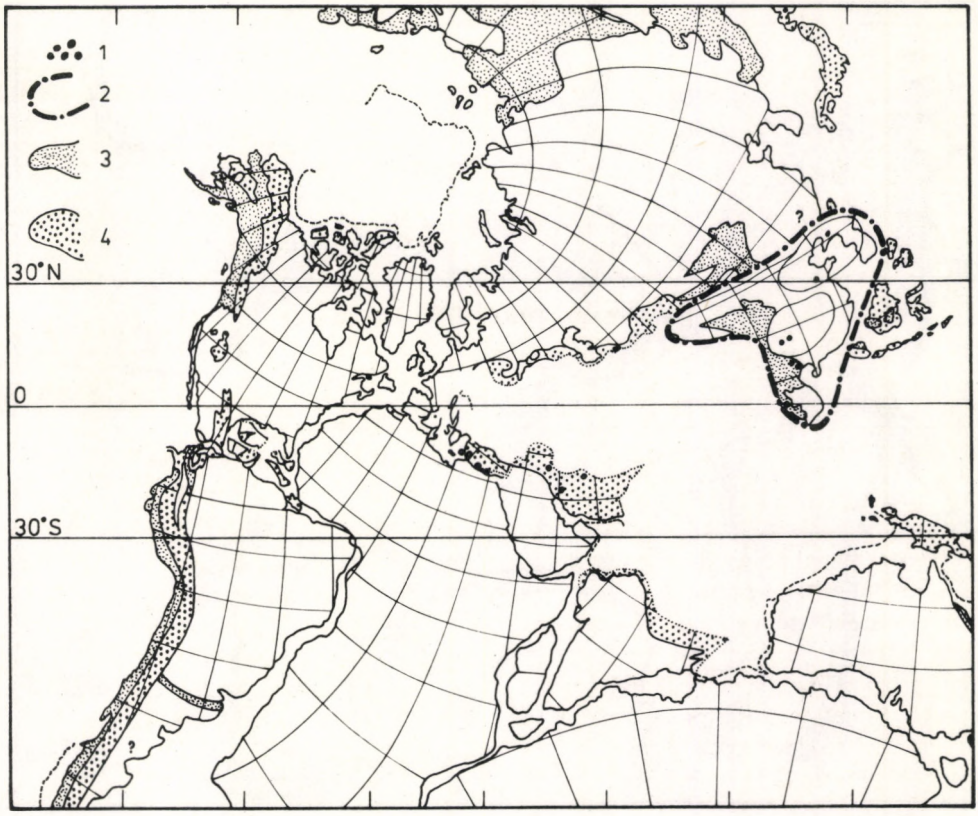


Figure 173. Middle Permian continent reconstruction (about 250 m.y.). After Smith *et al* (1973) completed by bauxite deposits

(1) bauxite deposits; (2) positions of plates suggested by the present writer; (3) Hercynian orogenic belts; (4) Mesosoic and Tertiary orogenic belts

In the Upper Permian, karst bauxites were formed in Asia Minor and in Central and Northern Iran. These lay on the southern shore of the Tethys, in the tropical climatic zone; the Alborz Range deposits, on the other hand, lay on the northern shore of the Tethys, presumably also in the tropical climatic zone.

In the *Triassic*, bauxite formation ceased in the East Asian region, as a result of a continued northward migration, as revealed clearly by figure 174. Karst bauxite formation, begun in the Permian on the southern and northern shores of the Tethys, also continued in the Triassic. It expanded to the Balkan Peninsula in the west; on the north shore, minor bauxite deposits are found in Soviet Central Asia. These latter lay somewhat to the north of 30° latitude north, but in a zone where bauxite formation is still climatically conceivable.

The only mystery is the Bogoslovsk–Volchansk deposit group in the Northern Urals, USSR. According to the reconstruction, this region lay between 50° and 55° latitude north in the Triassic; that is, far to the north of the tropical climatic zone. The resolution of this contradiction must remain a task for the future.

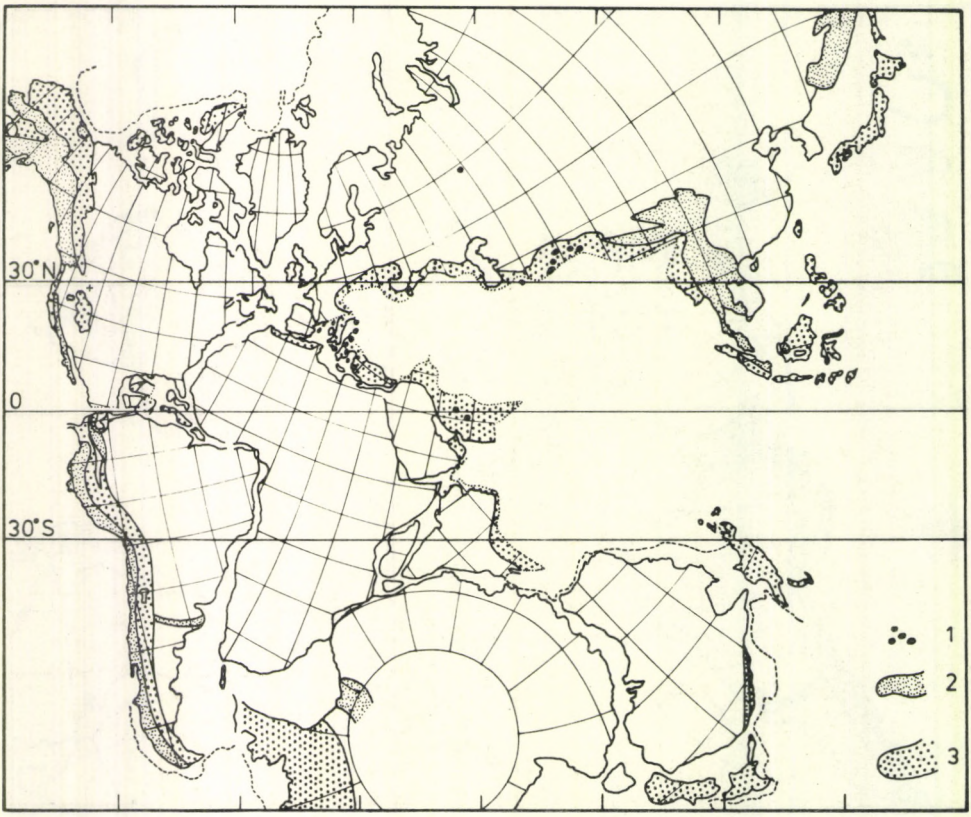


Figure 174. Lower Triassic continent reconstruction (about 220 m.y.). After Smith *et al* (1973) completed by bauxite deposits

(1) bauxite deposits; (2) Hercynian orogenic belts; (3) Mesozoic and Tertiary orogenic belts

In the *Jurassic*, bauxite formation continued on the north and south shores of the Tethys. Karst bauxites were formed on the Balkan Peninsula, and in minor deposits also in the West Alps, in the Northeastern Carpathians and, in the USSR, in the Crimea and in the Ferghana. On the southern shores of the Tethys, iron-rich laterites came to exist in the Taurus range: minor karst bauxite deposits formed in Pakistan, around Campbellpore and Muzaffarabad. As shown in figure 175, all these lay in the zone favourable to bauxite formation.

The areal extent of bauxite formation gradually increased in the *Cretaceous*. On both shores of the gradually shrinking Tethys, karst bauxite deposits came to exist, from Spain to Afghanistan. The centre of gravity of karst bauxite formation was on the Balkans and in the Transdanubian Hills, Hungary. Figure 176 reveals how all these regions fell into the tropical climatic zone. The same holds for the extensive bauxite deposits of the Southern Urals and Siberia, USSR (those along the Angara River, around Ubagansk, and around Verchne-Tobolsk). It was probably in the Upper Cretaceous that large-scale lateritic bauxite formation first began on the South American and West African continental plates.

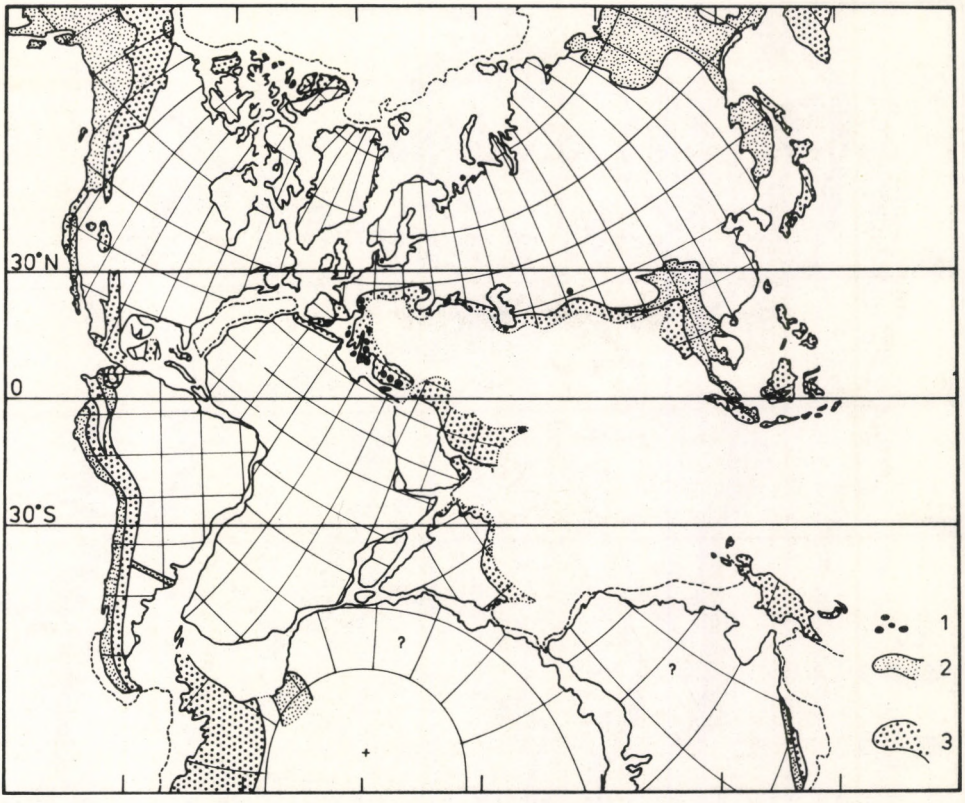


Figure 175. Middle Jurassic continent reconstruction (about 170 m.y.). After Smith *et al* (1973) completed by bauxite deposits

(1) bauxite deposits; (2) Hercynian orogenic belts; (3) Mesozoic and Tertiary orogenic belts

In the *Paleocene and Eocene*, bauxite formation continued in the Mediterranean region; its centre of gravity continued to be in Dalmatia and in the Transdanubian Hills of Hungary. Lateritic bauxite formation intensified on the African, South-American and Indian continental plates. Lateritic bauxites formed also in Northern and Eastern Australia and in Tasmania. Laterite and karst bauxite deposits came to exist in the Southeastern US.

Frakes and Kemp (1972) not only prepared continental reconstructions for the Eocene and Oligocene, but also constructed paleoclimatological maps based on calculated reconstructions of the most probable trajectories of the marine currents, on oxigen isotope determinations of paleotemperatures, and on paleobotanical data. According to them, continental configuration in the Eocene resulted in a highly favourable pattern of marine currents: as a result, the climate could grow much warmer than it is today (fig. 177). This made it possible for bauxite formation to continue in the Mediterranean region, in Dalmatia and in the Transdanubian Hills of Hungary above all. Lateritic bauxites were forming extensively in South America, West Africa, India, Australia and Tasmania. Lateritic and karst bauxite deposits kept forming in the

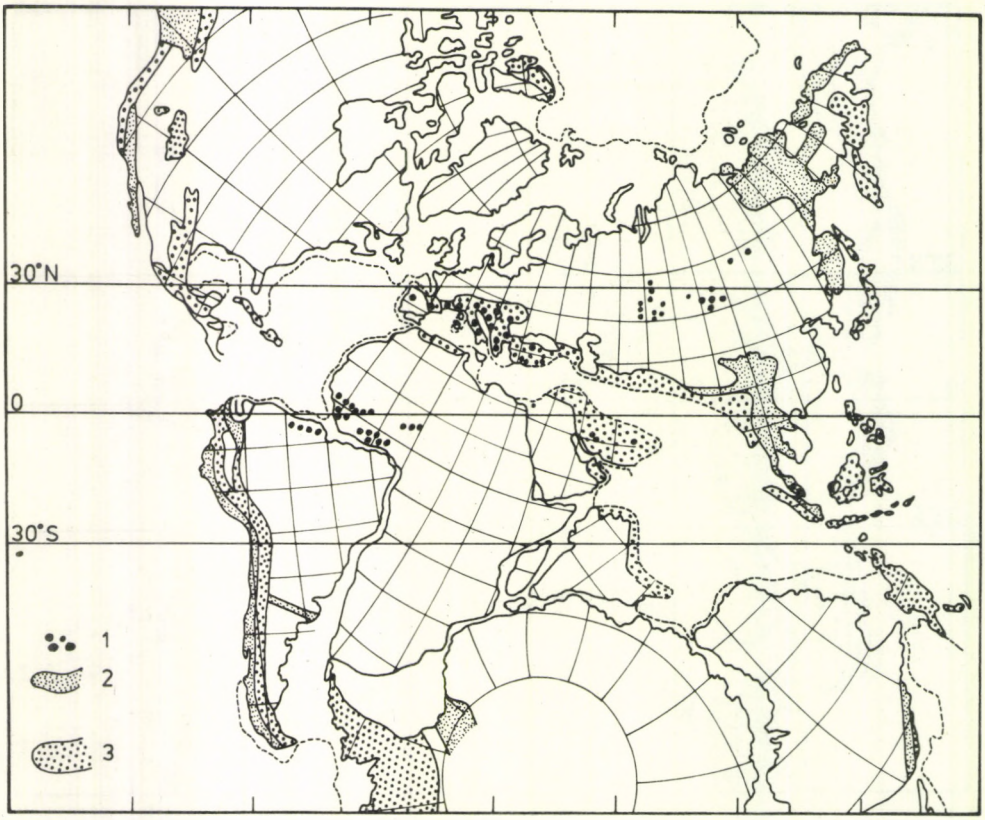


Figure 176. Middle Cretaceous continent reconstruction (about 100 m.y.). After Smith *et al* (1973) completed by bauxite deposits

(1) bauxite deposits; (2) Hercynian orogenic belts; (3) Mesozoic and Tertiary orogenic belts

Southeastern US. Under these favourable conditions, bauxite formation extended to 50 and even 55° latitude north.

The further drift of the continents in the *Oligocene* changed the pattern of marine currents, resulting in a cooler and drier global climate (fig. 178). This was the probable reason for the cessation of bauxite formation in the Mediterranean region and in the Southeastern US, as well as in Southeast Australia and Tasmania. Lateritic bauxite formation continued in South America, Africa, India and Eastern Australia. Far to the north of the optimum climate of bauxitisation, in Autrim, North-Ireland, minor lateritic bauxite deposits came to exist, probably aided by the Gulf Stream. This issue is unclarified as yet in its palaeoclimatological context.

In the *Neogene*, the formation of lateritic bauxites continued in the above-named regions. Furthermore, huge tonnages of lateritic bauxites came to be formed in the Caribbean. All these regions were within the climatic zone favourable to bauxitisation. A deposit hard to fit into this pattern, however, is the clayey lateritic bauxite of Miocene age on the Vogelsberg, Federal Republic of Germany; another one is the lateritic

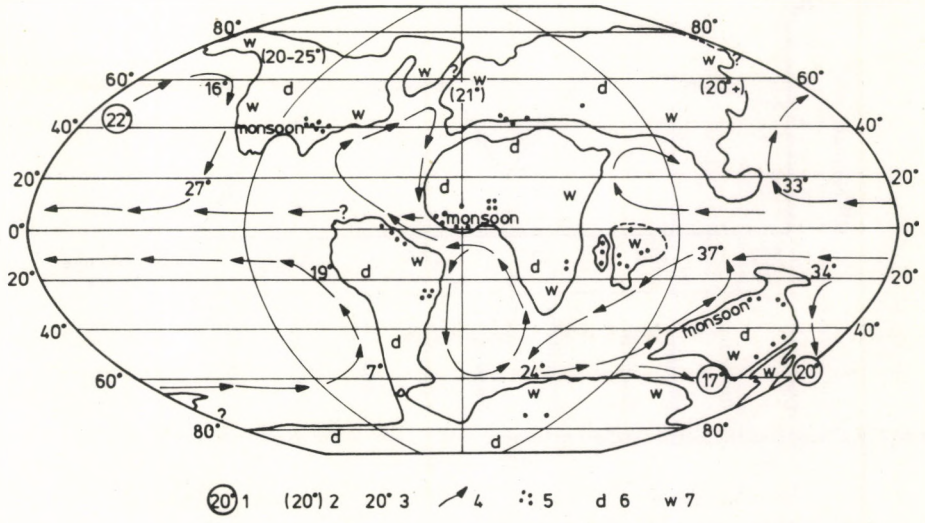


Figure 177. Eocene continent reconstruction after Frakes and Kemp (1972) completed by bauxite deposits

(1) paleotemperatures based on oxygen isotope measurements; (2) mean annual temperature (air); (3) calculated paleotemperature; (4) ocean currents; (5) bauxite deposits; (6) dry climate; (7) wet climate

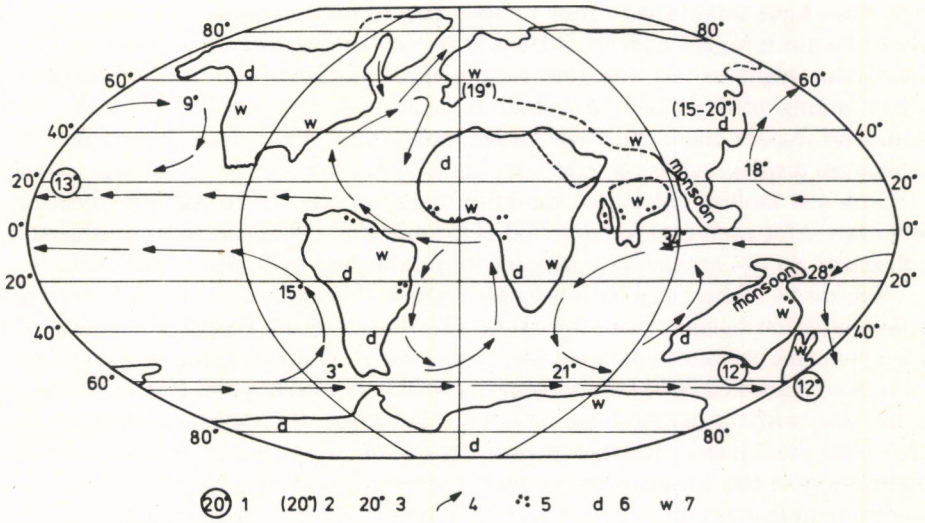


Figure 178. Middle Oligocene continent reconstruction after Frakes and Kemp (1972) completed by bauxite deposits

(1) paleotemperatures based on oxygen isotope measurements; (2) mean annual temperature (air); (3) calculated paleotemperature; (4) ocean currents; (5) bauxite deposits; (6) dry climate; (7) wet climate

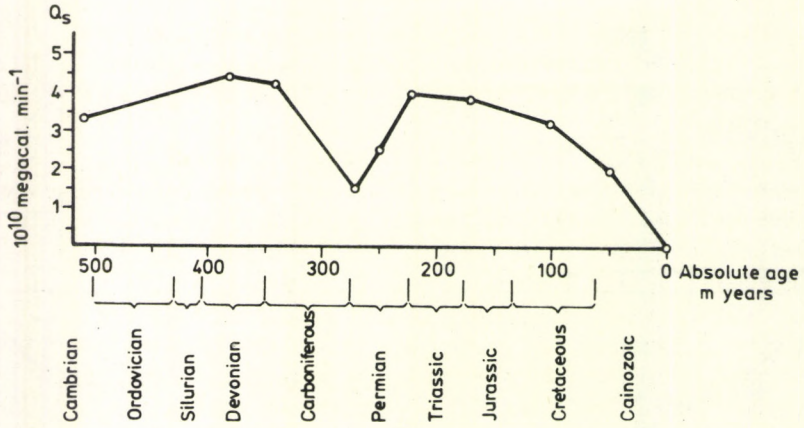


Figure 179. Changes of the Earth's radiation balance with respect to the recent continent/ocean distribution. After Dobosi

bauxite in Oregon State, US. Both regions lay far to the north of the zone of optimum climate.

Quaternary glaciation further confined the regions suitable for bauxitisation. Currently, the formation of lateritic bauxites is going on exclusively under the tropical humid climates near the equator, e.g. in Southeast Guinea, in Cameroun, in the Congo and Zaire, Brazil, Panama, Costa Rica, Hawaii and the Fiji Islands. No karst bauxite is being formed today except on some islands of the Western Pacific.

The plate tectonic and palaeoclimatological considerations just presented are usefully complemented by the satellite-borne meteorological observations of recent years. They have established ratios of absorbed to re-radiated solar energy, finding them to be much higher over the oceans than over the continents. A *global radiation balance* was calculated on this basis for the entire surface of the Earth. This finding implies among others that the radiation balance is the more positive, the more continental masses there are close to the poles, and vice versa. This implies that continental displacements of necessity also affect the global radiation balance. Meteorologist Dobosi calculated radiation balances referred to today's continental configuration for each geological period of the past (fig. 179), using for the purpose the reconstructions of continental configurations published by Smith *et al* (1973).

According to Dobosi's radiation-balance curve, there was a single great period of surplus radiation balance from the Middle Devonian to the Early Carboniferous; it gives a plausible explanation for intensified bauxitisation in this epoch. It was followed by a period of marked deficit at the boundary of Carboniferous and Permian, again in fair harmony with the bauxitological evidence. Subsequently, a substantial surplus had built up by the Triassic, which has then been gradually depleted by today. Intense bauxitisation in the Mesozoic is in good agreement with this hypothesis which is, however, unable to account on its own for the maximum of bauxitisation in the Upper Cretaceous and Eocene. It is the above-mentioned analysis by Frakes and Kemp (1972) that contributes evidence clarifying the situation. The drop in radiation surplus in the Neogene is rather striking. It was clearly one of the reasons resulting in the above-mentioned constriction of the zone suitable for bauxitisation.

In summary, the world bauxite formation pattern over the ages was controlled by the fluctuating extent of the zone of optimum climate, combined with the displacements of the continents, which incessantly modified the positions of the land areas relative to the optimum zone.

Even in the optimum zone, however, bauxite could not form or be preserved everywhere. This circumstance is ascribed by the present writer above all to the diverging structural evolution of the great geotectonic units. The concentration of lateritic bauxite formation on the continental platforms can be explained by the plausible assumption that lateritic weathering required, in addition to a favourable climate, also quiet periods of weak erosion. Under a favourable climate, lateritic weathering did, of course, take place also in the orogenic belts, but structural mobility made it much more likely for the weathering products to be eroded.

Karst bauxites are found mainly in orogenic belts, because the formation of pure carbonate rocks was confined to these belts. Furthermore karst bauxites had more of a chance of preservation than the laterite bauxites in their more exposed sites. The writer wishes to emphasise once again that, also in the orogenic belts, bauxite was a product of relatively quiet periods: no bauxites formed during and after the phases of orogeny. Nevertheless, only some 15% of total world bauxite tonnage is found in the orogenic belts, in confirmation of the above considerations, that is the optimum areas for bauxite formation and preservation being the quiet platform areas.

Studying the structural evolution of the orogens, the writer has gained the conviction that karst bauxites formed primarily where colliding continental plates caught some microcontinents or oceanic microplates between them. Cases in point are the range of the Uralids and Turkestanids in the Paleozoic, and the Mediterranean and the Caribbean regions during the Alpine orogeny. This succession of events with its islands, coastal peneplains, repeated upliftings and widespread carbonate rocks provided a favourable setting for karst bauxite formation.

Hardly any bauxites have been found so far in the Andes and in the Cordilleras, even where the climate was favourable. The likely reason is that the frontal overriding of an extensive oceanic plate by an extensive continental one creates a relief and a sedimentation different from the above-outlined ones. Considerable altitude differences are created: orogens such as these are characterised by the deposition, in repeated phases, of coarse clastic continental beds of molasse facies. The role of carbonate rocks is more subordinate; there are practically no quiet periods of emersion with low-intensity erosion (Aubouin 1972). Such conditions are unfavourable to the evolution of bauxite deposits.

Even in the Mediterranean zone of Alpine type orogeny which has been found to be so favourable geotectonically, bauxite did not form everywhere. Most of it concentrates in the southern arc system (fig. 166), the region which underwent the most complicated plate tectonical evolution. Torsions of microplates, orogens piling up around them, oceanic microplates largely consumed in the zones of subduction: this complex association created a varied relief in which repeated emergences offered the best chances for bauxitisation; erosion and the unfavourable influence of clastic sedimentation were relegated farthest into the background.

In those places, on the other hand, where a simple 'frontal' folding took place (e.g. in the Maghrebids), clastic sedimentation was more extensive, and the condition of quiet

sedimentation required by bauxitisation remained unfulfilled. This is also the reason for the absence of bauxites from the Himalayas, and for the scarcity of karst bauxites in Iran and Afghanistan. Erosion also was much intenser here: the majority of such deposits as did form fell prey to it.

Another fact not to forgotten is that, as a result of horizontal structural displacements, a number of bauxite deposits are not found today in their original sites. Horizontal displacements in the range from a few to a hundred kilometres may be reckoned with for the deposits in overthrust and nappe zones. A case in point is the nappe system of the East Alps, which includes the bauxite deposits of the Tirol, Salzburg, Unterlaussa and Dreistätten. According to Oberhauser (1968) and Ruttner (1970, oral communication), nappe formation began in the Campanian, immediately after the accumulation of the bauxites, and continued to the Aquitanian stage of the Miocene. The nappes travelled some 130 to 150 km in a northward direction. Overthrusts of similar size took place also elsewhere in the orogenic belts; these displacements must by all means be reckoned with in the palaeogeographic reconstruction.

8.6.2. INTERPRETATION OF BAUXITE FORMATION IN THE OROGENS

If the relationship between the geographic distribution of bauxite formation and mountain building is such a close one, there arises quite spontaneously the question whether there is also a temporal linkage between them. In order to explore the answers to this question, the writer has constructed diagrams of the evolution of a few geotectonic units on the basis of data provided by the relevant literature. He has added the stratigraphic positions of the bauxite horizons found in them and the constitution of the bauxitic formation as a whole. He has found bauxite formation and accumulation to coincide by and large with certain well-defined phases of structural evolution in the main bauxite regions. There are some minor differences only according to whether the bauxite region in question is situated in a pericontinental or an island-arc orogen.

The relationship is most compelling in the *geosynclinal zones* of the continental margins. For an example, let us survey the regions of the Dinarids, the Hellenids and the Appennines. All these ranges belong to the southern arc system of the Alpine–Mediterranean geosynclinal zone (fig. 166). The Dinarids and Hellenids are contiguous; it is the structural line of Skutari-Peć that is usually regarded as the boundary between them. In both ranges, the southwestern zone is miogeosynclinal, the northeastern is eugeosynclinal. Parallel to their strike, there are elongate swells, alternating with, deeper-water zones ('furrows'). Bauxite formation was confined to the swells. From the early Mesozoic on, sedimentation on the swells, where bauxite is found, is characterised by pure carbonate formations.

In the Dinarids, the Pfalzian orogenic phase (Lower Triassic), interpreted as the final event of the Hercynian orogeny, was followed by fairly weak deformations of the so-called neo-Labinian phase, local emersions above all, at the boundary of the Middle and Upper Triassic (Herak 1971). In the emerged areas, in Slovenia, Lika and Crna Gora (Montenegro), Yugoslavia, sizeable bauxite deposits came to exist. The Late

Kimmerian tectonic phase also was a weak one; it resulted in local emergences similar to the above ones. It is with this phase that the bauxite deposits accumulated in the Malm and at the boundary of the Jurassic and Cretaceous are associated. The Austrian, Subhercynian and Laramian phases also gave rise to similar weak movements and local emergences, each of which gave rise to the accumulation of a bauxite horizon. Herak interpreted these phases as precursors to the geotectonic 'cataclysm' at the end of the Paleogene.

It was at the beginning of the Miocene, in the Savian phase, that the overthrusting of the structural zones towards the southeast took place, and the great nappes came to exist. These large-scale structural movements greatly increased the relief energy and hence also the rates of erosion in the region. Carbonate sedimentation was replaced by clastic sedimentation in ever-expanding areas. As a result, the setting suitable for bauxite formation disappeared. Even if some bauxite had formed locally, it was soon eroded and mixed into the huge masses of clastic sediment. This tendency was but amplified by further orogenic phases, following in the Neogene. This is why no bauxite deposits younger than the Upper Oligocene are known from the Dinarid region.

Structurally, then, the Dinarids are characterised by an absence of absolute structural quietness at all times after the Hercynian orogeny. The repetition of the first, comparatively weak orogenic phases gave rise to local emergences in the zones of the swells. This is why there are so many bauxite horizons in this region. The tonnage distribution of bauxite and bauxitic clays over the geological periods is as follows:

Oligocene	11 million tons
Eocene	78 million tons
Paleocene	66 million tons
Cretaceous	123 million tons
Jurassic	134 million tons
Middle and Upper Triassic	187 million tons

That is, the rate of accumulation of bauxite decreased as time passed: on the other hand, it is the bauxite of the Triassic horizon that is the most clayey.

It is a familiar fact that mountain building within a geosynclinal zone starts in the interior of the eugeosynclinal zone, to proceed outward by degrees through the miogeosynclinal zone, towards the margin of the orogen. This relationship is valid by and large also for the accumulation of the karst bauxite deposits. For example, in the eugeosynclinal zone of the Dinarids, no bauxite deposits younger than Upper Cretaceous are known. On the inward side of the miogeosynclinal zone, in the so-called 'Dinaric swell zone', bauxite horizons extend from the Middle Triassic to the Middle Eocene. In the outward, so-called 'Adriatic swell zone', on the other hand, only Palaeocene and Eocene deposits are found (except for the small Tithonian deposit group of Vrsar in Istria). It is not accidental that the bauxitic formation of the Vlasenica deposit group in the innermost, so-called 'Lim zone' is the one to deviate farthest from the pure carbonate type, exhibiting as it does the greatest similarity to 'contaminated' carbonatic bauxite formation of the Transdanubian Hills, Hungary.

Structural evolution took a similar course in the *Hellenids*. Here, it is in the innermost part of the eugeosynclinal zone, in the so-called 'Pelagonian zone', that the

oldest (Permo–Triassic, Triassic and Jurassic) bauxite deposits are found. In the next zone outward (to the west), called the Subpelagonian Zone, bauxite is of Lower Malm age. Bauxite is most abundant in the next, so-called 'Parnassos zone', which has three bauxite horizons: Lower Malm, Upper Malm and Turonian. Each bauxite horizon is associated with gentle structural movements and local emergence (Aubouin 1969). Flysch, accumulated in great thickness in the Paleocene and Eocene, is the result of a more substantial structural uplift and an associated erosion, which put an end to bauxite accumulation in this particular structural zone. In the outermost, so-called 'Gavrovo zone', already of a miogeosynclinal character, reef-type carbonate sedimentation went on from the Triassic to the Middle Eocene. In this swell zone, a local emergence and bauxite accumulation took place in the Middle Eocene, after some weak structural movements. Here too, bauxite accumulation was stopped by the deposition of flysch.

The succession from older to younger bauxite deposits from the inside outward can thus be demonstrated also in the Hellenids. The only bauxite deposits not fitting into this picture are the minor, Senonian ones on Euboea Island in the Aegean, whose palaeogeographic and structural interpretation will require further studies. In the Hellenids, the paroxysm of structural evolution was in the Oligocene and early Miocene, when the structural zones were overthrust in a southwesterly direction (Aubouin 1969). The bauxite deposits, then, came to be formed before the paroxysm, in the precursory phases, also in the Hellenids.

In the *Appennine* geosyncline also, there were zones of basins and swells parallel to the strike of the geosynclinal axis. As distinct from the Dinarids and Hellenids, there was here a period of structural quiet lasting from the end of the Paleozoic right up to the Albain, with a continuous neritic and deeper-water marine sedimentation. No bauxite could form anywhere during this period. A slight mobilisation entailing an emergence took place in the Cenomanian and in the Turonian. During it, bauxite deposits were formed in one or two horizons in the central and external swell zones. A stronger uplift and erosion initiated a flysch-like sedimentation in the Paleogene, which precluded any further opportunities of bauxite formation. The paroxysm of structural evolution took place early in the Miocene (Savian phase), at which time the swell zones were overthrust in an easterly direction onto the basin zones intercalated between them (Ippolito *et al* 1973). Compression was three- to fivefold in the southwest –northeasterly direction.

In summary, bauxite deposits are confined in each of these three ranges to the precursory phases of orogeny, characterised by a slight structural mobility. The largest tonnages of bauxite and the largest number of bauxite horizons were formed when this period was longest (in the Dinarids: 600 million tons). In the Hellenids, where the precursory phase was somewhat shorter, the number of bauxite horizons is less, and so is the aggregate bauxite tonnage (500 millions tons). In the Appennines, where the precursory phase was the shortest, having been preceded by a long period of rest, there is just one bauxite horizon (at most two in places), and aggregate bauxite tonnage is a mere 50 million tons. If the bauxite tonnages formed in these periods but eroded subsequently could be added, the differences would be even more striking.

In the Hercynian geosyncline of the Urals, bauxite deposits accumulated in the Middle Devonian in the eugeosynclinal zone, and in the Upper Devonian in the

miogeosynclinal zone. The progress from older to younger bauxites from the interior outward is thus also confirmed here.

Of the *intracontinental orogens*, the structural evolution of the Pyreneo–Provençal range is known in greatest detail. We shall take this range as our example, in view of the fact that it contains sizeable bauxite deposits. In the region, several phases of the Hercynian orogeny have been demonstrated, the last spasm of which is reflected by a hiatus at the boundary of Permian and Triassic. Subsequently, a continuous marine sedimentation prevailed over most of the region right up to the Aptian. Most of the bauxite was accumulated in the Aptian, Albian, Turonian and Senonian stages, and on the boundary between Cretaceous and Palaeocene (colour figure 23).

According to the structural analyses set forth by Aubouin and Menessier (1960–62) and Mattauer and Proust (1967), the region was the site of a gradual subsidence and structural quiescence in the Triassic and in the Jurassic. Early in the Cretaceous, structural mobility took an upswing, and subsidence was interrupted by local emergences. The germs of the subsequent structural zones, the axes of anticlines and synclines developed at the time. This is when bauxite formation first started. Its intensification was associated with the emergence of the so-called Durancian Isthmus. Increasing mobility is indicated in the Upper Cretaceous by further subsidences and emergences.

At the boundary of the Cretaceous and Palaeogene, the Laramian phase brought folding and dynamic metamorphism to the axial zone of the Pyrenees. As a result of relief sculpture by these movements, carbonate sedimentation was succeeded by the accumulation of clastic sediments all over the range, which at the same time put a stop to bauxite accumulation. The strongest phase of orogeny in this region was the Pyrenean, at the boundary of Eocene and Oligocene. It resulted in strong folding, imbrications and overthrusts. Subsequently, most of the region was uplifted, and sedimentation was replaced by widespread erosion. There were phases of orogeny, weakening by degrees, even in the early and late Miocene, at a time, however, when no bauxite came to be formed anywhere.

In terms of absolute ages, the structural rest succeeding the Hercynian orogeny lasted for some 90 million years here. During it, no bauxite was formed. Subsequently, there were 70 million years of gentle structural mobility, with local emergences. All the bauxite deposits were formed in this period. It was followed by a period characterised by strong phases of orogeny, strong erosion and clastic sedimentation, and no bauxite. This temporal succession agrees very well with our findings concerning the geosynclinal zones.

As an example for the *median masses and basins*, let us review the geological evolution of Hungary. In this region, the Hercynian orogeny gave rise to a zoned structural arrangement of southeast–northwesterly strike. The paroxysm of the orogeny was at the boundary of the Devonian and Carboniferous, in the so-called Bretonian phase, but the Sudetian phase was also a strong one. The end of the orogeny is marked by the Asturian phase, a weaker one already (Wein 1973). According to Wein, this is when the consolidation of the Hercynian structures gave rise to 'a crustal entity made up of rigid Precambrian and early Paleozoic rocks which subsequently became the nucleus of the structural unit known by the name of Tisia or Pannonian median mass.' The erosion of the Hercynian ranges started in the Upper Carboniferous: it was strongest in the

Permian, and gradually died down in the Triassic. In each of the structural zones developed during the Carboniferous, sedimentation of a different type began in the Mesozoic.

As regards bauxite formation, the zone of the greatest importance is the one occupied by the Transdanubian Hills. There, subsidence and repeated weak oscillations took place in the Triassic, Liassic and Dogger. In the Malm, the Late Kimmerian phase, which resulted in faulting and gentle doming, was of a slight intensity overall. Carbonate sedimentation continued practically uninterrupted afterwards. The Austrian phase was somewhat stronger, causing faultings, domings and local emergence. With this phase is associated the Lower Aptian bauxite horizon of Alsópere. No significant relief sculpture took place even then, and, in fact, the carbonate sedimentation could go on unchanged.

Subsequent to the Subhercynian orogenic phase in the Turonian, the entire zone was uplifted; this resulted in sizeable bauxite accumulations in the Southwestern Bakony Hills. Subsequent to the Gosau transgression, the Laramian orogenic phase brought a further emergence. It is as a result of this phase that the geographically most extensive bauxite accumulation took place, from Sümeg in the Southwest up to Nézsa in the northeast. Only a few insignificant accumulations of bauxitic clay are associated with a local emergence owing to the Illyrian orogenic phase in the Middle Eocene. Strong compression and imbrications in the Pyrenean orogenic phase at the end of the Eocene sculptured an entirely new relief, the more uplifted parts of which were exposed to intense erosion. As a result, prior carbonate sedimentation was superseded by sandy–clayey sedimentation, which at the same time meant the end of bauxite accumulation.

In the Eastern Alps and in the Carpathians, the paroxysm of mountain building coincided with the Savian and Styrian orogenic phases, bringing huge overthrusts and nappes and a general uplifting. In the Pannonian Basin, the forces acting after these phases were largely tensional, with large fault blocks starting to subside along the faults.

Stegena *et al* (1975) regarded the Pannonian Basin region as an interarc basin, the Paleo–Mesozoic nucleus of which was the product of the collision of a Gondwanian and a Laurasian microplate. The folding up of the Carpathians and of the Dinarids was the result of subduction towards the interior of the basin. The plate undergoing subduction was partially melted by the heat of friction, and the volatiles of Earth-surface origin started to migrate upward: being held up by the bottom of the crust, they partially re-melted it. This process, called 'subcrustal erosion', reduced by some 8 km the thickness of the lower crust according to those authors. The subsidence of the crust thus thinned was an isostatic process, promoted by the load of the superincumbent Neogene and Pleistocene strata.

Also in this geotectonic unit, bauxite formation took place in a phase of increased mobility succeeding a rather quiet period, and preceding the great turnabout in structural evolution, when the compressive stresses changed over into tensional ones. Since the nucleus of the region is a microplate partly consolidated in the course of the Hercynian orogeny, it reacted more weakly to the orogenic phases than the surrounding geosynclinal zones. The sluggishness of the oscillations is indicated by relatively longer hiatuses of sedimentation..

The structural zone of Villány–Tótkomlós–Bihar in the southernmost part of the country is a transition towards the geosynclinal zones. It is characterised by greater

mobility and strong compressive movements in the Austrian and Subhercynian orogenic phases. It is with a preceding, much weaker phase, the Late Kimmerian, that the accumulation of the bauxite deposits of Nagyharsány and Pădurea Craiului are associated. The stratigraphic successions of the two sites are strikingly similar, and so are the mineralogical and chemical features of their bauxites. No bauxites were formed, on the other hand, in the Mecsek–Kiskőrös and Igal–Bükk structural zones which, although adjacent, have a completely different evolution history.

In summary: in the pericontinental orogens and in the intracontinental ones, bauxite accumulated in periods of weaker structural mobility preceding the paroxysm of mountain building. There are no differences in this respect between the individual second-order geotectonic units.

The structural evolution of the *island-arc orogens is* different, and this difference had a bearing on bauxite formation. Let us consider the Caribbean region for an example. Structurally, Jamaica and Hispaniola may be regarded as the east end of the so-called Nicaraguan Rise (Arden 1969). A state of relative structural rest prevailed in this region up to the middle of the Jurassic; the crust here presumably was a thin, oceanic one. At that time, the Atlantic lithospheric plate of oceanic character started to underthrust the Caribbean lithospheric plate. A strong volcanism arose along the zone of subduction, and the erupted volcanics constituted an island arc (Horsfield and Roobol 1974). At the end of the Cretaceous, the first phase of the Laramian orogeny resulted in a strong uplifting of the island arc. Simultaneously, synorogenic granodiorite masses were intruded, and high-pressure low-temperature metamorphism also took place along the axis of the arc. An orogenic phase in the early Eocene resulted in strong folding, up- and overthrusting.

From the Middle Eocene to the middle of the Miocene, for 30 million years, there was a period of relative structural quiescence in this region. A gradual slow subsidence and a carbonate sedimentation keeping step with it resulted in the formation of a thick limestone succession. Weak orogenic movements in the middle of the Miocene caused Jamaica to emerge in its entirety and to fall apart into fault blocks in the process. Most of the blocks were tilted southward. At the end of the Pliocene, further strong structural movements took place in Eastern and Northern Jamaica, with substantial folding and doming. The principal fault lines were also rejuvenated.

It is a fact of prime importance for structural evolution that the deep-drilling and seismic evidence precludes the assumption of an early Mesozoic or Paleozoic geosyncline in this region (Arden 1969).

Bauxite formation probably started in the Upper Oligocene in Northern and in the Miocene in Southern Jamaica, going on in both places presumably up to the end of the Pliocene. Also on Hispaniola, bauxite formation presumably took up the entire time span from the Upper Miocene to the Pliocene.

The Caribbean Island Arc, and its northern wing which comprises Hispaniola and Jamaica, are still structurally active regions today, but the subduction of an oceanic lithospheric plate has been succeeded by the lateral displacement of neighbouring lithospheric plates along the Cayman deep-sea Trench. This is implied by frequent earthquakes in our days (e.g. Guatemala 1976). Further orogenic movements are likely also in future at the boundary of the two lithospheric plates.

Bauxite formation started on the freshly emerged surfaces at the end of a period of relative rest succeeding the Laramian orogeny, and was presumably stopped by the orogenic movements of the Pliocene. Bauxite formation currently going on the islands is insignificant. In the writer's opinion, bauxite formation has not yet come to a close in this region: further phases of orogeny and climatic changes may lead to the formation of more bauxite, and if there will be temporary inundations by the sea, several bauxite horizons may come to exist, just as in the Mediterranean region.

The island arcs in the Western Pacific represent even younger stages of mountain building: the underthrusting by the oceanic lithospheric plate is still going on full tilt. Accordingly, volcanic activity and pyroclastic sedimentation are predominant. Karst bauxites could only form in the relatively quiet parts of the region, where reef limestones emerged above the sea level (e.g. in the Solomons, in the Loyalty Group, etc). Wherever it was permitted by a suitable climate and weak enough erosion, even some minor laterite bauxite deposits could develop (Guam, Fiji etc). Bauxite formation is continuing in these regions even today, and will presumably also continue in the geohistorical future, possibly producing several bauxite horizons.

Bauxite formation in similar island-arc regions presumably took place also in the Mesozoic and Paleozoic, in the region of the Tethys, which has disappeared since by crustal-consumption.

In the orogenic belts, bauxite formation could restart in the so-called *post-orogenic periods*, subsequent to the structural consolidation of the region, provided the region remained within the climatic zone favourable to bauxitisation. This second set of bauxite deposits appears at the base of the post-orogenic succession, which is of an epeirogenic nature: it includes bauxites of laterite and Tikhvin type on alumosilicate rocks and of karst type on carbonate rocks.

The geotectonic evolution of the Earth, then, controlled the formation of karst bauxites not only in space but also in time. This is essentially the root cause for the tonnage of karst bauxites formed in the course of geohistory varying cyclically rather than linearly (table 1). The strong upswing in karst bauxite quantity in the Devonian is due to bauxite formed partly in the Hercynian orogens, and partly on the paraplatforms of China. The next maximum in the Upper Cretaceous corresponds to the weakish precursory movements of the Alpine orogeny in the Mediterranean region. The Miocene–Pliocene maximum of karst bauxite formation can also be interpreted as an analogous phase, occurring in the Caribbean region.

8.6.3. GENERAL GEOHISTORICAL PROBLEMS OF BAUXITE FORMATION

In the two foregoing sections, we were concerned with the spatial and temporal relationships between bauxite formation and the tectonic evolution of the Earth. The evolution of bauxite formation, however, was clearly influenced by numerous other factors. The questions as to which where these factors and whether they had acted in the geohistorical past in the same way as they are acting now is an unsolved question as yet. The opinions of the interested researchers diverge widely as to how far the *principle of uniformitarianism* (actualism) can be applied to the older bauxites. Finally, opinions

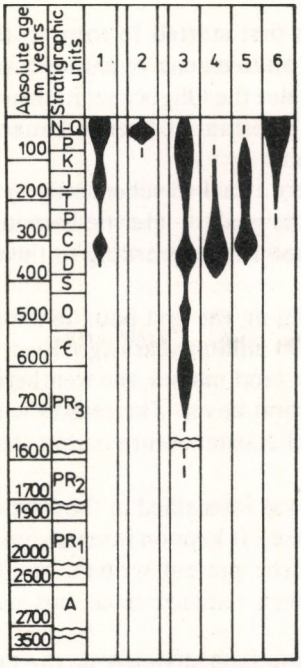


Figure 180. Evolution of bauxites and laterites. After Mikhailov (1977)

(1) lateritic iron ores; (2) lateritic manganese ores; (3) karst bauxites; (4) reworked lateritic bauxites ('sublateritic bauxites'); (5) Tikhvin-type or sedimentary bauxites; (6) lateritic bauxites

diverge greatly also as to the overall evolution of bauxite formation. Let us cite here only the most noteworthy opinions.

According to Strakhov (1960, 1963), bauxite formation on the globe first started as long as 4·5 billion years ago. All over the Archaeozoic and Proterozoic, bauxite deposits were formed in the seas by chemogenic processes, similarly to the immense deposits of Precambrian iron ore. Strakhov was led to this assumption by geochemical considerations based on some similarities in the behaviour of Al, Fe and Mn. He attributed the fact that the huge bauxite deposits postulated by him had not yet been discovered to the inadequacy of prospection in Precambrian regions.

This chemogenic bauxite formation in the sea was superceded according to Strakhov by the accumulation of clastic bauxites dumped into the seas about one billion years ago. In the land areas, the formation of karst bauxites has been going on since the beginning of the Paleozoic, that of lateritic bauxites since the beginning of the Mesozoic, both at gradually accelerating rates. These, however, are insignificant with respect to tonnage as compared with the ancient, marine bauxite deposits.

Mikhailov (1975, 1977) refuted Strakhov's theory on the grounds that aluminium delivered to the sea necessarily combines with the silica available there, forming clay minerals. Geological prospection in recent years has found no sizable bauxite deposits in any of the Precambrian successions, in further refutation of Strakhov's theory. According to Mikhailov, bauxite formation first started some 800 million years ago; both its intensity and its variety of forms has increased since (fig. 180). According to

Mikhailov, lateritic bauxites first started to form in the Devonian, in thin elongate strips, and from the Triassic on in extensive uninterrupted areas. The peak of bauxite formation on the globe was since the Oligocene; its centre of gravity kept shifting in the favour of lateritic bauxites. Substantial laterite bauxite formation is going on even today in the equatorial zones.

Mikhailov's statements were attacked vehemently by Akayemov and his co-workers (1975), according to whom the peak of lateritic bauxite formation was in the Eocene. The process has greatly slowed since according to them, as a result of global climatic changes.

In the present writer's opinion, the first bauxite deposits came to be formed in the early Precambrian about 3000 million years ago. Even if lateritic weathering could in fact take place on the ancient land masses, the weathering products were mostly soon eroded in the absence of a plant cover. The reason why the bulk of Early Paleozoic bauxites is stratified and is of clastic texture is that they had been swept together by local streams.

Continental lateritisation was intensified in the Carboniferous by the appearance of a plant cover on the land masses. It kept on intensifying also subsequently, with a peak in the Paleocene–Eocene in the present writer's opinion. The geographic extent of lateritisation has certainly been reduced since, but not to the extent postulated by Akayemov *et al* (1975).

Karst bauxite formation kept intensifying from the Devonian up to the Quaternary, apart from the cyclical changes referred to above. This holds even if the erosion of larger tonnages of older (Mesozoic and Paleozoic) bauxites is admitted. Although continental drift and mountain building in the orogens did control the events of bauxite formation in time and space, it exhibits no feature which would account for this steady increase. The present writer feels that it was the gradually increasing spread of carbonate rocks that gave the opportunity for increasing quantities of karst bauxites to form and to accumulate. It was especially the formation of biogenic limestones that has increased since the end of the Precambrian (Strakhov 1960, 1963).

Another fundamental problem is whether the karst bauxite formation in the past was different from karst bauxite formation today. On the strength of the evidence presented in chapters on Depositology and Petrography, the writer feels that, the older a deposit, the less can the principle of uniformitarianism be applied to it, even though the minerogenetic processes of geochemistry have changed but little since the formation of the oldest bauxite deposits. As regards the mode and facies of accumulation of bauxite, profound changes have in fact taken place. The greatest difference is the tendency of the Paleozoic deposits to accumulate in neritic and lagoon facies, the fair stratification of the deposits proper, and their frequent grading into carbonatic cover strata. These different conditions of sedimentation were presumably controlled by a plant cover and, in part-connexion therewith, by the relief conditions, different from those prevailing today. Bauxitisation could be influenced in the early Paleozoic by an atmosphere whose composition also differed from today's.

Bauxite formation, as we have seen, is an exceptional and rare event in the evolution of Earth, a process that has produced a mere thousandth of a per cent of all sedimentary rocks of the globe. A favourable conjunction of numerous factors was required for bauxite to be produced, and even more numerous factors must have combined to

permit the accumulation of karst bauxite deposits and their preservation to this day. Successful bauxite prospection requires an increasingly profound understanding of these factors. It is this understanding, combined with the quantitative analysis of the material of bauxite, by means more sophisticated than today's, that may permit us to solve those problems of bauxite formation that are still controversial today. It is to this endeavour that the writer wishes to contribute this book.

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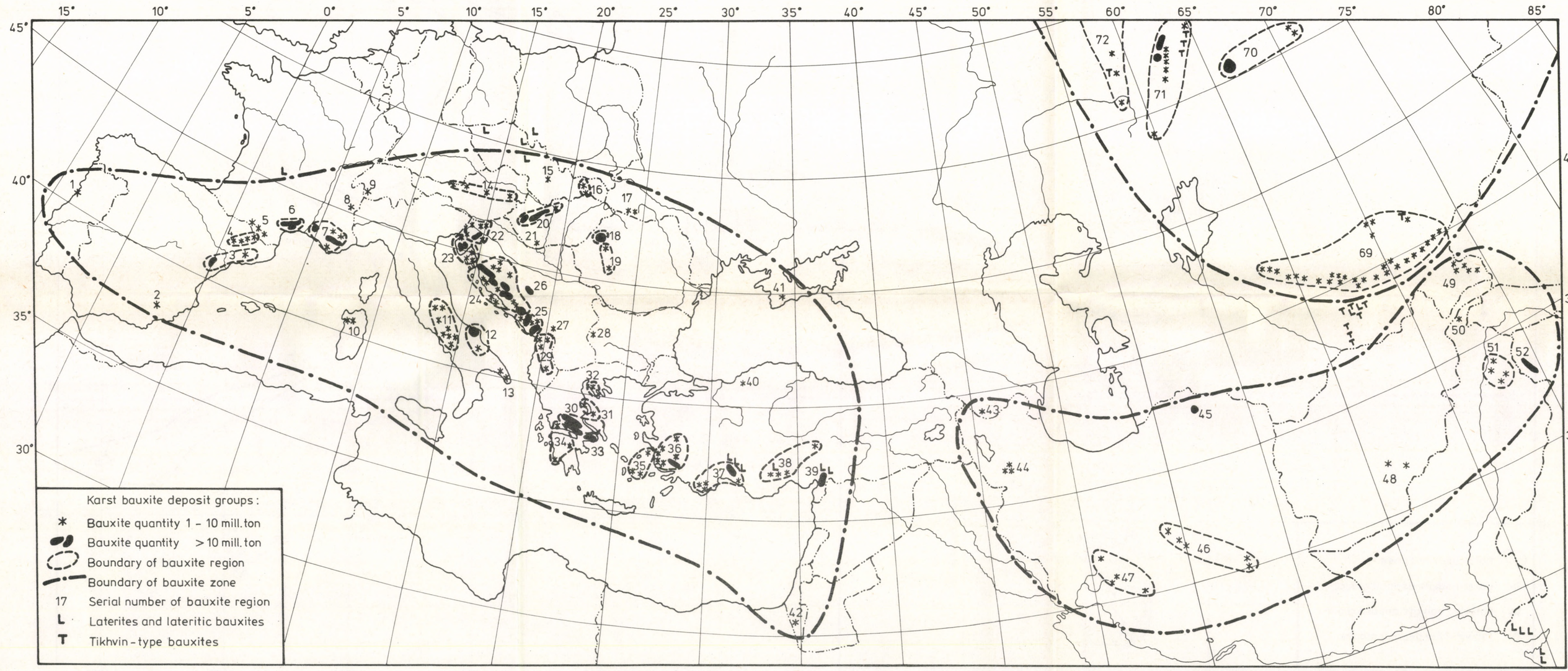


Figure 1. Karst bauxite deposit groups of the Mediterranean and Irano–Himalayan bauxite belt

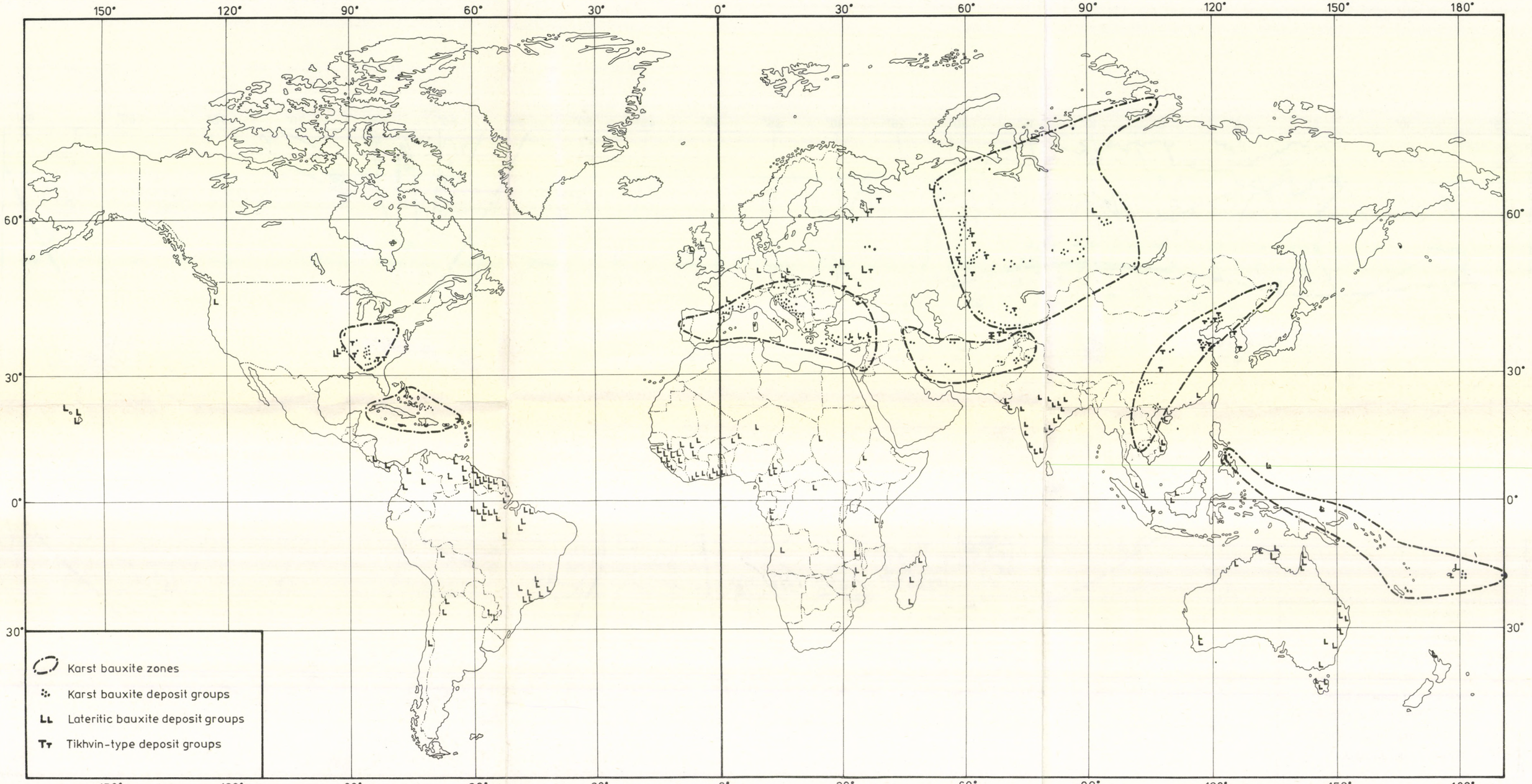


Figure 5. Bauxite deposit groups of the world

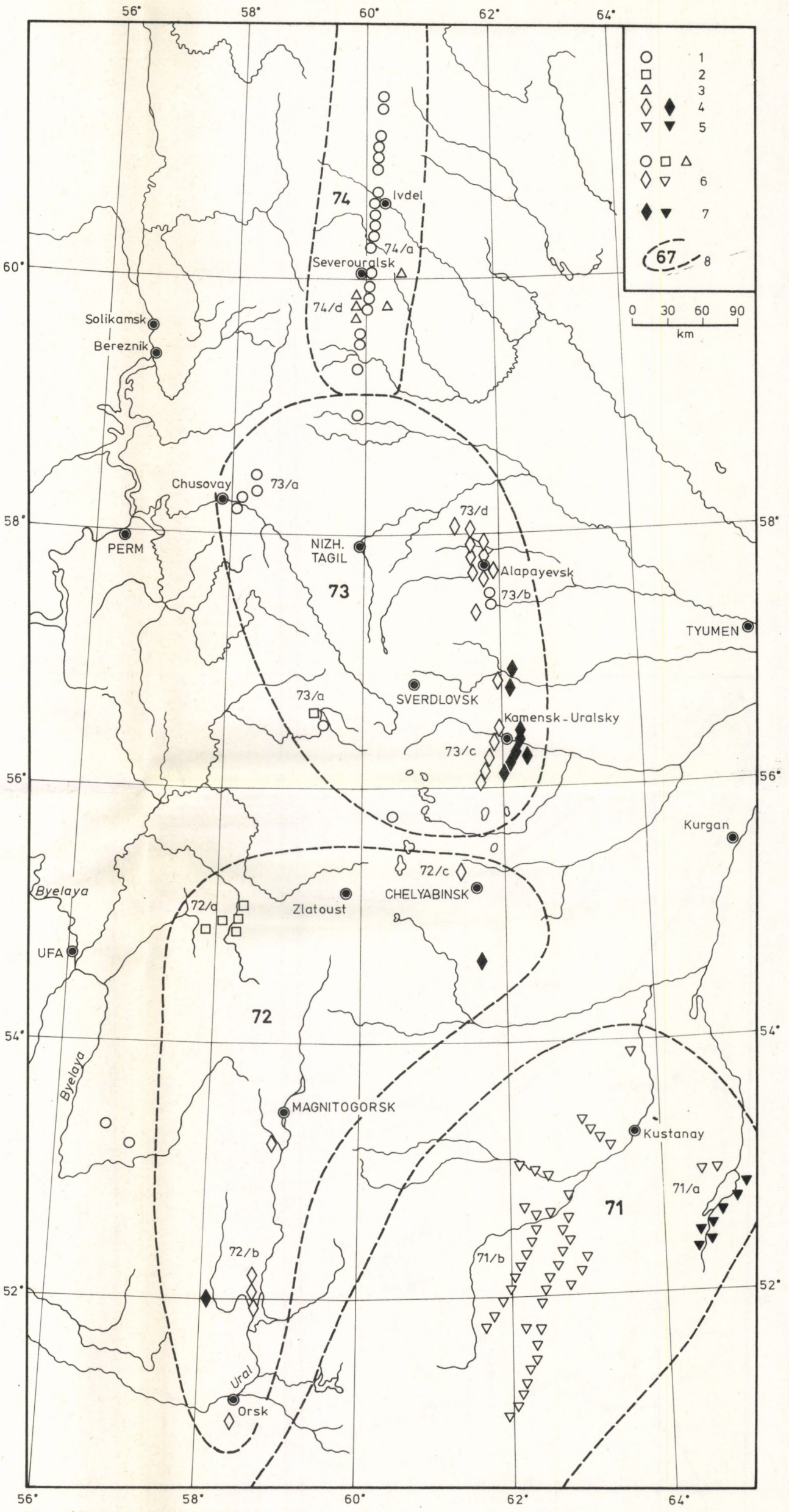


Figure 24. Bauxite geological map sketch of the bauxite regions of the Urals, USSR

(1) Middle Devonian bauxites; (2) Upper Devonian bauxites; (3) Upper Triassic bauxites; (4) Apt–Albian bauxites; (5) Cenomanian–Turonian bauxites; (6) karst bauxite deposits; (7) Tikhvin-type and lateritic bauxite deposits; (8) serial number and areal extent of bauxite deposit group

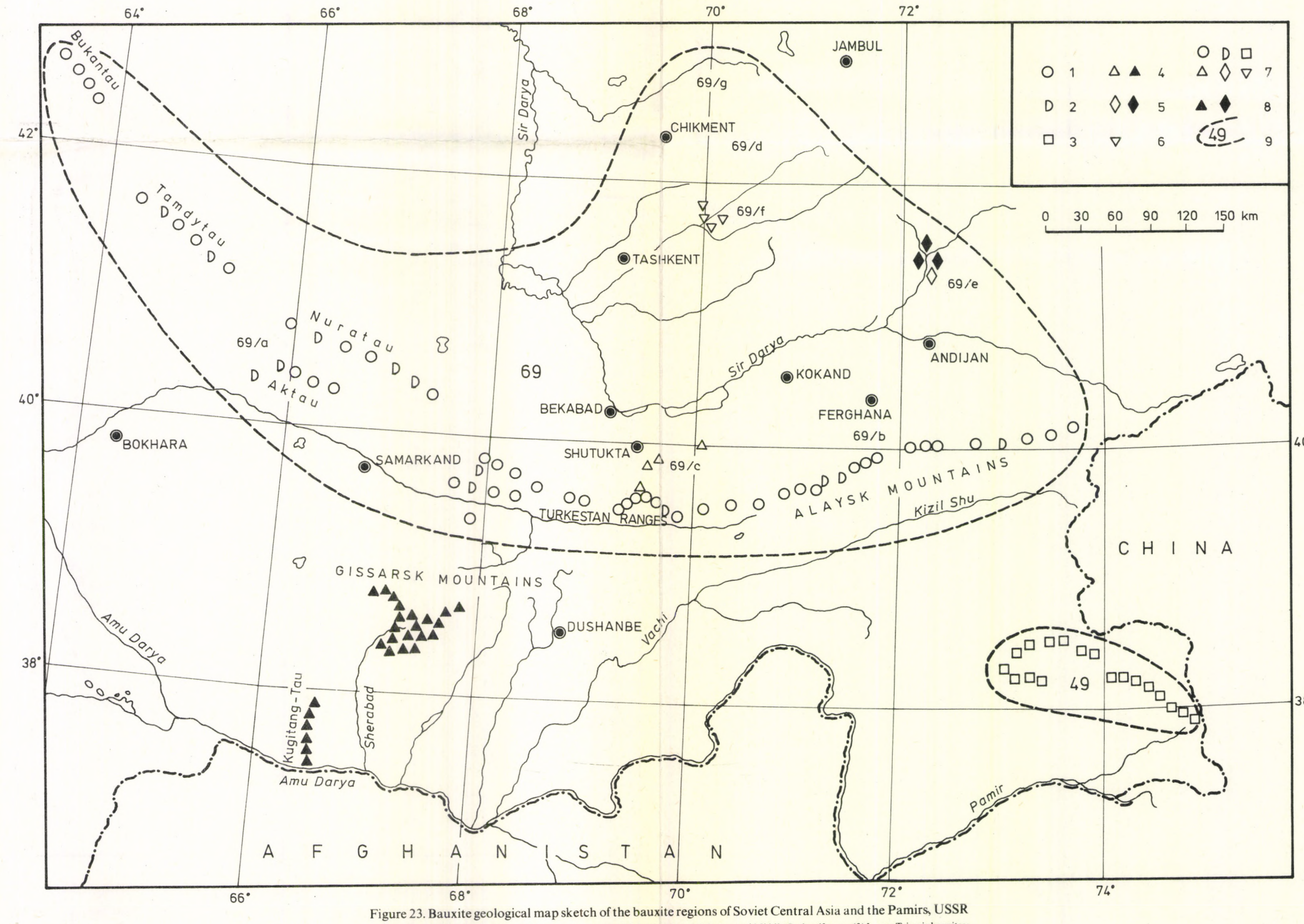


Figure 23. Bauxite geological map sketch of the bauxite regions of Soviet Central Asia and the Pamirs, USSR

(1) Lower and Middle Carboniferous bauxites; (2) metamorphic bauxite (emery), Lower and Middle Carboniferous; (3) Lower Triassic bauxites; (4) Upper Triassic bauxites; (5) Lower Liassic bauxites; (6) Lower Cretaceous bauxites; (7) karst bauxite deposits; (8) Tikhvin-type and lateritic bauxite deposits; (9) serial number and areal extent of bauxite deposit group

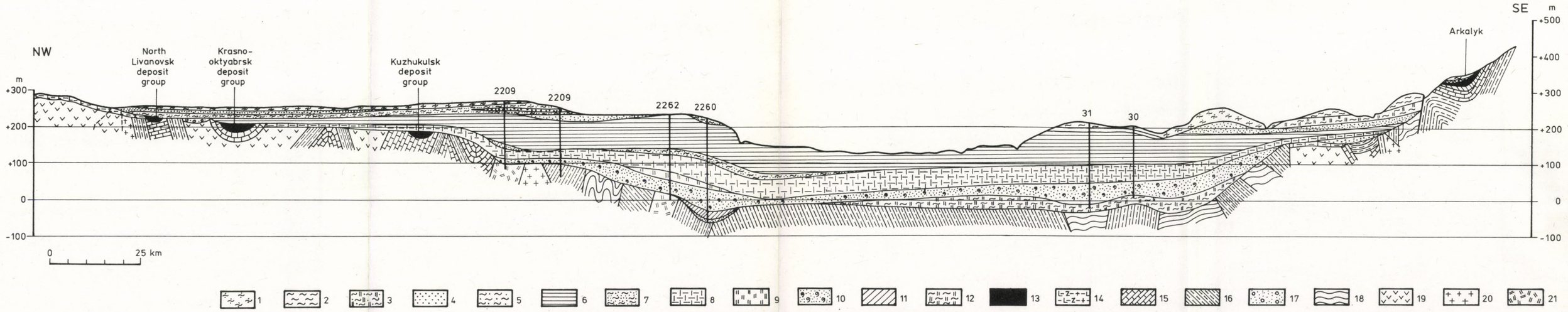


Figure 61. A synoptic geological profile of the Turgay Plateau, USSR, according to Krivtsov (1968, 1969)

(1) soil and clay, Quaternary; (2) clay, Neogene; (3) multicoloured clay, Paleogene; (4) quartz sand; (5) sandy clay; (6) lamellar clay; (7) clayey sand; (8) diatomaceous earth and clayey diatomaceous earth (3 to 8 — Paleogene); (9) clay, marl; (10) glauconitic sand; (11) clay, Albian–Cenomanian; (12) multicoloured kaolinitic clay; (13) bauxite, bauxitic clay; (14) a mesozoic zone of weathering on top of the Paleozoic; (15) carbonate rocks; (16) clayey–sandy beds; (17) coarse clastic rocks (15 to 17 – Paleozoic); (18) early Paleozoic metamorphics; (19) basic and ultrabasic igneous rocks; (20) granite; (21) acid volcanics

Appendix 1. (cont.

Serial number	Bauxite region (country) Bauxite deposit group	Deposit type	Bauxite tonnage (10^6 t)	Deposit form	Structural position	Degree of structural deformation	Age of underlying beds	Period of bauxite accumulation	Age of overlying	Reliability of stratigraphy	Facies and lithology of immediate bedrock	Facies and lithology of immediate cover	Bedrock karstification	Thickness of deposit avg. (m)	Thickness of deposit max.	Reference
1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
25	Crna Gora (Yugoslavia)															
	A. Gornje Polje, Seljoni	Med	10	Lent.	Miogeosyn. zone	Strong	Ladinian	U. Ladinian–L. Carnian	Carnian	Cert.	Neritic limestone	Swampy carbonaceous clay	Medium	3–5	10	Burić 1966
	B. Viduša Mountains	Med	5	Lent. > nest	Miogeosyn. zone	Strong	Callovian	L. Malm	Tithonian	Hyp.	Neritic limestone	Neritic limestone	Medium			Burić 1966
	C. Nikšićka Župa	Med	120	Lent. > strip	Miogeosyn. zone	Strong	Norian–Rhaetic	Malm	Valanginian	Hyp.	Neritic limestone	Brackish calcareous marl	Strong > medium	10–15	65	Burić 1966; Grubić 1975
	D. Dragalj	Med	5	Lent. > nest	Miogeosyn. zone	V. strong	Oxfordian–Kimmeridgian	Malm	Berriasian–Valanginian	Hyp.	Neritic limestone	Neritic limestone	Medium	2–4	10	Burić 1966
	E. Stara Crna Gora (Lješev stup)	Med	10	Lent. > nest	Miogeosyn. zone	V. strong	Oxfordian–Tithonian	Cenomnian	Cenomanian	Hyp.	Neritic limestone	Neritic limestone	Medium	2–4	10	Burić 1966
	F. Petrovići, Velimlje, Crvena Kita	Med	30	Lent. > nest	Miogeosyn. zone	Strong	Tithonian–Hauterivian	Cenomnian	Cenomanian	Hyp.	Neritic limestone	Neritic limestone	Medium	4–8	40	Burić 1966
	G. Bijele Poljane	Med	70	Lent.	Miogeosyn. zone	Strong	Liassic–Dogger	Cenomnian	U. Cenomanian	Hyp.	Neritic limestone	Swampy carbonaceous clay	Weak > medium	3–6	15	Burić 1966; Dragović 1965, 1966
	H. Coastal zone (Ulcinj)	Med	50	Lent. > strat.	Miogeosyn. zone	Strong	Campanian	Palaeocene	M. Eocene	Hyp.	Neritic limestone	Neritic limestone	Medium > strong	2–5	40	Burić 1966
26	Vlasenica (Yugoslavia)	Med	70	Lent. > strat.	Eugeosyn. zone	Medium	Ladinian–Carnian	L. Cretceous	Cenomanian	Hyp.	Neritic limestone	Freshwater sandstone	Medium > strong	8–10	40	Grubić 1975
27	Grebnička Planina (Yugoslavia)	Med	10	Lent.	Eugeosyn. zone	Strong	Turonian	Senonia	Maestrichtian	Cert.	Neritic limestone	Neritic limestone	Medium			Misirlić and Mitrović 1969
28	Tran (Bulgaria)	Med	3	Lent. > nest	Eugeosyn. zone	Strong	Tithonian	Cenomnian	Turonian	Hyp.	Neritic limestone	Brackish clay	Medium	1–3	10	Karanov 1965
29	Albania															
	A. Albanian Alps	Med	1	Lent. > nest	Eugeosyn. zone	V. strong	Ladinian	Ladinian–Carnian	Carnian	Quest.	Neritic limestone	Neritic limestone	?	?	?	Peza *et al.* 1969
	B. Coastal Range	Med	3	Nest > lent.	Miogeosyn. zone	Strong	Senonian	L. and M. Eocene	M. Eocene	Hyp.	Neritic limestone	Neritic limestone	Strong > medium	3–5	30	Duro *et al.* 1969
	C. Gramsh	Med	1	Nest	Miogeosyn. zone	Strong	M. Eocene	M. Eocne	U. Eocene	Hyp.	Neritic limestone	Neritic limestone	Medium	3–5	10	
30	Central Greece															
	A. Atalanti, Kallidromon, Helicon Range	Med	25	Lent. > nest	Eugeosyn. zone	Strong	Dogger	L. Malm	Kimmeridgian	Cert.	Neritic limestone	Neritic limestone	Medium > weak	2–5	10	Papastamatiou 1964, 1965
	B. Euboia Island (Lower Malm)	Med	2	Nest > lent.	Eugeosyn. zone	Strong	Dogger	L. Malm	Kimmeridgian	Cert.	Neritic limestone	Neritic limestone	Medium	1–3	5	Papastamatiou 1964, 1965; Guernet and Robert 1973
	C. Kiona, Parnassos, Helicon Range (Malm)	Med	140	Lent. > nest	Eugeosyn. zone	Strong	Kimmeridgian	U. Malm	Tithonian	Cert.	Neritic limestone	Neritic limestone	Medium > strong	8–12	25	Papastamatiou 1964, 1965; Bárdossy and Mack 1967; Nia 1968
	D. Kiona, Parnassos, Helicon Range (Turonian)	Med	250	Dol. > lent.	Eugeosyn. zone	Strong	Cenomanian–L. Turonian	Turonin	U. Turonian	Cert.	Neritic limestone	Neritic limestone	Strong > medium	10–15	25	Papastamatiou 1964, 1965; Bárdossy and Mack 1967; Nia 1968
	E. Euboia Island (Lower Senonian)	Med	10	Nest	Eugeosyn. zone	Strong	L. Turonian	L. Senonian	U. Senonian	Hyp.	Neritic limestone	Neritic limestone	Medium	?	?	Papastamatiou 1964, 1965; Robert 1971
31	Pelagonian Zone (Greece)															
	A. Skopelos Island	Ar > lat	1	Nest	Eugeosyn. zone	Strong	L. Liassic	Liassic	U. Liassic	Quest.	Neritic dolostone	Neritic shale	Medium	0.5–1	2	Papastamatiou 1963, 1964
	B. Volos	Med	1	Nest	Eugeosyn. zone	V. strong	Permo–Triassic	Triassic	Jurassic	Quest.	Metamorphic marine limestone	Metamorphic slate	Medium	1–2	4	Tataris 1971
32	Vardar Zone (Greece)															
	A. Chalkidike Peninsula	Med	1	Dol. > nest	Eugeosyn. zone	V. strong	Kimmeridgian	U. Malm	Tithonian–L. Cretaceous	Hyp.	Neritic limestone	Neritic limestone	Medium	2–5	20	Bárdossy *et al.* 1973
33	Eleusis, Mandra, Megara (Greece)	Med	50	Lent.	Eugeosyn. zone	V. strong	U. Triassic	Albian	L. Cenomanian	Hyp.	Neritic limestone	Brackish limestone conglomerate	Weak	5–8	4	Papastamatiou 1964, 1965; Aronis 1955; Papastamatiou and Maksimović 1970
34	Gavrovo Zone (Greece)															
	A. Peloponnese Peninsula	Med	3	Lent. > nest	Miogeosyn. zone	Medium	M. Eocene	M. Eocne	U. Eocene	Cert.	Neritic limestone	Brackish limestone	Weak	0.5–2	4	Kiskyras 1958, 1962
	B. Klokova	Med	1	Nest	Miogeosyn. zone	Medium	M. Eocene	M. Eocne	U. Eocene	Cert.	Neritic limestone	Neritic limestone	Weak	0.5–1	3	Kiskyras 1962
35	Aegean Islands															
	A. Amorgos	Med	1	Lent.	Eugeosyn. zone	V. strong	Liassic	Malm	Malm	Quest.	Marine limestone	Neritic limestone	Medium	1–3	10	Marinos 1954; Papastamatiou 1964
	B. Samos	Med	1	Lent. > nest	Eugeosyn. zone	V. strong + reg. metam.	Permian	Permo-Triassic	Triassic	Quest.	Metamorphic limestone	Metamorphic limestone	?	1–3	6	de Lapparent 1937
	C. Naxos	Med	2	Lent. > nest	Eugeosyn. zone	V. strong + reg. metam.	Permian	Permo-Triassic	Triassic	Quest.	Metamorphic limestone	Metamorphic limestone	?	1–3	6	Papastamatiou 1951
36	Menderes Massif (Turkey)	Med	60	Lent. > strat.	Eugeosyn. zone	V. strong + reg. metam.	U. Permian	Permo-Triassic	L. Triassic	Hyp.	Metamorphic marine limestone	Metamorphic neritic limestone	Weak	2–5	10	Önay 1949
37	Western Taurus (Turkey)															
	A. Alanya	Med	10	Lent.	Miogeosyn. zone	Strong	L. Permian	U. Perian	U. Permian–L. Triassic	Cert.	Neritic limestone	Neritic shale	Weak	2–5	8	de Peyronnet 1970
	B. Akseki, Seydisehir, Sütlegen	Med	60	Lent. > nest	Miogeosyn. zone	Strong	Cenomanian	Cenomnian	Cenomanian	Cert.	Neritic limestone	Brackish-lacustrine limestone	Strong > medium	5–10	50	Özlü 1977
	C. Sütlegen	Med	1	Lent.	Miogeosyn. zone	Strong	Senonian	L. Eocne	L. Eocene	Hyp.	Marine limestone	Marine limestone	?	?	?	Wippern 1964
38	Eastern Taurus (Turkey)															
	A. Adama, Bolkardag, Saimbeyli	Med	10	Lent. > nest	Miogeosyn. zone	Strong	L. Permian	U. Perian	U. Permian–L. Triassic	Hyp.	Neritic limestone	Neritic shale, dolostone	Weak	2–5	8	Wippern 1964
	B. Küçük Koras, Sebilköy	Med	5	Lent.	Miogeosyn. zone	Strong	Albian–Cenomanian	Turonia	Senonian	Hyp.	Neritic limestone	Neritic limestone	Strong	5–10	8	Göksu 1954; Wippern 1964
39	Payas, Islahye (Turkey)	Med	30	Lent.	Miogeosyn. zone	V. strong	Turonian	Senonia	Senonian	Hyp.	Neritic limestone	Neritic limestone	Medium	5–10	20	Petrascheck 1965
40	Zonguldak (Turkey)	Sal	5	Lent.	Eugeosyn. zone	Medium	Visean	Albian	Cenomanian	Hyp.	Neritic limestone	Neritic sandstone	Medium		?	Göksu 1954
41	Crimea (USSR)	Med	1	Lent.	Intracont. orogen	Medium	Oxfordian	Kimmedgian	Tithonian	Cert.	Neritic limestone	Neritic limestone	?	2–4	8	Malakovsky and Lysenko 1964
42	Maktesh Ramon (Israel)	Med	5	Lent. > dol.	Paraplatform	V. weak	Liassic	U. Cretceous	U. Cretaceous	Hyp.	Lagoonlimestone	Swampy carbonaceous claystone	Weak	5–10	20	Slatkine and Heller 1961
43	Little Caucasus (Julfa) (USSR)	Med	1	Lent.	Eugeosyn. zone	Strong	U. Carbon–L. Permian	L. Triaic	Triassic	Quest.	Marine dolostone	?	?	1–2	?	Mikhailov *et al.* 1973
44	Bukan (Iran)	Med	10	Strat.	Eugeosyn. zone	V. strong	U. Permian	Permo-Triassic	L. Triassic	Quest.	Neritic limestone	Neritic limestone	V. weak	2–4	12	
45	Alborz Range (Iran)															
	A. Jajarm (Permo–Triassic)	Med	5	Strat.	Eugeosyn. zone	Strong	U. Permian	Permo-Triassic	L. Triassic	Quest.	Marine dolostone	Marine claystone	V. weak	2–4	10	
	B. Jajarm (Upper Triassic)	Med	15	Strat.	Eugeosyn. zone	Strong	Triassic	U. Triaic	L. Liassic	Quest.	Marine dolostone	Marine claystone	V. weak	2–3	6	
46	Central Iranian Plateau															
	A. Yazd	Med	10	Transitional lent.–strat.	Eugeosyn. zone	Strong	U. Permian	U. Triaic	L. Liassic	Cert.	Marine dolostone	Carbonaceous claystone	Weak	2–4	12	
	B. Kerman	Med	10	Transitional lent.–strat.	Eugeosyn. zone	Strong	L. Triassic	U. Triaic	Rhaetic	Cert.	Marine dolostone	Brackish shale	Weak	2–3	9	Huckriede *et al.* 1962
47	Zagros Range (Iran)															
	A. Do-Polan, Do-Rag	Med	10	Lent.	Miogeosyn. zone	Strong	U. Permian	Permo-Triassic	L. Triassic	Quest.	Neritic dolostone	Freshwater claystone, dolostone	Weak	2–4	6	
	B. Behbehan (Sar-e-Fariab)	Med	10	Lent.	Miogeosyn. zone	Strong	Cenomanian–Turonian	L. Senonian	Santonian	Quest.	Reef limestone	Freshwater limestone	V. weak	1–2	5	
	C. Shiraz (Daryush Dam)	Med	1	Lent.	Miogeosyn. zone	V. strong	M. U. Cretaceous	U. Creceous	U. Cretaceous	Quest.	Marine limestone	Marine limestone	Weak	1–2	4	
48	Kandahar (Afganistan)	Med	5	Strat. > lenth.	Eugeosyn. zone	V. strong	Malm	L. Cretceous	L. Cretaceous	Quest.	Marine limestone	Dryland sandstone	Weak	1–5	10	Bushinsky 1971, 2nd ed. 1975
49	Pamirs (USSR)	Med	10	Lent.	Miogeosyn. zone	V. strong	U. Permian	L. Triaic	L. Triassic	Hyp.	Neritic dolostone	Swampy carbonaceous claystone	Medium > weak	2–3	10	Karapetov 1968; Konnov 1972; Krivtsov 1968–1969
50	Hindukush Range (Pakistan)															
	A. Damun Gol, Chitra State	Med	1	Lent.	Eugeosyn. zone	V. strong	L. M. Devonian	M. Devnian	U. Devonian	Quest.	Neritic dolostone, limestone	Neritic limestone	Medium	1–2	7	Stauffer 1969
51	Northeast Pakistan															
	A. Salt Range, Sargodha	Med	10	Lent.	Miogeosyn. zone	Medium	U. Permian	Palaeone	Palaeocene–Eocene	Quest.	Marine limestone	Swampy clay, sandstone	Medium	2–3	6	Ashraf *et al.* 1972
	B. Campbellpore, Muzaffabad	Med	10	Lent.	Miogeosyn. zone	Strong	U. Triassic–Liassic	Dogger	Malm	Quest.	Marine limestone	Lagoon clay, coal	Medium	2–4	8	Patterson 1967
52	Jammu (Kashmir-India)	Med	20	Lent.	Miogeosyn. zone	Strong	Permo–Carb	Palaeone–L. Eocene	Eocene	Quest.	Marine limestone	Swampy clay, coal	Medium	1–3	6	Caillère and Singh 1967; Patterson 1967
53	Samar Island (Philippines)	Med	50	Mantle > dol.	Eugeosyn. zone	Quiet	Plio–Pleistocene	Pleistone–recent	—	Cert.	Reef limestone	None	Medium > strong	2–4	8	de Weisse 1976
54	Solomon Islands															
	A. Wagina and other islands	Med	30	Mantle	Eugeosyn. zone	Quiet	Pleistocene	Pleistone–recent	—	Cert.	Reef limestone	None	Medium	1–2	4	
	B. Rennell Island	Med	30	Mantle	Eugeosyn. zone (ocean island arc)	Quiet	Pleistocene	Pleistone–recent	—	Cert.	Reef limestone	None	Medium	1–2	4	de Weisse 1970; Taylor and Hunghes 1975
55	Loyalty Group															
	A. Lifou Island	Med	5	Mantle	Eugeosyn. zone (ocean island arc)	Quiet	Pleistocene	Pleistone–recent	—	Cert.	Reef limestone	None	Medium > weak	0.1–0.5	2	Tercinier 1971; de Weisse 1976
	B. Maré Island	Med	5	Mantle	Eugeosyn. zone (ocean island arc)	Quiet	Pleistocene	Pleistone–recent	—	Cert.	Reef limestone	None	Medium > weak	0.1–0.5	2	Tercinier 1971; de Weisse 1976
56	Fiji-Tonga Group															
	A. Lau Islands (Fiji)	Med	1	Mantle	Eugeosyn. zone (ocean island arc)	Quiet	Pleistocene	Pleistone–recent	—	Cert.	Reef limestone	None	Medium > weak	0.1–0.5	1	Patterson 1967
	B. Niué Island (Tonga)	Med	1	Mantle	Eugeosyn. zone (ocean island arc)	Quiet	Pleistocene	Pleistone–recent	—	Cert.	Reef limestone	None	Medium > weak	0.1–0.5	2	Schofield 1959

Appendix 1. Principal geological and depositological features of the world's karst bauxite deposit groups

Serial number	Bauxite region (country) Bauxite deposit group	Deposit type	Bauxite tonnage (10^6 t)	Deposit form	Structural position	Degree of structural deformation	Age of underlying beds	Period of bauxite accumulation	Age of overlying	Reliability of stratigraphy	Facies and lithology of immediate bedrock	Facies and lithology of immediate cover	Bedrock karstification	Thickness of deposit avg. (m)	Thickness of deposit max.	Reference
1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
1	Andorinha (Portugal)	Med	1	Lent. > dol.	Cons. epi-Hercynian zone	Weak	Dogger	Barremian–Aptian	Aptian–Cenomanian	Hyp.	Neritic limestone	Swampy carbonaceous clay	Medium	2–5	15	Figueredo Gomes 1965
2	Betic Cordillera (Spain)	Med	1	Lent. > dol.	Miogeosyn. zone	Strong	Liassic	Dogger	Malm (?)	Quest.	Neritic limestone	Neritic limestone	Medium	2–5	10	Alias *et al.* 1972
3	Catalonia (Spain)															
	A. Maestrazgo	Med	10	Lent. > dol.	Intracont. orogen	Medium	Kimmeridgian–Oxfordian	Aptian	U. Aptian	Cert.	Neritic limestone	Swampy carbonaceous clay	Strong > medium	2–5	15	Combes 1969
	B. Ebro Massif	Med	5	Dol. > nest	Median mass	Quiet	L. Liassic	Palaeocene	Palaeocene–Eocene	Hyp.	Neritic limestone, dolostone	Freshwater conglomerate	Strong > medium	2–4	7	Combes 1969
4	Spanish Pyrenees (Spain)	Med	10	Lent. > nest	Intracont. orogen	Strong	Liassic–Dogger	Cenomanian–Turonian	Senonian (Santonian)	Cert.	Neritic limestone, dolostone	Freshwater sandstone	Medium	2–4	10	Combes 1969
5	French Pyrenees (France)															
	A. Ariège	Ar	10	Lent. > dol.	Intracont. orogen	Strong	Kimmeridgian–Barremian	Barremian–Albian	Aptian–Albian	Cert.	Reef limestone, dolostone	Reef limestone, swampy coloured clay	Medium > strong	3–5	20	Combes 1969
	B. Saint Paul de Fenouillet	Ar	1	Lent. > nest	Intracont. orogen	V. strong	Neocomian–Aptian	U. Aptian–Albian	Aptian–Albian	Cert.	Neritic limestone	Reef limestone	Medium	2–4	8	de Lapparent 1930
6	Languedoc (France)															
	A. Cazouls, St Chinian	Med	30	Lent.	Intracont. orogen	Strong	Dogger–Hettangian	Turonian–Senonian	Coniacian–Santonian	Cert.	Neritic dolostone	Brackish sandy clay marl	Medium	5–10	20	de Lapparent 1930
	B. Bédarieux	Med	50	Strat. > lent. > dol.	Intracont. orogen	Weak	Dogger–Malm	L. Cretaceous–Senonian	Campanian–Maestrichtian	Cert.	Neritic dolostone	River-laid conglomerate	Medium > strong	5–10	50	Combes 1971
	C. Villeveyrac	Med	100	Strat. > nest	Intracont. orogen	Medium	Oxfordian–Tithonian	Coniacian–Santonian	Campanian–Maestrichtian	Cert.	Neritic limestone	Brackish claystone, marl	Medium > strong	4–8	15	de Lapparent 1930
	D. La Boissière	Med	5	Lent. > nest	Intracont. orogen	Weak	Malm	Senonian–Danian	Palaeocene	Hyp.	Neritic limestone	River-laid conglomerate	Medium > strong	2–4	10	de Lapparent 1930
7	Provence (France)															
	A. Alpilles	Med	160	Lent. > strat.	Intracont. orogen	Medium	U. Hauterivian–L. Aptian	Senonian	Campanian–Maestrichtian	Cert.	Neritic limestone	Lacustrine limestone	Medium	10–15	30	de Lapparent 1930
	B. Revest and Allauch	Med	5	Lent.	Intracont. orogen	Strong	Barremian–L. Aptian	Albian	U. Cenomanian–Turonian	Cert.	Neritic limestone	Swampy clay	Medium	2–3	6	Nicolas 1970
	C. Brignoles	Med	150	Strip > lent.	Intracont. orogen	Strong > v. strong	Bathonian–Tithonian	Cenomanian–Turonian	U. Cenomanian–Santonian	Cert.	Neritic limestone, dolostone	Swampy carbonaceous clay, brackish clay marl	Medium	5–10	20	Nicolas 1970
	D. Haut–Var	Med	10	Dol.	Intracont. orogen	Medium	Kimmeridgian–Tithonian	Senonian–Danian	Maestrichtian–Palaeocene	Cert.	Neritic dolostone, limestone	River-laid sandy clay	Strong	10–15	50	Roch, 1959
8	Vanoise (France)	Med	1	Nest	Miogeosyn. zone	V. strong	Triassic	Dogger	Bottom of Malm	Hyp.	Neritic limestone, dolostone	Neritic limestone	Medium	0.5–2	3	Ellenberger 1955; Goffé and Saliot 1977
9	Dréveneuse (Switzerland)	Med	1	Lent.	Miogeosyn. zone	V. strong	Ladinian	L. Dogger	Bathonian	Cert.	Neritic limestone	Brackish limestone	Weak	2–4	7	Badoux and de Weisse 1959
10	Sardinia–Nurra (Italy)	Med	5	Lent. > nest	Median mass	Weak	Neocomian	Aptian–Albian	Cenomanian	Cert.	Neritic limestone	Neritic limestone	Medium	1–2	5	Cocco and Pecorini 1959
11	Central Italy															
	A. Abruzzi	Med	20	Lent. > nest	Miogeosyn. zone	V. strong	Albian–Cenomanian	Cenomaian	U. Cenomanian	Cert.	Neritic limestone	Nearshore limestone	Weak > medium	2–4	10	Bárdossy *et al.* 1977
	B. Campania	Med	5	Lent. > nest	Miogeosyn. zone	V. strong	Albian	Cenomaian–Turonian	U. Cenomanian–Turonian	Cert.	Neritic limestone	Brackish calcareous marl	Weak	1–2	5	Bárdossy *et al.* 1977
12	Apulia (Italy)															
	A. San Giovanni Rotondo	Med	15	Lent.	Miogeosyn. foreland	Weak	Albian–L. Cenomanian	Cenomaian–Turonian	Turonian	Cert.	Neritic limestone	Reef limestone	Medium	4–6	15	Bárdossy *et al.* 1977
	B. Spinazzola	Med	5	Dol. > cany.	Miogeosyn. foreland	Weak	L. Cenomanian	Cenomaian–L. Turonian	L. Turonian	Cert.	Neritic limestone	Brackish multicoloured clay	Strong	8–10	40	Bárdossy *et al.* 1977
13	Terra D'Otranto (Italy)	Sal	5	Lent.	Miogeosyn. foreland	Weak	Senonian–Turonian	Oligocene–Miocene	L. Miocene	Hyp.	Neritic limestone	Swampy carbonaceous clay	Medium	5–8	14	Cotecchia and Dell' Anna 1959
14	Eastern Alps (Austria)															
	A. Tirol, Salzburg	Med	1	Lent.	Miogeosyn. zone	V. strong	Norian–Rhaetic	Turonia–Coniacian	Coniacian–Santonian	Hyp.	Neritic dolostone, limestone	Swampy carbonaceous clay	Medium	2–3	5	Ruttner 1970
	B. Unterlaussa, Dreistätten	Med	5	Lent. > nest	Miogeosyn. zone	V. strong	Norian–Rhaetic	Turonia–Coniacian	Senonian	Cert.	Neritic dolostone	Swampy carbonaceous clay	Medium > strong	2–5	20	Ruttner 1970; Haberfelner 1951
15	Mojtin (Czechoslovakia)	Med	1	Dol. > nest	Miogeosyn. zone	V. strong	M. Triassic	Turonia–Senonian	Eocene	Hyp.	Neritic limestone, dolostone	?Conglomerate	Medium	1–3	15	Orlov 1937; Konta and Kužvart 1962
16	Markušovce, Drienovec (Czechoslovakia)	Med	1	Dol. > nest	Miogeosyn. zone	Strong	M. Triassic	Turonia–Senonian	Palaeocene	Quest.	Neritic limestone	?Conglomerate	Medium	1–2	5	Konta and Kužvart 1962
17	Rahov Massif (USSR)	Med	2	Lent.	Miogeosyn. zone	Strong	U. Triassic	L. Dogger	Bajocian	Hyp.	Neritic limestone	Neritic limestone	Medium	1–2	?	Denisevich 1958
18	Pădurea Craiului (Rumania)	Med	50	Dol. > lent.	Eugeosyn. zone	Strong	Tithonian	Neocomian	Neocomian	Hyp.	Neritic limestone	Freshwater limestone	Strong	4–8	20	Papiu *et al.* 1969, 1970a
19	Southern Transylvania (Rumania)															
	A. Sohodol-Cîmpeni	Med > Kaz	5	Lent. > dol.	Eugeosyn. zone	Weak-medium	Precambrian	Senonian	Senonian	Hyp.	Metamorphic limestone	Swampy carbonaceous clay	Medium	2–5	10	Papiu *et al.* 1970a
	B. Hațeg	Med > Kaz	5	Lent. > dol.	Eugeosyn. zone	Weak-medium	Malm-L. Cretaceous	U. Aptian–Albian	Cenomanian	Hyp.	Neritic limestone	Nearshore standstone	Medium	3–6	20	Papiu *et al.* 1970a, 1971
20	Transdanubian Hills															
	A. Alsópere, Tés	Med	5	Lent.	Interarc basin	Weak	Rhaetic–L. Liassic	L. Aptian	U. Aptian	Cert.	Neritic limestone	Brackish clay, marl	Medium	1–3	15	Barnabás 1970
	B. Halimba, Padragkút, Sümeg	Med	75	Strat. > lent.	Interarc basin	Weak	Norian, Rhaetic–L. Liassic	L. Senonian	Santonian	Cert.	Neritic dolostone, limestone	Lacustrine, river-laid marl, conglomerate	Medium	5–10	35	Barnabás 1970
	C. Iharkút, Bakonyjákó, Bakonybél	Med	15	Grab. > lent. > dol.	Interarc basin	Weak	Norian	L. Senonian	U. Senonian–M. Eocene	Cert.	Neritic dolostone	Freshwater multicoloured clay, sandstone	Strong > medium	20–30	100	
	D. Nyirád, Nagytárkány	Med	60	Dol. > lent.	Interarc basin	Weak	Norian–Carnian	Senonian–Palaeocene	L. Eocene–Miocene	Cert.	Neritic dolostone	Swampy carbonaceous clay	Strong > medium	5–10	50	Barnabás 1970; Károly *et al.* 1970
	E. Szőc Basin, Halimba–Malomvölgy	Med	25	Lent.	Interarc basin	Weak	Norian–Carnian	Senonian–Palaeocene	L. Eocene–M. Eocene	Cert.	Neritic dolostone	Swampy carbonaceous clay	Medium	4–7	20	Barnabás 1970
	F. Kislőd, Ajka	Med	10	Lent.	Interarc basin	Weak	Norian–Carnian	Palaeocene	M. Eocene	Cert.	Neritic dolostone	Brackish clay, marl	Medium	5–10	20	Barnabás 1970
	G. Fenyőfő, Dudar, Bakonyoszlop	Med	35	Lent. > grab.	Interarc basin	Weak	Norian–Rhaetic	L. Eocene	M. Eocene, U. Oligocene	Cert.	Neritic dolostone, limestone	Swampy clay, marl	Medium	5–10	60	Barnabás 1970
	H. Iszkaszentgyörgy	Med	50	Strat. > lent.	Interarc basin	Weak	Ladinian–Carnian–Norian	L. Eocene	M. Eocene	Cert.	Neritic dolostone	Swampy carbonaceous clay	Medium	5–8	16	Komlóssy 1970
	I. Gánt	Med	60	Strat. > lent. > grab.	Interarc basin	Weak	Carnian–Norian	L. and M. Eocene	M. Eocene	Cert.	Neritic dolostone	Freshwater multicoloured clay, limestone	Medium > strong	10–20	40	Kiss 1955
	J. Nagyegyháza, Mány	Med	30	Lent.	Interarc basin	Weak-medium	Norian	L. Eocene–Palaeocene	M. Eocene	Hyp.	Neritic dolostone	Dryland dolostone detritus	Medium	3–6	20	Sóki 1973; Tóth 1976
	K. Nézsa	Med	1	Lent.	Interarc basin	Weak	Rhaetic	L. Eocene	M. Eocene	Cert.	Neritic limestone	Brackish clay, limestone	Medium	5–8	28	Kiss 1952
	L. Óbarok, Újbarok	Sal	5	Lent. > dol.	Interarc basin	Weak	Norian	L. and M. Oligocene	U. Oligocene	Cert.	Neritic dolostone, limestone	Brackish clay, sand	Medium	2–5	20	Jaskó 1957a
	M. Nagyvázsony, Öcs	Sal	30	Strat. > lent.	Interarc basin	Weak	Norian	Neogene	Pliocene	Cert.	Neritic dolostone	Freshwater limestone	Medium	10–20	35	Jaskó 1957b
21	Nagyharsány	Med	5	Nest > lent.	Interarc basin margin	Medium	L. Tithonian	Valanginian–Hauterivian	L. Barremian	Cert.	Neritic limestone	Nearshore limestone	Weak	1–3	8	Fülöp 1966
22	Slovenia (Yugoslavia)															
	A. Podlipa, Borovnica	Med > Ar	50	Lent. > strat.	Miogeosyn. zone	Medium	U. Ladinian	Ladinian–Carnian	Carnian	Cert.	Neritic dolostone, limestone	Nearshore claystone	Weak > v. weak	2–3	23	Bušer and Lukács 1970
	B. Hrušica Mountains, Žužemberk	Med	3	Lent. > nest	Miogeosyn. zone	Strong	Oxfordian–L. Kimmeridgian	Kimmeridgian	U. Kimmeridgian	Cert.	Neritic limestone	Brackish limestone	Weak > medium	1–3	5	Bušer and Lukács 1970
	C. Vinica, Metlika	Med	1	Lent. > dol.	Miogeosyn. zone	Weak	L. Cretaceous	U. Cretaceous	U. Cretaceous	Quest.	Neritic limestone	Neritic limestone	Medium > strong	2–5	25	Bušer and Lukács 1970
	D. Nadrt	Med	1	Nest	Miogeosyn. zone	Medium	L. Turonian	Turonian	U. Turonian	Cert.	Neritic limestone	Nearshore limestone breccia	Weak	0.3–0.5	4	Bušer and Lukács 1970
	E. Kozina	Med	3	Nest	Miogeosyn. zone	Strong	Turonian–Santonian	Palaeocene	Palaeocene	Cert.	Neritic limestone	Lacustrine limestone	Medium > weak	0.5–1	3	Bušer and Lukács 1970
	F. Savinja Valley	Med	3	Lent. > dol.	Miogeosyn. zone	Medium	U. Ladinian	Oligocene	M. Oligocene	Quest.	Neritic limestone, dolostone	Nearshore limestone	Medium	2–3	8	Bušer and Lukács 1970
	G. Velika planina	Med	1	Nest > dol.	Miogeosyn. zone	Strong	U. Ladinian	Oligocene	M. Oligocene	Quest.	Neritic dolostone	Nearshore limestone, marl	Medium > strong	1–3	5	Bušer and Lukács 1970
	H. Bohinj	Med	1	Lest. > nest	Miogeosyn. zone	Strong	Rhaetic	Oligocene	U. Oligocene	Quest.	Neritic limestone	Freshwater limestone	Medium	1–2	3	Bušer and Lukács 1970
23	Istria (Yugoslavia)															
	A. Rovinj, Vrsar	Med	1	Lent.	Miogeosyn. foreland	Weak	Kimmeridgian–Tithonian	Tithonian	U. Tithonian–Berriasian	Hyp.	Neritic limestone	Neritic limestone	Weak > medium	5–10	11	Grubić 1975
	B. Central and Southern Istria	Med	40	Deep dol.	Miogeosyn. foreland	Weak	Campanian	Palaeocene	Palaeocene–L. Eocene	Cert.	Neritic limestone	Freshwater limestone	Strong	10–15	50	Bušer and Lukács 1972
24	Dalmatia, Bosnia, Hercegovina															
	A. Lika, Bjelaj-Lipa	Ar > Med	130	Strat. > lent.	Miogeosyn. zone	Medium	Ladinian	Ladinian–Carnian	Carnian	Cert.	Neritic limestone, marl	Freshwater limestone conglomerate	Weak	5–10	25	Šinkovec 1970
	B. Knin, Marići	Med	1	Lent.	Miogeosyn. zone	Medium	U. Ladinian–Carnian	Rhaetic	Hettangian	Hyp.	Neritic limestone	Neritic limestone	Medium	?	?	de Weisse 1948
	C. Grmeć Mountains, Bosanska Krupa (Lower Cretaceous)	Med	5	Lent. > nest	Miogeosyn. zone	Strong	Tithonian–Valanginian	L. Cretaceous	Barremian	Cert.	Neritic limestone	Neritic limestone	Medium	2–5	15	Jurković 1965
	D. Ogorje	Med	1	Lent. > nest	Miogeosyn. zone	Strong	L. Cretaceous	Albian–Cenomanian	Cenomanian–Turonian	Hyp.	Neritic limestone	Nearshore limestone breccia	Medium	?	?	Sakač 1972
	E. Grmeć Mountains, Bosanska Krupa (Senonian)	Med	25	Lent. > nest	Miogeosyn. zone	Strong	Turonian	Senonian	Campanian–Maestricht.	Cert.	Neritic limestone	Nearshore limestone conglomerate	Weak > medium	0.5–3	5	Grubić 1975
	F. Jajce	Med	25	Dol. > lent.	Miogeosyn. zone	Strong	Santonian	Campanian	Maestrichtian	Cert.	Neritic limestone	Nearshore limestone conglomerate	Strong > medium	2–12	40	Grubić 1975
	G. Obrovac, Novigrad, Drniš (Palaeocene)	Med	3	Lent. > nest	Miogeosyn. zone	V. strong	Campanian	Palaeocene	Palaeocene	Cert.	Neritic limestone	Freshwater limestone	Medium	2–4	10	Grubić 1975
	H. Imotski, Mostar (Palaeocene)	Med	25	Dol. > nest	Miogeosyn. zone	V. strong	Santonian–Campanian	Palaeocene	Palaeocene	Hyp.	Neritic limestone	Brackish marl, clay	Strong	5–15	30	Grubić 1975
	I. Obrovac, Novigrad, Drniš (Eocene)	Med	50	Lent. > cany.	Miogeosyn. zone	V. strong	Campanian–L. Eocene	M. Eocene	M. Eocene	Cert.	Neritic limestone	Nearshore limestone conglomerate	Strong	5–10	40	Grubić 1975
	J. Imotski, Mostar (Eocene)	Med	25	Dol. > nest	Miogeosyn. zone	Strong	Santonian–L. Eocene	M. Eocene	M. and U. Eocene	Cert.	Neritic limestone	Brackish marl, sandstone	Strong	5–15	30	Grubić 1975
	K. Jajce-Baraći	Sal	5	Lent. > nest	Miogeosyn. zone	Strong	Santonian	Oligocene	Miocene	Hyp.	Neritic limestone	Freshwater limestone	Medium > strong	5–10	20	Trubelja and Vasiljević 1971

Appendix 1. (cont.)

Serial number	Bauxite region (country) Bauxite deposit group	Deposit type	Bauxite tonnage (10^6 t)	Deposit form	Structural position	Degree of structural deformation	Age of underlying beds	Piod of bauxite ccumulation	Age of overlying	Reliability of stratigraphy	Facies and lithology of immediate bedrock	Facies and lithology of immediate cover	Bedrock karstification	Thickness of deposit avg. (m)	Thickness of deposit max.	Reference
1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
79	Dominican Republic	Med	120	Mantle > lent.	Eugeosyn. zone	Weak	M. U. Eocene–Oligocene	U. Miocene–Pliocene	None	Hyp.	Neritic limestone	None	Medium > strong	2–4	10	Goldich and Bergquist 1974
80	Haiti	Med	60	Mantle > lent.	Eugeosyn. zone	Weak	Eocene	U. Miocene–Pliocene	None	Hyp.	Neritic limestone	None	Medium > strong	2–3	10	Goldich and Bergquist 1948
81	Jamaica															
	A. St Ann Plateau	Med	500	Mantle > lent.	Eugeosyn. zone	V. weak	M. U. Eocene–Oligocene	L. Miocene–Pliocene	None	Cert.	Neritic limestone	None	Strong > medium	5–10	20	Anderson 1971
	B. Cockpit Country	Med	100	Dol.	Eugeosyn. zone	V. weak	Oligocene	L. Miocene–Pliocene	None	Cert.	Neritic limestone	None	V. strong	10–20	30	Hill 1973
	C. Essex Valley	Med	200	Mantle > lent.	Eugeosyn. zone	V. weak	Oligocene–Miocene	U. Miocene–Pliocene	Pleistocene	Cert.	Neritic limestone	None > alluvium	Medium	5–10	15	Hill 1973
	D. Manchester Plateau	Med	300	Lent. > mantle	Eugeosyn. zone	V. weak	Eocene–Oligocene–Miocene	U. Miocene–Pliocene	Pleistocene	Cert.	Neritic limestone	None > lacustrine	Strong > medium	5–10	25	Hill 1973
	E. Clarendon Plateau	Med	70	Lent.	Eugeosyn. zone	V. weak	Oligocene–Miocene	U. Miocene–Pliocene	None	Cert.	Neritic limestone	None	Medium	5–10	20	Hill 1973
	F. St Catherine Plateau	Med	30	Lent.	Eugeosyn. zone	V. weak	Oligocene–Miocene	U. Miocene–Pliocene	None	Cert.	Neritic limestone	None	Medium	4–8	15	Hill 1973
82	Cayman Islands	Med	3	Mantle	Eugeosyn. zone	Quiet	Miocene	Pliocen	None	Quest.	Neritic limestone	None	Medium > weak	0.5–1	2	Ahmad and Jones 1969
83	Bahamas	Med	5	Mantle	Miogeosyn. zone	Quiet	Miocene–Pliocene	Pliocen–Pleistocene	None	Quest.	Reef limestone	None	Medium > weak	0.5–1	3	Ahmad and Jones 1969
84	Missouri (Gasconade County) (US)	Kaz > Med	20	Dol. > lent.	Continental plate	Quiet	Cambrian–Ordovician	M. Carboniferous	U. Carboniferous	Cert.	Neritic cherty dolostone	Neritic claystone	Strong > medium	5–10	70	Keller *et al.* 1954; Keller 1964
85	Western rim of Costal Plain (US)	Kaz	1	Strat. > lent.	Continental plate	Quiet	L. Carboniferous	Palaeocene–Eocene	None	Hyp.	Neritic limestone	None	Medium > strong	40–50	80	Dunlap *et al.* 1965; Bergquist and Overstreet 1965
86	SE rim of Coastal Plain (US)	Tim > Tikh	10	Strat.	Cons. epi-Hercyn. zone	Quiet	Palaeocene	Palaeocene–L. Eocene	L. Eocene	Hyp.	Neritic limestone	Nearshore glauconitic sand	Weak > medium	15–30 (2–4)	40 (6)	Clark 1965; Clarke 1966
87	SW Valley and Ridge Prov. (US)	Kaz	5	Dol. > funn.	Cons. epi-Hercyn. zone	Quiet	U. Cambrian–Ordovician	Palaeocene–L. Eocene	Eocene, Quaternary	Hyp.	Marine limestone, dolostone	Nearshore sand, clay	Strong > medium	10–40 (2–5)	130 (10)	Cloud 1967; Dunlap *et al.* 1965
88	NE Valley and Ridge Prov. (US)	Kaz	1	Dol. > funn.	Cons. epi-Hercyn. zone	Quiet	Cambrian–L. Ordovician	L. Eocene	Quaternary	Hyp.	Marine limestone, dolostone	Dryland scree	Strong > medium	?	?	Warren *et al.* 1965

Explanation of symbols:

To Column 3

Ar = Ariège
Lat = laterite
Kaz = Kazakhstan
Med = Mediterranean
Sal = Salento
Tikh = Tikhvin
Tim = Timan
Tul = Tulsk

To Column 4

Tonnages are totals of ore-grade bauxite and sub-grade bauxitic rocks, with total worked-out bauxite tonnage added wherever applicable

To Column 5

lent. = lenticular
dol. = doline
strat. = stratiform
cany. = canyon
grab. = graben
comp. = composite
vall. = valley
funn. = funnel
cont. = contact

To Column 6

miogeosyn. = miogeosynclinal
intracont. = intracontinental
eugeosyn. = eugeosynclinal
cons. = consolidated
Hercyn. = Hercynian
Baikal. = Baikalian
Caled. = Caledonian
geosyn. = geosynchnal

To Column 7

v. = very
reg. metam. = regional metamorphism

To Columns 8, 9 and 10

L. = Lower
M. = Middle
U. = Upper

To Column 11

hyp. = hypothetical
quest. = questionable
cert. = certain

To Columns 15 and 16

In Kazakhstan-type deposits: 15–30 and 80: thickness of entire bauxite complex
(2–5) and (10) aggregate thickness of bauxite ore bodies within the complex

Appendix 1. (cont.)

Serial number	Bauxite region (country) Bauxite deposit group	Deposit type	Bauxite tonnage (10^6 t)	Deposit form	Structural position	Degree of structural deformation	Age of underlying beds	Period of bauxite accumulation	Age of overlying	Reliability of stratigraphy	Facies and lithology of immediate bedrock	Facies and lithology of immediate cover	Bedrock karstification	Thickness of deposit avg. (m)	Thickness of deposit max.	Reference
1	2	3	4	5	6	7	8		9	10	11	12	13	14	15	16	17
57	Sisophon, Battambang (Campuchea)	Med	3	Lent.	Miogeosyn. zone	Medium > strong	U. Permian	U. Permian	U. Permian	Hyp.	Neritic limestone	Neritic siliceous limestone	V. weak	0.5–1.5	3	Nirkov and Gorovoy 1974; Saurin 1965
58	Socialist Republic of Vietnam															
	A. Lang-Son (Upper Permian)	Med	20	Lent.	Miogeosyn. zone	Strong	U. Permian (Kazanian)	U. Permian	U. Permian	Hyp.	Neritic limestone	Paralic swampy claystone	Medium	4–8	50	Komlóssy 1976; Chebotarev and Buy Fu-Miy 1961
	B. Lang-Son (Neogene)	Sal	50	Lent.	Miogeosyn. zone post-orogenic succession	Quiet	U. Carbon.–U. Permo	Neogene	None or Quaternary	Hyp.	Neritic limestone	Dryland scree	Medium > strong	5–10	54	Komlóssy 1976; Chebotarev and Buy Fu-Miy 1961
	C. Cao-Bang (Upper Permian)	Med	100	Lent. > strat.	Miogeosyn. zone	V. strong	L. Permian	U. Permian	U. Permian	Hyp.	Neritic limestone	Neritic limestone	Medium > weak	5–9	20	Komlóssy 1976; Chebotarev and Buy Fu-Miy 1961
	D. Cao-Bang (Neogene)	Sal	30	Lent.	Miogeosyn. zone post-orogenic succession	Quiet	U. Carbon.–U. Permian	Neogene	None or Quaternary	Hyp.	Neritic limestone	Dryland scree	Strong	4–7	30	Komlóssy 1976; Chebotarev and Buy Fu-Miy 1961
59	Kunming (Yünnan Province) (China)	Tim	130	Strat.	Paraplatform	Medium	Devonian	Carboniferous	M. U. Carboniferous	Hyp.	Neritic limestone	?	Weak	10–15	20	Bushinsky 1971; 2nd ed. 1975; Patterson 1967
60	Kweichow Province (China)															
	A. Kveyang (Carboniferous)	Tim	150	Strat. > lent.	Paraplatform	Medium	M. and U. Cambrian	M. Carbon	M. Carbon	Hyp.	Neritic limestone	Neritic limestone	Weak	5–10	15	Wang Chunghsing 1965
	B. Kveyang (Permian)	Tim	150	Strat.	Paraplatform	Medium	Cambrian–M. Carbon.	L. Permian	L. Permian	Hyp.	Neritic limestone	Paralic coal measures	Weak > medium	?	?	Wang Chunghsing 1965
61	Gun District (Honan Province) (China)	Tim	30	Strat. > lent.	Paraplatform	Weak	Ordovician	Carboniferous	U. Carboniferous	Hyp.	Neritic limestone	Paralic coal measures	Weak	2–6	?	Schüller 1957
62	Nanting, Po-San (Shantung Province) (China)	Tim	250	Strat.	Paraplatform	Medium	Ordovician	L. Carboniferous	M. Carboniferous	Hyp.	Neritic limestone	Paralic coal measures	Weak > medium	1–4	?	Bushinsky 1971; 2nd ed. 1975; Patterson 1967
63	Penhsihu (Liaoning Province) (China)	Tim	100	Strat.	Paraplatform	Medium	Ordovician	Carboniferous	M. Carboniferous	Hyp.	Neritic limestone	Paralic coal measures	Weak	1–3	6	Bushinsky 1971; 2nd ed. 1975; Patterson 1967
64	Sikhote-Aliny Range (USSR)	Kaz	1	Dol.	Miogeosyn. zone	Weak	U. Cambrian	U. Cretaceous (?)	U. Cretaceous–Paleogene	Quest.	Neritic limestone	?	Strong	?	?	Kirikov and Mikhailov 1971
65	Bokson (Eastern Sayan) (USSR)	Med	25	Strat. > lent.	Miogeosyn. zone	Strong	U. Proterozoic	L. Cambrian	L. Cambrian	Hyp.	Reef limestone	Swampy carbonaceous clay	V. weak > weak	4–5	30	Ilina 1958; Orlova 1958; Kirpal' and Tenyakov 1974
66	Chadobetsk Area (USSR)	Kaz	25	Dol. > comp.	Continental plate	Quiet	L. Cambrian	Albian–Cenomanian	Oligocene (?) Quaternary	Hyp.	Neritic limestone, dolostone	Dryland sand, clay	Strong > v. strong	10–80 (2–10)	150 (60)	Krivtsov 1968–1969; Peltek 1971; Slukin 1971
67	Region Along Yenisei (USSR)															
	A. Angara area	Kaz	25	Cont. > dol.	Cons. epi-Baikal. zone	Quiet	Precambrian–Cambrian	Albian–Cenomanian	Quaternary	Hyp.	Neritic limestone, dolostone	Dryland scree	Strong > v. strong	40–100 (5–10)	300 (50)	Krivtsov 1968–1969; Peltek 1971
	B. Ivanovsk area	Kaz	25	Cont. > dol.	Cons. epi-Baikal. zone	Quiet	Proterozoic	Palaeocene	Neogene–Quaternary	Hyp.	Neritic limestone, dolostone	Dryland clay	Strong > v. strong	40–100 (5–10)	150 (60)	Krivtsov 1968–1969; Peltek 1971
	C. Tatarsk area	Kaz	20	Cont. > dol.	Cons. epi-Baikal. zone	Quiet	Proterozoic	Palaeocene	Quaternary	Hyp.	Neritic limestone	Dryland clay, sand	Strong > v. strong	40–120 (5–10)	260 (200)	Krivtsov 1968–1969; Peltek 1971; Kirpal' and Tenyakov 1974
	D. Sulakshinsk area	Kaz	3	Cont. > dol.	Cons. epi-Baikal. zone	Quiet	Proterozoic	Palaeocene	Quaternary	Hyp.	Neritic limestone	Dryland clay	Strong > v. strong	40–100 (2–10)	150 (35)	Krivtsov 1968–1969; Kirpal' and Tenyakov 1974
68	Altay–Sayan Region (USSR)															
	A. Alatau of Kuznetsk, Gornaya Shoriya	Med	5	Lent. > nest	Miogeosyn. zone	Strong + reg. metam.	U. Proterozoic	L. Cambrian	L. Cambrian	Hyp.	Metamorphic limestone, dolostone	Metamorphic limestone, dolostone	?	1–2	7	Mikhailov *et al.* 1973
	B. Berdsk, Obukhovsk	Med	10	Lent. > nest	Miogeosyn. zone	Strong + reg. metam.	Coblenzian	Eifelian	U. Eifelian	Cert.	Reef limestone	Neritic limestone	Strong	1–3	20	Nagorsky 1958; Kharin 1969; Mikhailov *et al.* 1973
	C. Salair, Barzassk	Kaz	10	Lent. > dol.	Cons. epi-Hercyn. zone	Quiet	Cambrian	Aptian–Albian	U. Cretaceous	Hyp.	Neritic limestone	Dryland sandy clay	Medium > strong	10–20 (2–5)	50 (35)	Krivtsov 1968–1969; Sapozhnikov 1971
69	Soviet Central Asia															
	A. Bukantau, Tamdytau, Nuratau	Med	5	Lent. > nest	Miogeosyn. zone	Strong + reg. metam.	L. Carbon.–U. Devonian	L. and M. Carboniferous	U. Carboniferous (Bashkirian)	Cert.	Neritic limestone	Neritic calcareous marl	Medium > weak	2–3	40	Abdullaev 1970; Konnov 1972
	B. Turkestan and Aliysk Ranges	Med	5	Lent. > nest	Miogeosyn. zone	Strong + reg. metam.	U. Devonian–L. Carbon.	L. and M. Carboniferous	Bashkirian–Moscovian	Hyp.	Neritic limestone	Neritic limestone	Medium > weak	1–2	60	Abdullaev 1970; Konnov 1972
	C. Southern Ferghana	Med > Kaz	3	Dol.	Cons. epi-Hercyn. zone	Weak	L. and M. Devonian	Norian–Rhaetic	Rhaetic–L. Liassic	Hyp.	Neritic limestone	River-laid conglomerate	Strong	1–3	8	Krivtsov 1968–1969
	D. Chimkent area (Liassic)	Tim	2	Lent.	Cons. epi-Hercyn. zone	Weak	L. Carboniferous	L. Liassic	Liassic	Hyp.	Neritic limestone	Swampy carbonaceous clay	Weak	1–4	5	Mikhailov *et al.* 1973
	E. Northern Ferghana	Med > Kaz	3	Lent. > nest	Partly cons. epi-Hercynian zone	Strong	L. Carboniferous	L. Liassic	Liassic	Hyp.	Neritic limestone	Swampy carbonaceous clay	Medium	3–5	12	Krivtsov 1968–1969
	F. Kansaysk area	Med	1	Lent.	Cons. epi-Hercyn. zone	Medium	Visean	L. Cretaceous	L. Cretaceous	Hyp.	Neritic limestone	Dryland sandy clay	Weak	1–2	5	Krivtsov 1968–1969
	G. Chimkent area (Valanginian)	Kaz	5	Dol. > lent.	Cons. epi-Hercyn. zone	Quiet	L. Carboniferous	Valanginian	L. Cretaceous	Hyp.	Neritic limestone with tuffaceous interbeddings	Lacustrine limestone, clay marl	Medium > weak	1–4	5	Mikhailov *et al.* 1973
70	Central Kazahstan (USSR)															
	A. Tselinograd area	Med	1	Nest	Geosyn. zone	Strong	U. Ordovician	U. Ordovician	U. Ordovician	Hyp.	Reef limestone	Dryland claystone, aleurolite	Medium	0.1–1	1	Dolgopolov *et al.* 1971
	B. Tselinograd, Ekibastuz	Kaz	10	Dol. > cont.	Cons. epi-Caled. zone	Quiet	U. Proterozoic–U. Ordovician	Cenomanian–Palaeocene	M. and U. Oligocene	Hyp.	Neritic limestone	Dryland clay, sandstone	Strong > v. strong	20–140 (2–50)	250 (66)	Krivtsov 1968–1969; Lisitsyna and Pastukhova 1963; Tyurin 1971
	C. Arkalyk	Kaz	180	Strip > dol.	Cons. epi-Caled. zone	Quiet	U. Denovian–L. Carboniferous	Palaeocene–L. Eocene	Eocene	Hyp.	Neritic limestone, dolostone	River-laid sandy clay	Medium > strong	20–60 (5–10)	180 (65)	Krivtsov 1968–1969
71	Turgay Plateau (USSR)															
	A. Naurzumsk	Tim > Tikh	10	Strat. > lent.	Cons. epi-Hercyn. zone	Quiet	L. Carbon.–L. M. Triassic	Albian–Cenomanian	Maestrichtian	Hyp.	Neritic limestone with tuffaceous interbeddings	Neritic glauconitic sand	Weak > v. weak	20–40 (3–5)	70 (35)	Krivtsov 1968–1969; Kirpal' and Tenyakov 1974
	B. Western Turgay area	Kaz	250	Dol. > cont.	Cons. epi-Hercyn. zone	Quiet	L. Carboniferous	Albian–Cenomanian–Turonian	Cenomanian–Palaeocene	Hyp.	Neritic limestone with tuffaceous interbeddings	Swampy carbonaceous clay	Strong > medium	100–150 (5–10)	300 (130)	Krivtsov 1968–1969; Kirpal' and Tenyakov 1974; Tyurin 1971; Venkov and Fatkhudinov 1966
	C. Mugojar area	Kaz	10	Strip > dol.	Cons. epi-Hercyn. zone	Quiet	L. and M. Carboniferous	Aptian–Albian	U. Cretaceous–Palaeocene	Hyp.	Neritic limestone, dolostone	Freshwater clayey sand	Strong > medium	30–60 (5–15)	120 (45)	Krivtsov 1968–1969; Kirpal' and Tenyakov 1974; Tyurin 1971
72	Southern Urals (USSR)															
	A. Novo-Pristansk	Med	50	Lent.	Miogeosyn. zone	Medium	L. Frasnian	M. U. Frasnian	U. Frasnian	Cert.	Reef limestone	Neritic marl	Strong	1–3	10	Kirpal' and Tenyakov 1974
	B. Orsk, Magnitogorsk	Kaz	25	Dol. > lent.	Cons. epi-Hercyn. zone	Quiet	L. Carboniferous	Aptian–Albian	Neogene	Hyp.	Neritic limestone	Dryland multicoloured clay	Strong	20–50 (2–5)	80 (30)	Krivtsov 1968–1969
	C. Chelyabinsk	Kaz	5	Dol. > lent.	Cons. epi-Hercyn. zone	Quiet	Devonian–Carboniferous	Aptian–Albian	Senonian	Hyp.	Neritic limestone	Neritic glauconitic sandstone	Medium	10–20 (2–5)	87 (10)	Krivtsov 1968–1969
73	Central Urals (USSR)															
	A. Nizhne-Serginsk, Pashiya	Med	25	Lent. > nest	Miogeosyn. zone	Strong	L. Eifelian	U. Eifelian	Givetian	Cert.	Neritic limestone	Neritic limestone and dolostone	Weak	1–2	5	Kirpal' and Tenyakov 1974; Mikhailov *et al.* 1973
	B. Kosikovsk	Med	10	Lent. > nest	Eugeosyn. zone	Strong	L. Eifelian	U. Eifelian	Givetian	Cert.	Neritic limestone	Neritic limestone	Weak	1–2	5	Kirpal' and Tenyakov 1974; Mikhailov *et al.* 1973
	C. Kamensk-Uralsk	Kaz	5	Dol.	Cons. epi-Hercyn. zone	Weak	L. and M. Carboniferous	Aptian–Albian	Albian–Cenomanian	Hyp.	Neritic limestone	Dryland clay, sand	Strong	10–40 (1–6)	60 (20)	Krivtsov 1968–1969; Tkachenko and Bolshun 1971
	D. Alapaevsk, Mugaysk	Kaz	10	Dol. > vall.	Cons. epi-Hercyn. zone	Quiet	Devonian–Visean	Aptian–Albian	Albian–Cenomanian–U. Cretaceous	Hyp.	Neritic limestone	Swampy carbonaceous clay	Strong > medium	40–80	150	Krivtsov 1968–1969; Tkachenko and Bolshun 1971
74	Northern Urals (USSR)															
	A. Subrovsk bauxite horizon	Med	200	Strat. > lent.	Eugeosyn. zone	Strong	Gedinnian	L. Eifelian	L. Eifelian	Cert.	Reef limestone	Neritic bituminous shale, limestone	Medium > strong	2–5	60	Gutkin 1978; Bushinsky 1971; 2nd ed. 1975; Kirpal' and Tenyakov 1974
	B. Bogoslovsk bauxite horizon	Med	10	Lent.	Eugeosyn. zone	Strong	U. Eifelian	U. Eifelian–L. Givetian	L. Givetian	Cert.	Reef limestone	Neritic bituminous limestone, marl	Medium > strong	?	?	Gutkin 1978; Bushinsky 1971, 2nd ed. 1975; Kirpal' and Tenyakov 1974
	C. Pripolyarny, Polyarny Ural	Med	10	Lent. > dol.	Eugeosyn. zone	Strong	Gedinnian	Eifelian	U. Eifelian	Hyp.	Reef limestone	Neritic limestone	Medium > strong	1–2	15	Mikhailov *et al.* 1973
	D. Bogoslovsk, Volchansk	Tim	10	Lent.	Partly cons. epi-Hercyn. zone	Medium	Eifelian–Givetian	Rhaetic	Liassic	Hyp.	Neritic limestone	Swampy carbonaceous clay	Medium > weak	10–15	25	Krivtsov 1968–1969; Tkachenko and Bolshun 1971
75	Taimyr Peninsula (USSR)	Tim	10	Strat.	Geosyn. zone	Medium	Famennian	L. Visean	Visean	Hyp.	Neritic limestone	Lagoon limestone	V. weak	5–8	10	Budnikov 1971; Mikhailov *et al.* 1973
76	Timan Range (USSR)															
	A. Central Timan	Tim > Ar	100	Lent.	Cons. epi-Baikal. zone	Weak	U. Rifean	L. Frasnian, M. Devonian	L. Frasnian	Cert.	Neritic limestone, dolostone	Swampy carbonaceous clay, basalt	Medium	5–10	30	Dyemina 1977; Kirpal' 1977
	B. Tobyssk, Kedvinsk	Tim	40	Lent.	Cons. epi-Baikal. zone	V. weak	L. Famennian	Visean	Visean	Cert.	Neritic limestone	Freshwater sandy clay	Weak	1–3	6	Dyemina 1971
	C. Southern Timan	Tim	40	Strat. > lent.	Cons. epi-Baikal. zone	V. weak	L. Frasnian	L. Carboniferous	Visean	Cert.	Neritic limestone	Swampy carbonaceous clay	V. weak > weak	2–4	20	Dyemina 1971, 1977; Gulyanitsky 1969
77	Tulsk, Ryazansk (USSR)	Tul	1	Nest	Continental plate	Quiet	L. Carboniferous	Pliocene	L. Carboniferous	Quest.	Neritic limestone	Swampy coal complex	Medium	0.1–0.2	8	Trubina 1958
78	Puerto Rico	Med	2	Dol. > nest	Eugeosyn. zone	Weak	M. Oligocene–M. Miocene	Miocene–Pliocene	None	Quest.	Reef limestone	None	Strong	?	?	Hildebrand 1960; de Weisse 1976

Appendix 2. Mineralogical data of the world's karst bauxite deposit groups

Serial number	Bauxite region Bauxite deposit group	Average percentage distribution of bauxite types												Percentage abundance of bauxite colours								Bauxite textures (in decreasing abundance)	Stratification of deposits	Fracture	Hardness	Packing
		bauxite	low-iron bauxite	ironrich bauxite	clayey bauxite	calcareous dolomitic bauxite	bauxitic limestone, dolostone	bauxitic limestone, dolostone breccia	bauxite iron ore	clayey bauxitic iron ore	bauxitic clay	carbonaceous bauxitic clay	bauxite like clay	reds violet, pink	yellow yellowish brown	multicoloured	white	dark brown	green	grey	black			of bauxite types		
1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27
1	Andorinha (Portugal)	–	–	–	10	–	–	–	–	–	50	–	40	50	50	–	–	–	–	–	–	pel > pel–pis	0	C	h_1	mp
2	Betic Cordillera (Spain)	10	10	–	30	–	–	–	–	–	30	–	20	60	10	10	20	–	–	–	–	pis > pel > pi–oo–plt	0	R > C > Rh	$h_2 > h_1 > U$	mp > p
3	Catalonia (Spain)																									
	A. Maestrazgo	10	5	–	25	–	–	–	–	–	40	–	20	70	–	25	5	–	–	–	–	plt–pis–mcl > con–ar–mcl	0	R > RC	$h_2 > h_1$	mp > p
	B. Ebro Massif	10	–	–	20	–	–	–	–	–	30	–	40	90	–	10	–	–	–	–	–	con–br–oo	0	R	h_2	mp
4	Spanish Pyrenees (Spain)	15	–	5	40	–	–	–	–	–	30	–	10	90	–	10	–	–	–	–	–	con–oo > cor–ar > pel	0	R > S	$h_2 > h_3$	mp
5	French Pyrenees (France)																									
	A. Ariège	20	5	5	20	–	–	–	–	–	40	–	10	70	–	10	–	–	5	15	–	pis > pis–plt > pel	0	R > L > C	$h_2 > h_3 > h_4$	mp > p–vac
	B. Saint Paul de Fenouillet	10	10	10	30	–	–	–	–	–	30	–	10	60	–	20	10	–	10	–	–	pis > pis–pel > pel	0	R > L > C	$h_1 > h_2 > h_3$	mp > p > vac
6	Languedoc (France)																									
	A. Cazouls, St Chinian	25	5	10	20	–	–	–	–	–	30	–	10	60	–	35	5	–	–	–	–	pel–pis > pis–plt > pel	0 ≫ well	R > L > C	$h_2 > h_3 > h_1$	mp > p > vac
	B. Bédarieux	40	5	10	10	–	–	–	–	–	20	–	15	90	–	5	5	–	–	–	–	mcl–pel > ar > con–br > oo–pis	0 > v > gb	R > L	$h_2 > h_1 > h_3$	mp > p
	C. Villeveyrac	45	5	10	20	–	–	–	–	–	15	–	5	60	10	30	–	–	–	–	–	con–pis > pis–pel–pel > oo	0 > v > gb	R	$h_2 > h_1 > h_3$	mp > p
	D. La Boissière	20	5	10	20	25	5	–	–	–	10	–	5	70	–	30	–	–	–	–	–	oo > oo–pis > pel	0 > v	R	$h_2 > h_3$	mp
7	Provence																									
	A. Alpilles	30	10	10	25	5	–	–	–	–	10	–	10	50	10	40	–	–	–	–	–	con–br > ag > orb–ar > pel	0 > v	R	$h_2 > h_1 > U$	mp > p > vac
	B. Revest and Allauch	55	5	10	15	–	–	–	–	–	10	–	5	60	10	30	–	–	–	–	–	pis–orb > cor–ar > pel	0 > v > well	R	$h_3 > h_2$	mp > p
	C. Brignoles	50	10	10	10	3	–	–	–	–	10	2	5	50	10	25	5	–	–	10	–	con–br > oo–pis–orb–ar > pel	0 > v > gb	R > S > C	$h_2 > h_3 > h_1$	mp > p
	D. Haut-Var	40	5	10	20	–	–	–	–	–	15	–	10	90	–	10	–	–	–	–	–	oo > oo–pis > pel	0 > v	R > C	$h_2 > h_1 > U$	mp > p
8	Vanoise (France)	–	–	–	–	30	–	–	–	–	40	–	30	80	–	–	–	–	–	20	–	mic > pan > pan–oo	0	R	h_4	mp
9	Dréveneuse (Switzerland)	–	–	–	50	–	–	–	–	–	40	–	10	60	30	–	–	–	–	10	–	mic–pel	0	R	$h_4 > h_3$	mp
10	Sardinia-Nurra (Italy)	10	5	–	35	–	–	–	–	–	30	–	20	45	50	–	5	–	–	–	–	con–orb > pe–mcl > oo–ar	0 ≫ well	R > E	$h_1 > U > h_2$	mp > p
11	Central Italy																									
	A. Abruzzi	30	2	8	30	–	–	10	–	–	15	–	5	50	30	15	–	–	–	5	–	ar > ar–oo > con–br–pel–ar	0	R	$h_3 > h_2$	mp
	B. Campania	30	2	5	30	3	–	10	–	–	15	–	5	74	10	15	–	–	1	–	–	oo–ar > oo–orb–ar–con–br	0	R	$h_3 > h_2$	mp
12	Apulia (Italy)																									
	A. San Giovanni Rotondo	30	2	10	33	10	–	–	–	–	10	–	5	60	10	25	–	–	–	5	–	ar > orb–ar > orb–oo–con–orb	0 > v	R	$h_2 > h_1 > h_3$	mp > p
	B. Spinazzola	40	1	10	30	4	–	–	–	–	10	–	5	90	10	–	–	–	–	–	–	ar > con–br > orb > con–br–pel	0	R	$h_2 > h_3 > h_1$	mp > p
13	Terra D' Otranto (Italy)	10	–	10	20	–	–	–	–	–	20	–	40	80	20	–	–	–	–	–	–	orb–con > pe	0	E > R	$h_1 > U > h_3$	p
14	East Alps (Austria)																									
	A. Tirol, Salzburg	20	–	10	40	–	–	–	–	–	20	–	10	80	–	–	–	–	–	20	–	pel > oo–pis	0 > v	S > C	$h_2 > h_3$	mp > p
	B. Unterlaussa, Dreistätten	20	–	10	20	10	2	3	–	5	20	–	10	75	–	10	–	–	–	15	–	oo–pis > pel > con–orb	0 > v	R > C > Rh	$h_3 > h_2 > h_1$	mp
15	Mojtin (Czechoslovakia)	20	5	5	30	–	–	–	–	–	30	–	10	60	10	25	5	–	–	–	–	mcl > con–br > orb–mcl	0	S > R > C	$h_3 > h_2$	mp
16	Markušovce, Drienovec (Czechoslovakia)	20	–	10	30	–	–	–	–	–	30	–	10	70	10	20	–	–	–	–	–	mcl > ar > co	v > well	C > R	$h_2 > h_3 > h_1$	mp
17	Rahov Massif (USSR)	20	–	20	20	–	–	–	–	–	30	–	10	60	10	–	–	–	15	15	–	mcl > pel–oo	v	C > R	$h_4 > h_3 > h_2$	mp
18	Pădurea Craiului (Rumania)	30	5	15	30	–	–	–	–	–	15	–	5	65	10	–	5	5	10	5	–	orb–oo > orb–ar–pel–orb	0	R > Rh > S	$h_4 > h_3 > h_2$	mp > vac
19	Southern Transylvania (Rumania)																									
	A. Sohodol, Cîmpeni	20	–	20	20	–	–	–	–	–	20	10	10	60	–	–	–	–	20	20	–	mcl > ar–mcl > con–ar	0 > v	R	$h_3 > h_2$	mp
	B. Hateg	20	–	20	15	–	–	–	5	–	30	–	10	100	–	–	–	–	–	–	–	pel–mcl > ar–mcl	0 > v	R	$h_3 > h_2$	mp
20	Transdanubian Hills																									
	A. Alsópere, Tés	10	5	5	40	5	–	–	–	–	30	–	5	80	5	10	–	–	–	5	–	mcl–ar > oo–pis–orb–con	0	R > C	$h_2 > h_3$	mp > p–vac
	B. Halimba, Padragkút, Sümeg	25	–	10	20	15	5	–	–	–	20	1	4	80	–	10	–	–	–	10	–	mcl > mcl–or > oo > con	0 > v > gb	R > C	$h_2 > h_1 > h_3$	mp > p
	C. Iharkút, Bakonyjákó, Bakonybél	50	–	–	30	–	–	5	–	–	10	–	5	80	10	10	–	–	–	–	–	oo–orb–mcl > pel–oo–orb–oo–pis	0	R	$h_1 > h_2$	p > mp
	D. Nyírád, Nagytárkány	50	5	10	20	–	–	–	–	–	10	1	4	55	10	30	–	–	–	5	–	pel > mcl > co–n–br (boul)	0	R > C > E	$h_1 > h_2 > U$	p > mp
	E. Szőc Basin, Halimba–Malomvölgy	25	5	10	35	–	–	–	–	–	20	–	5	50	15	30	–	–	–	5	–	pel > mcl–orb > con (boul)	0 ≫ v	R > E > C	$h_1 > U > h_2$	p > mp
	F. Kislőd, Ajka	30	2	10	33	–	–	–	–	–	20	–	5	35	12	50	–	–	–	3	–	pel–mcl > cor–orb > ar	0	R > E	$h_1 > U$	p > vac
	G. Fenyőfő, Dudar, Bakonyoszlop	40	5	5	25	–	–	–	–	–	20	–	5	70	–	30	–	–	–	–	–	mcl > mcl–ar > con–pel	0	R > E	$h_1 > h_2 > U$	p > vac
	H. Iszkaszentgyörgy	50	2	3	30	–	–	–	–	–	10	1	4	25	30	30	–	–	–	15	–	mcl > mcl–ar > con–mcl	0	R > E	$h_1 > h_2 > U$	p > vac
	I. Gánt	40	10	5	30	–	–	–	–	–	10	1	4	20	40	30	5	–	3	2	–	con–orb–mcl > ag–con > pel	0 > v > gb	R > S > E	$h_1 > h_2 > U$	p > vac > mp
	J. Nagyegyháza, Mány	30	–	20	25	10	–	–	–	–	10	–	5	70	–	10	–	–	–	20	–	mcl–ar > ar–con–mcl–pel	0 ≫ v	R	$h_2 > h_1 > U$	p > mp
	K. Nézsa	15	10	5	30	–	–	–	–	–	30	–	10	60	15	20	5	–	–	–	–	con > con–br > ar	0 ≫ gb	R	$h_2 > h_4 > h_2$	mp > p > U
	L. Óbarok, Újbarok	20	–	5	25	–	–	–	–	–	30	–	20	60	10	30	–	–	–	–	–	br–con > ar–mcl	0	R > E	$h_1 > U > h_2$	p > mp
	M. Nagyvázsony, Öcs	10	–	–	30	–	–	–	–	–	30	–	30	80	20	–	–	–	–	–	–	con–orb > pe > pel–pis	0	E > R	$h_1 > U > h_2$	p > mp
21	Nagyharsány	20	20	2	28	–	–	–	–	–	25	–	5	50	10	30	5	–	5	–	–	ar–oo > mcl–pel > ar	0	R > S > Rh	$h_3 > h_2 > h_4$	mp
22	Slovenia (Yugoslavia)																									
	A. Podlipa, Borovnica	30	1	10	40	–	–	–	–	2	12	–	5	75	–	–	–	5	10	10	–	oo > oo–pis > pel–oo > pel	v > 0	S > C	$h_4 > h_3$	mp
	B. Hrusica Mountains, Zuzemberk	40	5	5	40	–	–	–	–	–	10	–	–	90	–	–	–	–	5	5	–	pel > pel–mcl > mcl–orb	0	C > S	$h_3 > h_4$	mp
	C. Vinica, Metlika	35	–	5	30	–	–	–	–	–	20	–	10	100	–	–	–	–	–	–	–	con–pel > pel	0 (well)	C > R	$h_1 > h_2$	p
	D. Nadrt	20	40	–	35	–	–	–	–	–	5	–	–	–	100	–	–	–	–	–	–	oo > oo	0	R > S	$h_4 > h_3$	mp > vug
	D. Nadrt	20	40	–	35	–	–	–	–	–	5	–	–	10	10	–	–	–	40	30	10	pel–oo > oo	0	R > Rh	$h_4 > h_3$	mp > vac
	E. Kozina	50	20	–	25	–	–	–	–	–	5	–	–	70	10	20	–	–	–	–	–	pel > pel–orb	0	C > R	$h_3 > h_2$	mp > p
	F. Savinja Valley	30	5	5	40	–	–	–	–	–	20	–	–	10	60	–	–	–	20	10	–	con–orb > co–n > con–pel	0	R	$h_3 > h_2 > h_4$	p > vac
	G. Velika planina	40	10	–	30	–	–	–	–	–	20	–	–	40	20	30	10	–	–	–	–	pel–col > pel	0	R > C	$h_3 > h_4$	mp > p
	H. Bohinj	30	–	20	30	–	–	–	–	–	20	–	–													
23	Istria (Yugoslavia)													70	–	30	–	–	–	–	–	ar–mcl > ar–pel	0	C > R	h_2	mp > p
	A. Rovinj, Vrsar	20	–	5	50	–	–	–	–	–	25	–	–	50	20	20	–	–	–	10	–	oo–pel > pel	0	R > E	$h_1 > h_2 > h_3$	mp > p
	B. Central and Southern Istria	45	5	20	25	–	–	–	–	–	5	–	–													
24	Dalmatia, Bosnia, Hercegovina																									
	A. Lika, Bjelaj-Lipa	25	–	10	15	–	–	–	–	5	25	–	20	65	–	20	–	5	5	5	–	oo > oo–ar > ppor > pan	0 > v (sch)	R > S > C	$h_3 > h_4$	mp
	B. Knin, Marici	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?
	Grmeć Mountains, Bosanska Krupa (lower Cretaceous)	30	–	10	40	–	–	–	–	–	20	–	–	80	–	–	–	–	20	–	–	pis–oo > pel–mcl	0	R	$h_3 > h_4$	mp
	D. Ogorje	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?
	E. Grmeć Mountains, Bosanska Krupa (Senonian)	60	10	10	15	–	–	–	–	–	5	–	–	40	30	20	–	–	–	10	–	oo > oo–mcl–ar	0	C > Rh > R	$h_4 > h_3$	mp
	F. Jajce	75	5	5	10	–	–	–	–	–	5	–	–	70	10	20	–	–	–	–	–	oo > oo–ar–con > br–con	0	Rh > R	$h_3 > h_4$	mp
	G. Obrovac, Novigrad, Drnis (palaeocene)	60	5	15	15	–	–	–	–	–	5	–	–	70	20	10	–	–	–	–	–	pel > pel–oo	0	R	$h_3 > h_2$	mp
	H. Imotski, Mostar (Palaeocene)	70	5	10	10	–	–	–	–	–	5	–	–	70	20	10	–	–	–	–	–	con > orb > pel	0	R	$h_2 > h_1$	mp > p
	I. Obrovac, Novigrad, Drnis (Eocene)	75	5	5	10	–	–	–	–	–	5	–	–	70	10	20	–	–	–	–	–	pel > pel–orb > pel–ar > pis	0 > v	R > C	$h_2 > h_1 > h_3$	mp > p
	J. Imotski, Mostar (Eocene)	75	2	5	15	–	–	–	–	–	3	–	–	60	10	30	–	–	–	–	–	pel–con > pe	v > 0	R	$h_2 > h_1$	mp > p
	K. Jajce-Baraći	50	–	5	30	2	–	–	–	–	10	–	3	60	–	40	–	–	–	–	–	con > con–pel	0	C > R	$h_2 > h_3 > h_1$	p > mp
25	Crna Gora (Yugoslavia)																									
	A. Gornje Polje, Seljoni	50	15	5	20	–	–	–	–	–	10	–	–	75	–	–	15	–	–	10	–	oo–pis > pel	0 > v	R > Rh	$h_3 > h_4$	mp
	B. Viduša Mountains	60	5	10	20	–	–	–	–	–	5	–	–	80	20	–	–	–	–	–	–	oo–pis > pel	0	R	$h_3 > h_4$	mp
	C. Niksicka Zupa	63	2	10	20	–	–	–	–	–	5	–	–	80	5	10	–	–	–	5	–	mcl–pel > ar–con > ar–orb	0	C > Rh > R	$h_3 > h_2$	mp > p
	D. Dragalj	40	5	10	30	–	–	–	–	–	15	–	–	70	10	15	–	–	–	5	–	?	0	R	h_3	mp
	E. Stara Crna Gora (Ljesev stup)	40	5	10	30	–	–	–	–	–	15	–	–	70	10	20	–	–	–	–	–	?	0	R	h_3	mp
	F. Petrovići, Velimlje, Crvena Kita	30	10	10	10	10	–	20	–	–	10	–	–	70	15	10	5	–	–	–	–	pel–orb > co–n–orb > br > ag	0	C > Rh > R	$h_3 > h_2$	mp > vac
	G. Bijele Poljane	15	10	–	55	–	–	–	–	–	20	–	–	30	–	20	40	–	–	10	–	pel–orb > pe–con	0	C > Rh > R	$h_3 > h_2$	mp > s
	H. Coastal zone (Ulcinj)	60	–	5	25	–	–	–	–	–	10	–	–	20	80	–	–	–	–	–	–	con–nod > o–o–pis–nod	0	R	$h_3 > h_2$	mp > p
26	Vlasenica (Yugoslavia)	40	–	10	35	–	–	–	–	–	10	–	5	80	–	20	–	–	–	–	–	con > pel > orb–ar–oo–pis	0	R > E	$h_1 > U > h_2$	p
27	Grebnička Planina (Yugoslavia)	20	–	45	20	–	–	–	5	5	5	–	–	90	–	–	–	10	–	–	–	pis–oo > pel	0	R > C	$h_3 > h_4$	mp
28	Tran (Bulgaria)	30	5	15	25	–	–	–	–	5	15	–	5	75	20	5	–	–	–	–	–	oo > oo–pis > pel	0	R > C	$h_3 > h_2 > h_1$	mp
29	Albania																									
	A. Albanian Alps	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	pel–oo–pis	?	?	?	?
	B. Coastal Range	15	10	5	30	–	–	–	–	–	30	–	10	40	25	15	–	–	–	10	–	con > pel–co–n > orb–ar–pis	0 ≫ v	R > S > C	$h_2 > h_1 > h_3$	mp > p
	C. Gramsh	5	5	–	40	–	–	–	–	–	40	–	10	40	30	30	–	–	–	–	–	pis > pis–oo	0	R	$h_2 > h_3$	mp > p
30	Central Greece																									
	A. Atalanti, Kallidromon, Helicon Range	30	–	20	40	–	–	–	–	–	10	–	–	50	20	–	–	10	10	10	–	ar–orb–pel–co–n–ar–pel	0 > v	R > S	h_4	mp
	B. Euboia Island, (Lower Malm)	40	10	10	30	–	–	–	–	–	10	–	–	70	20	–	–	–	–	10	–	pis–oo > pis > br	0	R	h_4	mp
	C. Kiona, Parnassos, Helicon Range (Malm)	60	2	13	20	–	–	–	–	–	5	–	–	85	10	–	–	–	2	3	–	pel–mic > or–o–mic	0	Rh > S > C	$h_4 > h_3$	mp
	D. Kiona, Parnassos, Helicon Range (Turanian)	70	5	10	10	–	–	–	–	–	5	–	–	40	30	15	–	–	5	10	–	oo–pis > oo–pis–ar–ar–con	0 > v	R > S > C	$h_4 > h_3$	mp > vac
	E. Euboia Island (Lower Senonian)	40	–	40	15	–	–	–	–	–	5	–	–	80	–	20	–	–	–	–	–	oo–pis > br–con	0	R > S > C	$h_4 > h_3$	mp
31	Pelagonian Zone (Greece)																									
	A. Skopelos Island	15	–	–	10	10	5	–	–	–	40	–	20	55	10	5	–	–	–	30	–	pel > pel–orb > con	0	R > C	$h_4 > h_3$	mp
	B. Volos	50	–	20	20	–	–	–	–	–	10	–	–	100	–	–	–	–	–	–	–	?	0	R	h_4	mp
32	Vardar Zone (Greece)																									
	A. Chalkidike Peninsula	30	–	–	50	–	–	10	–	–	10	–	–	30	–	–	–	50	20	–	–	pan	0	R	h_4	mp
33	Eleusis, Mandra, Megara (Greece)	40	–	30	15	–	–	–	–	–	10	–	5	80	–	20	–	–	–	–	–	pel–mic > oc–pis–ar–orb	0 > v	Rh > S > R	$h_4 > h_3 > h_2$	mp
34	Gavrovo Zone (Greece)																									
	A. Peloponnese Peninsula	55	5	10	20	–	–	–	–	–	10	–	–	60	–	–	–	–	20	20	–	pis > pis–plt > ar–orb	0 > v	R	$h_4 > h_3$	mp
	B. Klokova	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	0 > v	?	?	?
35	Aegean Islands																									
	A. Amorgos	50	–	10	30	–	–	–	–	–	10	–	–	95	5	–	–	–	–	–	–	ar–pan > orb > pan	0	R > C	h_4	mp
	B. Samos	45	–	5	30	–	–	–	–	–	20	–	–	10	–	30	–	–	–	60	–	orb–ar–pan > pis–pan	v > sch	Rh > C > R	$h_4 > h_2$	mp
	C. Naxos	40	–	20	30	–	–	–	–	–	10	–	–	–	–	–	–	–	–	70	30	gbl > lbl > pbl	0 > sch	R > C > S	h_5	mp
36	Menderes Massif (Turkey)	30	–	10	40	–	–	–	–	–	20	–	–	20	–	10	–	20	–	40	10	pis–orb–spa > gbl > lbl–pbl	0 > sch > v	R > C > S	$h_5 > h_4$	mp
37	Western Taurus (Turkey)																									
	A. Alanya	30	10	10	20	–	–	–	–	–	10	–	–	20	–	10	–	20	–	40	10	con–ar > oo > pis	0 > v	R > C > S	$h_4 > h_5$	mp
	B. Akseki, Seydisehir, Sütlegen	50	–	10	20	–	–	10	–	–	10	–	–	60	10	10	–	10	–	10	–	mic–pel > ar–orb > oo–orb	0 > v > gb	Rh > C > R	$h_3 > h_2$	mp
	C. Sütlegen	10	–	–	40	–	–	–	–	–	40	–	10	?	?	?	?	?	?	?	?	?	?	?	?	?
38	Eastern Taurus (Turkey)																									
	A. Adama, Bolkardag, Saimbeyli	20	10	10	20	–	–	–	5	5	30	–	–	20	–	10	–	20	–	40	10	con–ar > oo > pis > gbl	0 > v > sch	R > C > S	$h_4 > h_5$	mp > vac–vug
	B. Kücük Koras, Sebilköy	30	10	10	30	–	–	–	–	–	10	–	–	70	–	30	–	–	–	–	–	ar–orb > oo–pis > mic–pan	0	Rh > C > R	$h_3 > h_4$	mp
39	Payas, Islahye (Turkey)	10	–	10	10	–	–	–	20	40	10	–	–	100	–	–	–	–	–	–	–	ar > ar–con > mcl	0	C > R	$h_4 > h_3$	mp
40	Zonguldak (Turkey)	20	–	20	10	–	–	–	–	–	30	–	20	100	–	–	–	–	–	–	–	pel–pis	0	R > E	U > h_3	p > mp
41	Crimea (USSR)	25	–	5	35	5	–	–	–	–	30	–	–	100	–	–	–	–	–	–	–	pis–plt > pis–plt–orb	0	R	h_3	mp
42	Maktesh Ramon (Israel)	–	10	–	30	–	–	–	–	–	40	–	20	35	10	40	10	–	5	–	–	oo–plt > oo–pel > pel	0	C > R	$h_2 > h_1$	mp > p
43	Little Caucasus (Julfa)	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	pis	?	?	?	?
44	Bukan (Iran)	10	5	15	5	5	–	–	–	10	30	–	20	30	–	10	5	20	25	10	–	mic–mcl > pis > mic > con–nod	v > 0 > well	R	$h_3 > h_4$	mp > p
45	Alborz Range (Iran)																									
	A. Jajarm (Permo-Triassic)	5	–	5	–	–	–	–	–	–	40	–	40	30	–	–	–	50	10	10	–	pis–oo > mi–pel	0	R	$h_3 > h_4$	mp
	B. Jajarm (Upper Triassic)	15	5	10	20	–	–	–	–	–	20	–	30	20	–	–	5	50	5	20	–	pis–oo > mi–pel	0	R	$h_3 > h_4$	mp
46	Central Iranian Plateau																									
	A. Yazd	10	–	40	20	–	–	–	–	–	20	–	10	50	–	–	–	30	–	20	–	pis > pis–oo	0 > v	R	$h_3 > h_4$	mp
	B. Kerman	–	–	20	30	–	–	–	10	10	20	–	10	90	–	–	–	10	–	–	–	mic > mic–orb	0	R > S > C	$h_3 > h_4$	mp
47	Zagros Range (Iran)																									
	A. Do-Polan, Do-Rag	5	20	–	25	–	–	–	–	–	40	–	10	20	20	–	10	–	–	50	–	con–br > ar–mcl > pis–oo	0 > v	R	$h_4 > h_3$	mp
	B. Behbahan (Sar-e-Fariab)	20	10	–	30	5	–	–	–	–	30	–	5	60	10	–	–	20	–	10	–	pis–oo > pis–oo–ar	0 > v	R	h_3	mp > vac
	C. Shiraz (Daryush Dam)	25	–	5	40	5	–	–	–	–	20	–	5	80	–	–	–	20	–	–	–	pis–oo > pis–mcl > pis–ar	0	R	$h_4 > h_3$	mp

Appendix (cont.)

Serial number	Bauxite region Bauxite deposit group	Average percentage distribution of bauxite types												Percentage abundance of bauxite colours								Bauxite textures (in decreasing abundance)	Stratification of deposits	Fracture	Hardness	Packing
		bauxite	low-iron bauxite	ironrich bauxite	clayey bauxite	calcareous dolomitic bauxite	bauxitic limestone, dolostone	bauxitic limestone, dolostone breccia	bauxite iron ore	clayey bauxitic iron ore	bauxitic clay	carbonaceous bauxitic clay	bauxite like clay	reds violet, pink	yellow yellowish brown	multicoloured	white	dark brown	green	grey	black			of bauxite types		
1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27
48	Kandahar	–	–	20	40	–	–	–	–	20	20	–	–	50	–	–	–	10	40	–	–	plt–ar > plt–pis–ar	0	R > C	$h_4 > h_3$	mp
49	Pamirs (USSR)	20	–	10	30	–	–	–	–	10	30	–	–	40	–	–	–	10	20	30	–	plt–ar > plt–pis–ar	v > well	R > C	$h_3 > h_4$	mp
50	Hindukush Range (Pakistan)																									
51	Northeast Pakistan																									
	A. Salt Range, Sargodha	–	15	–	30	5	–	–	–	–	30	–	20	30	30	20	10	–	10	–	–	oo–pis > pel	0	R > E	$h_2 > h_1$	p > mp
	B. Campbellpore, Muzaffarabad	15	5	–	30	–	–	–	–	–	30	10	10	50	10	10	20	–	–	5	5	pis	?	?	?	?
52	Jammu (Kashmir-India)													–	30	–	10	–	–	60	–	pis–oo > pel–oo > pis–plt	0	R	$h_4 > h_3 > h_2$	mp
53	Samar Island (Philippines)																									
54	Solomon Island																									
	A. Wagina and other islands																									
	B. Rennell Islands	90	–	–	10	–	–	–	–	–	–	–	–	100	–	–	–	–	–	–	–	pel > pel–pis	0	E	Us	p
55	Loyalty Group																									
	A. Lifou Island	40	–	20	–	40*	–	–	–	–	–	–	–	100	–	–	–	–	–	–	–	pel	0	E	Us	p
	B. Maré Island	40	–	10	–	50*	–	–	–	–	–	–	–	100	–	–	–	–	–	–	–	pel	0	E	Us	p
56	Fiji-Tonga Group																									
	A. Lau Islands (Fiji)	50	–	–	–	50*	–	–	–	–	–	–	–	100	–	–	–	–	–	–	–	pel	0	E	Us	p
	B. Niué Island (Tonga)	50	–	30	–	20*	–	–	–	–	–	–	–	100	–	–	–	–	–	–	–	pel	0	E	Us	p
57	Sisophon, Battambang (Campuchea)	20	–	–	30	–	–	–	–	10	20	–	20	60	–	40	–	–	–	–	–	pel–mic–mcl > pel–oo	0	R	h_3	mp
58	Socialist Republic of Vietnam																									
	A. Lang-Son (Upper Permian)	20	–	25	20	5	–	–	5	5	20	–	–	10	–	–	–	30	20	35	5	ar–mic > ar–oo–orb–con	v > 0 > well	C > R	$h_4 > h_3 > h_1$	mp
	B. Lang-Son (Neogene)	5	–	25	10	–	–	–	5	5	40	–	10	45	10	10	–	10	5	20	–	ar–mic > ar–oo–orb > pel	0	C > R > E	$h_4 > h_1 > h_3$	mp > p
	C. Cao-Bang (Upper Permian)	20	–	30	20	–	–	–	5	5	20	–	–	10	–	–	–	30	20	35	5	mic > pan > ar–orb > con	v > 0 > well	C > R	$h_4 > h_5 > h_3$	mp
	D. Cao-Bang (Neogene)	5	–	25	10	–	–	–	5	5	40	–	10	40	10	5	–	15	10	20	–	mic > pan > ar–orb > pel	0	R > C > E	$h_4 > h_5 > h_1$	mp > p
59	Kunming (Yünnan Province) (China)	15	5	10	20	–	–	–	5	15	10	10	10	–	–	–	–	–	20	70	10	ar–mcl > oo–pis > pan	v	R	$h_4 > h_3$	mp
60	Kweichow Province (China)																									
	A. Kveyang (Carboniferous)	20	10	10	20	–	–	–	5	5	10	10	10	–	–	–	–	10	10	70	10	mcl–mic > oo–pis > pan	v	R > E	$h_3 > h_1 > h_4$	mp > p
	B. Kveyang (Permian)	10	10	10	25	–	–	–	10	5	10	10	10	–	–	–	–	10	10	70	10	mcl–mic > oo–pis > pan	v	R > E	$h_3 > h_1 > h_4$	mp > p
61	Gun District (Honan Province) (China)	10	20	–	20	–	–	–	10	10	10	10	10	–	–	–	–	10	10	70	10	ar–mcl > ar–con > ar–pan	v > sch	R	$h_4 > h_3$	mp
62	Nanting, Po-San (Shantung Province) (China)	20	5	5	20	–	–	–	5	15	10	10	10	–	–	–	–	–	30	70	–	ar–con > pis > mic–pan	v > well	R	$h_4 > h_3$	mp
63	Penhsihu (Liaoning Province) (China)	20	5	5	20	–	–	–	5	15	10	10	10	–	–	–	–	–	–	100	–	ar–con > pis > mic–pan	v > well	R	$h_4 > h_3$	mp
64	Sikhote-Aliny Range (USSR)	10	–	–	50	–	–	–	–	–	40	–	–	100	–	–	–	–	–	–	–	pis	0	R	h_2	mp
65	Bokson (Eastern Sayan) (USSR)	10	–	10	40	–	–	–	–	5	30	–	5	80	–	–	–	–	10	5	5	br–ar > mcl > mcl–oo	v > well	R > C > S	$h_4 > h_3$	mp
66	Chadobetsk Area (USSR)	10	–	20	20	–	–	–	5	10	20	–	15	75	10	10	5	–	–	–	–	pis–con > pis–orb–ar–con–ar	0	R > E	$h_1 > U > h_2$	p > mp
67	Region Along Yenisei (USSR)																									
	A. Angara area	10	–	20	20	–	–	–	5	5	20	–	20	100	–	–	–	–	–	–	–	pis–orb > pis > orb–ar–ar	v > 0	R > E	$h_1 > h_2 > U$	p > mp
	B. Ivanovsk area	5	–	20	20	–	–	–	10	10	20	–	15	100	–	–	–	–	–	–	–	pis–orb > pis > orb–ar > ar	0	R > E	$h_1 > U > h_2$	p > mp
	C. Tatarsk area	5	–	15	15	–	–	–	15	15	20	5	10	90	–	–	–	–	–	5	5	pis–orb > ar–con–br	0	R > E	$h_1 > U > h_2$	p > mp
	D. Sulakshinsk area	5	–	15	15	–	–	–	15	15	20	–	15	100	–	–	–	–	–	–	–	pis–orb > ar–con	0	R > E	$h_1 > U > h_2$	p > mp
68	Altay Sayan Region (USSR)																									
	A. Alatau of Kuznetsk, Gornaya Shoriya	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?
	B. Berdsk, Obukhovsk	20	–	–	30	10	10	5	–	–	20	5	–	–	–	–	–	–	–	–	–	oo > oo–mcl	well > gb > cs	R	$h_4 > h_5$	mp
	C. Salair, Barzassk	10	–	10	15	–	–	–	5	5	25	5	25	70	5	–	–	–	–	20	5	pis > pis–orb–mcl–pel	0	R > E > C	$h_1 > U > h_2$	p > mp
69	Soviet Central Asia																									
	A. Bukantau, Tamdytau, Nuratau	20		20	25	–	–	5	–	–	25	–	5	40	–	20	–	–	20	20	–	ar–orb > ar–mcl–gbl > pan	0 > well–gb	R	$h_3 > h_4 > h_5$	mp
	B. Turkestan and Aliysk ranges	20	–	20	25	–	–	5	–	–	25	–	5	40	–	20	–	–	20	20	–	ar–mcl > orb–ar > oo–pel–gbl	0 > well–gb	R	$h_3 > h_4 > h_5$	mp
	C. Southern Ferghana	15	5	–	30	–	–	–	–	–	30	–	20	50	–	–	–	–	45	5	–	br > br–ag–ar–orb > oo	0 > v	R	$h_2 > h_1 > h_4$	mp > p
	D. Chimkent area (Liassic)	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?
	E. Northern Ferghana	10	–	10	25	–	–	–	10	5	20	–	20	60	20	–	–	–	20	–	–	orb–pis	0 > v	R	$h_2 > h_3 > h_1$	mp
	F. Kansaysk area	10	–	15	30	–	–	–	–	–	25	–	20	100	–	–	–	–	–	–	–	oo	0	R	h_1	mp > p
	G. Chimkent area (Valanginian)	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?
70	Central Kazakhstan (USSR)																									
	A. Tselinograd area	–	–	–	20	50	10	–	–	–	20	–	–	100	–	–	–	–	–	–	–	ar > mcl	0	R	$h_4 > h_5$	mp
	B. Tselinograd, Ekibastuz	10	–	5	30	–	–	–	5	5	30	–	15	70	10	20	–	–	–	–	–	mcl–oo > mcl–ar > pis–pel	0	E > R	$h_1 > U > h_2$	p > mp
	C. Arkalyk	15	5	5	30	–	–	–	5	5	20	5	10	40	20	20	15	–	–	5	–	mcl–pel > mcl–ar–pis–plt	0 > v	E > R	$h_1 > U > h_2$	p > vac–tv
71	Turgay-Plateau (USSR)																									
	A. Naurzumsk	10	5	5	20	–	–	–	5	5	30	–	20	50	10	20	–	–	–	20	–	mcl–orb > pis–orb	0	R > E	$h_1 > U > h_2$	p > mp
	B. Western Turgay area	15	5	5	25	–	–	–	5	5	20	5	15	60	20	15	–	–	–	–	5	mcl–ar > pis–ar > pis–plt–con	0 > v > gb	R > E	$h_1 > h_3 > h_2$	p > mp–vac
	C. Mugojar area	10	5	5	20	–	–	–	–	–	30	10	20	60	–	30	–	–	–	5	5	mcl–pis > ar–orb > pis–orb	0 > v	E > R	$U > h_1 > h_2$	p > mp
72	Southern Urals (USSR)																									
	A. Novo-Pritansk	20	5	5	30	–	–	5	–	–	20	5	10	65	–	10	–	5	5	10	5	ar–mcl > pis–oo > pis–plt	v > 0 > well	R > C > S	$h_3 > h_4 > h_2$	mp > p–vac
	B. Orsk, Magnitogorsk	15	–	5	30	–	–	–	–	–	30	–	20	75	20	–	–	–	–	5	5	pis–orb > mcl–ar	0 > v	E > R	$U > h_1 > h_2$	p > vac–tv
	C. Chelyabinsk	10	–	10	20	–	–	–	5	5	30	–	20	100	–	–	–	–	–	–	–	mcl–ar > ar–orb	0	E > R	$U > h_1 > h_2$	p
73	Central Urals (USSR)																									
	A. Nizhne-Serginsk, Pashiya	25	5	5	25	10	–	5	–	–	20	–	5	50	–	–	–	10	20	20	–	ar–orb > ar–mcl	v	R	$h_3 > h_4 > h_2$	mp > p
	B. Kosikovsk	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?
	C. Kamensk-Uralsk	5	–	20	30	–	–	–	5	5	20	–	15	100	–	–	–	–	–	–	–	mcl–ar > ar–orb	0	E > R	$h_1 > U > h_2$	p
	D. Alapaevsky, Mugaysk	20	5	5	20	–	–	–	–	–	30	–	20	85	–	–	5	–	–	10	–	mcl–pel > pis–plt	0 > v	E > R	$h_1 > U > h_2$	p > mp
74	Northern Urals (USSR)																									
	A. Subrovsk bauxite horizon	30	5	10	15	5	5	5	3	2	10	5	5	50	10	5	–	5	10	15	5	ar > ar–con > mcl > con–mic	v > well–gb	R > C > S	$h_3 > h_4 > h_2$	mp > vac–s
	B. Bogoslovsk bauxite horizon	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?	?
	C. Pripolyarny, Polyarny Ural	?	?	?	?	?	?	?	?	?	?	?	?	–	–	–	–	–	40	30	30	?	?	?	?	?
	D. Bogoslovsk, Volchansk	15	5	5	30	–	–	–	–	–	35	–	10	40	–	10	–	–	25	25	–	pis–ar > ar > mcl–ar–con	0 > v	R	$h_2 > h_1 > U > h_3$	mp > p
75	Taimyr Peninsula (USSR)	–	–	–	10	–	20	–	–	10	50	–	10	30	30	–	–	–	30	10	–	pel > mic > oo–pis	v > well	C > R > E	$h_3 > h_2 > h_1$	mp > p
76	Timan Range (USSR)																									
	A. Central Timan																									
	B. Tobyssk, Kedvinsk	5	5	–	40	–	–	–	–	10	20	–	20	80	–	20	–	–	–	–	–	ar > pel > mic	v > 0	R	$h_2 > h_1$	mp > p
	C. Southern Timan	5	20	–	25	5	–	–	–	–	20	5	20	15	10	–	5	–	–	70	–	pel > mic > oo–pis–ar > br	v > 0	R	$h_2 > h_1 > h_3$	mp > p–vac
77	Tulsk, Ryazansk (USSR)	–	20	–	40	–	–	–	–	–	40	–	–	–	10	–	60	–	–	30	–	pel–pan > pel	0	E	$U > h_1$	p > vac
78	Puerto Rico	–	–	–	20	–	–	–	–	–	50	–	30	70	30	–	–	–	–	–	–	ar–pel	0	R > C > E	$h_1 > U$	p
79	Dominican Republic	60	–	5	25	–	–	–	–	–	10	–	–	100	–	–	–	–	–	–	–	pel > pel–pis > pel–oo	0	E > R	$U > h_1$	p > vac
80	Haiti	60	–	5	25	–	–	–	–	–	10	–	–	70	30	–	–	–	–	–	–	pel > pel–pis–oo	0	E > R	$U > h_1$	p
81	Jamaica																									
	A. St Ann Plateau	70	–	–	20	–	–	–	–	–	10	–	–	70	30	–	–	–	–	–	–	pel	0	E > R	$U > h_1$	p
	B. Cockpit Country	80	–	–	15	–	–	–	–	–	5	–	–	100	–	–	–	–	–	–	–	pel	0 ≫ gb	E > R	$U > h_1$	p
	C. Essex Valley	60	5	–	13	–	–	2	–	–	20	–	–	60	30	5	5	–	–	–	–	pel > nod	0	E > R	$h_1 > U$	p
	D. Manchester Plateau	70	–	–	20	–	–	–	–	–	10	–	–	80	20	–	–	–	–	–	–	pel	0	E > R	$U > h_1$	p
	E. Clarendon Plateau	70	–	–	20	–	–	–	–	–	10	–	–	80	20	–	–	–	–	–	–	pel	0	E > R	$U > h_1$	p
	F. St Catherine Plateau	60	–	–	30	–	–	–	–	–	10	–	–	80	20	–	–	–	–	–	–	pel	0	E > R	$U > h_1$	p
82	Cayman Islands	–	–	–	10	–	–	–	–	–	90	–	–	100	–	–	–	–	–	–	–	pel	0	E > R	U	p
83	Bahamas	–	–	–	30	–	–	–	–	–	70	–	–	100	–	–	–	–	–	–	–	pel	0	E	U	p
84	Missouri (Gasconade) (US)	–	20	–	30	–	–	–	–	–	30	–	20	15	–	5	60	–	–	20	–	oo > oo–mic > pel	0 > well > v	R > C	$h_4 > h_3 > h_2$	mp > p–vac
85	Western rim of Coastal Plain (US)	–	5	–	20	–	–	–	–	–	45	–	30	25	35	–	40	–	–	–	–	pis–oo–plt > pel	0	R	$h_1 > U$	p > vac
86	SE rim of Coastal Plain (US)	3	20	2	25	–	–	–	–	–	30	–	20	10	30	–	40	–	–	20	–	pis–oo–plt > pel–pis–pel	0	R	$h_1 > U > h_2$	p > vac > s
87	SW Valley and Ridge Prov. (US)	5	15	3	25	–	–	–	–	–	30	2	20	10	25	15	30	–	–	20	–	pis–oo–plt > pel–con	0 > v	R	$h_1 > h_2 > h_3$	p > vac > s
88	NE Valley and Ridge Prov. (US)	2	10	–	28	–	–	–	–	–	30	–	–	–	30	–	40	–	–	30	–	pis–oo–plt > pel	0	R	$h_1 > h_2 > h_5$	p > vac > s

Explanation of symbols:

To Column 23

ag = agglomerate texture
ar = arenitic texture
plt = pisolithic texture
br = brecciated
orb = orbiculate particles in texture
(boul) = boulders (sporadic)
nod = nodular texture
con = conglomerate texture
col = collomorphic texture
pan = panidiomorphic texture
mic = microgranular texture
mcl = microclastic texture
oo = ooidic texture
pel = pelitomorphic texture
pis = pisoidic texture
ppor = pseudoporphyric texture
spas = spastoidic texture

In metamorphic bauxites:

gbl = granoblastic texture
lbl = lepidoblastic texture
pbl = porphyroblastic texture

To Column 24

0 = unstratified
v = weakly, vaguely stratified
cs = cross-stratification
gb = graded bedding
sch = schistosity, shale structure

To Column 25

R = rough
E = earthy
C = conchoidal
L = laminate
Rh = rhombohedral
S = smooth

The bauxites marked with an asterisk in Column 7 are phosphatic.

To Column 26

Us = unconsolidated, soil-like
U = unconsolidated
h_1 = low-hardness
h_2 = medium-hardness
h_3 = hard
h_4 = very hard
h_5 = corundum-hard

To Column 27

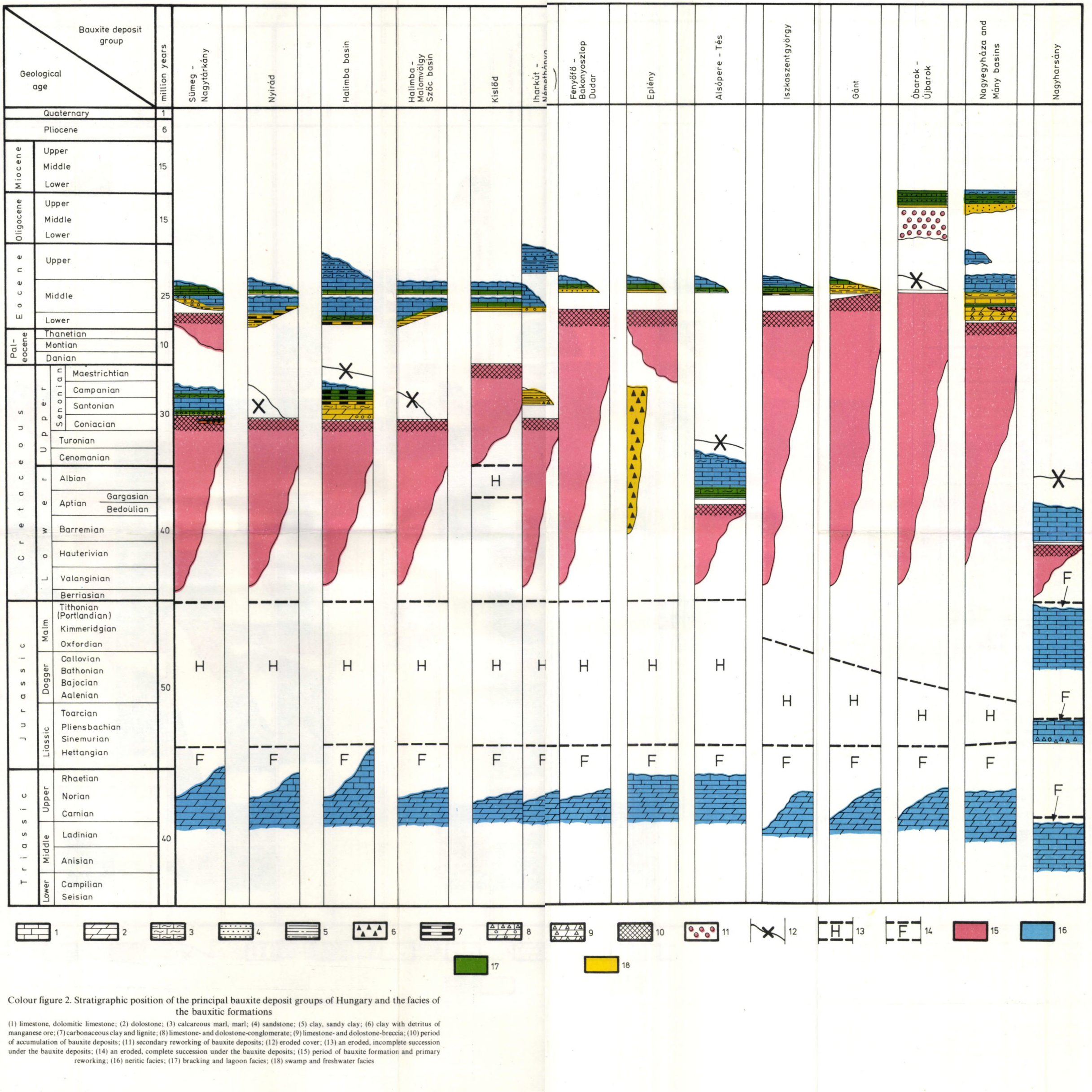
mp = microporous
p = porous
vac = vacuolar
s = spongy
vug = vugular
tv = tubulo-vugular

Appendix 3. Percentage abundances of aluminium mineral association in karst bauxites, by deposit groups

Serial number	Bauxite region Bauxite deposit group	Gibbsitic	Gibbsito Boehmito	Gibbsito–boehmito–diasporic	Gibbsito–corundous	Boehmitic	Boehmito diasporic	Diasporic	Diasporo–corundous	Corundous
1	2	3	4	5	6	7	8	9	10	11
1	Andorinha (Portugal)	–	100	–	–	–	–	–	–	–
2	Betic Cordillera (Spain)	10	80	–	–	10	–	–	–	–
3	Catalonia (Spain)									
	A. Maestrazgo	–	–	80	–	–	–	20	–	–
	B. Ebro Massif	–	–	–	–	100	–	–	–	–
4	Spanish Pyrenees (Spain)	–	–	–	–	100	–	–	–	–
5	French Pyrenees (France)									
	A. Arlège	–	5	–	–	5	60	30	–	–
	B. Saint Paul de Fenouillet	–	–	–	–	–	–	100	–	–
6	Languedoc									
	A. Cazouls, St Chinian	15	70	5	–	10	–	–	–	–
	B. Bédarieux	–	–	–	–	100	–	–	–	–
	C. Villeveyrac	–	–	–	–	100	–	–	–	–
	D. La Boissière	–	100	–	–	–	–	–	–	–
7	Provence (France)									
	A. Alpilles	–	40	–	–	60	–	–	–	–
	B. Revest and Allauch	–	–	–	–	100	–	–	–	–
	C. Brignoles	–	20	–	–	80	–	–	–	–
	D. Haut-Var	10	80	–	–	10	–	–	–	–
8	Vanoise (France)	–	–	–	–	–	–	100	–	–
9	Dréveneuse	–	–	–	–	–	100	–	–	–
10	Sardinia-Nurra (Italy)	–	–	–	–	100	–	–	–	–
11	Central Italy									
	A. Abruzzi	–	–	–	–	100	–	–	–	–
	B. Campania	–	–	–	–	100	–	–	–	–
12	Apulia (Italy)									
	A. San Giovanni Rotondo	–	–	–	–	100	–	–	–	–
	B. Spinazzola	–	–	–	–	100	–	–	–	–
13	Terra D'Otranto (Italy)	–	–	–	–	100	–	–	–	–
14	East Alps (Austria)									
	A. Tirol, Salzburg	–	–	–	–	100	–	–	–	–
	B. Unterlaussa, Dreistätten	–	–	–	–	100	–	–	–	–
15	Mojtin	–	30	–	–	70	–	–	–	–
16	Markušovce, Drienovec (Czechoslovakia)	–	40	–	–	60	–	–	–	–
17	Rahov Massif (USSR)	–	–	–	–	–	–	100	–	–
18	Pădurea Craiului (Rumania)	–	–	–	–	15	15	65	5	–
19	Southern Transylvania (Rumania)									
	A. Sohodol, Cimpeni	–	–	–	–	100	–	–	–	–
	B. Haţeg	–	–	–	–	100	–	–	–	–
20	Transdanubian Hills									
	A. Alsópere, Tés	–	40	–	–	60	–	–	–	–
	B. Halimba, Padragkút, Sümeg	–	10	–	–	90	–	–	–	–
	C. Iharkút, Bakonyjákó, Bakonybél	–	60	–	–	40	–	–	–	–
	D. Nyirád, Nagytárkány	–	30	–	–	70	–	–	–	–
	E. Szőc Basin, Halimba–Malomvölgy	40	60	–	–	–	–	–	–	–
	F. Kislőd, Ajka	10	80	–	–	10	–	–	–	–
	G. Fenyőfő, Dudar, Bakonyoszlop	20	60	–	–	20	–	–	–	–
	H. Iszkaszentgyörgy	10	70	–	–	20	–	–	–	–
	I. Gánt	–	70	–	–	30	–	–	–	–
	J. Nagyegyháza, Mány	–	10	–	–	90	–	–	–	–
	K. Nézsa	–	–	–	–	–	–	100	–	–
	L. Óbarok, Újbarok	30	60	–	–	10	–	–	–	–
	M. Nagyvázsony, Öcs	30	50	–	–	20	–	–	–	–
21	Nagyharsány	–	–	–	–	40	50	10	–	–
22	Slovenia (Yugoslavia)									
	A. Podlipa, Borovnica	–	–	–	–	90	10	–	–	–
	B. Hrušica Mountains, Žužemberk	–	–	–	–	100	–	–	–	–
	C. Vinica, Metlika	100	–	–	–	–	–	–	–	–
	D. Nadrt	–	–	–	–	100	–	–	–	–
	E. Kozina	–	–	–	–	60	30	10	–	–
	F. Savinja Valley	–	–	–	–	50	50	–	–	–
	G. Velika planina	–	–	–	–	–	–	100	–	–
	H. Bohinj	–	100	–	–	–	–	–	–	–
23	Istria (Yugoslavia)									
	A. Rovinj, Vrsar	–	–	–	–	100	–	–	–	–
	B. Central and Southern Istria	20	50	–	–	30	–	–	–	–
24	Dalmatia, Bosnia, Hercegovina									
	A. Lika, Bjelaj-Lipa	–	–	–	–	10	60	30	–	–
	B. Knin, Marići	?	?	?	?	?	?	?	?	?
	C. Grmeć Mountains, Bosanska Krupa (Lower Cretaceous)	–	–	–	–	100	–	–	–	–
	D. Ogorje	?	?	?	?	?	?	?	?	?
	E. Grmeć Mountains, Bosanska Krupa (Senonian)	–	–	–	–	80	20	–	–	–
	F. Jajce	–	–	–	–	100	–	–	–	–
	G. Obrovac, Novigrad, Drnis (Paleocene)	–	100	–	–	–	–	–	–	–
	H. Imotski, Mostar (Paleocene)	–	70	–	–	30	–	–	–	–
	I. Obrovac, Novigrad, Drniš (Eocene)	40	60	–	–	–	–	–	–	–
	J. Imotski, Mostar (Eocene)	10	90	–	–	–	–	–	–	–
	K. Jajce-Baraći	10	90	–	–	–	–	–	–	–
25	Crna Gora (Yugoslavia)									
	A. Gornje Polje, Seljoni	–	–	–	–	100	–	–	–	–
	B. Viduša Mountains	–	–	–	–	100	–	–	–	–
	C. Nikšićka Zupa	–	–	–	–	100	–	–	–	–
	D. Dragalj	–	–	–	–	100	–	–	–	–
	E. Stara Crna Gora (Ljesev stup)	–	–	–	–	100	–	–	–	–
	F. Petrovići, Velimlje, Crvena Kita	–	–	–	–	100	–	–	–	–
	G. Bijele Poljane	–	40	–	–	60	–	–	–	–
	H. Coastal zone (Ulcinj)	–	–	–	–	100	–	–	–	–
26	Vlasenica (Yugoslavia)	–	–	–	–	100	–	–	–	–
27	Grebnička Planina (Yugoslavia)	–	–	–	–	10	30	60	–	–
28	Tran (Bulgaria)	–	10	–	–	90	–	–	–	–
29	Albania									
	A. Albanian Alps	?	?	?	?	?	?	?	?	?
	B. Coastal Range	–	40	–	–	60	–	–	–	–
	C. Gramsh	?	?	?	?	?	?	?	?	?
30	Central Greece									
	A. Atalanti, Kallidromon, Helicon Range	–	–	–	–	30	70	–	–	–
	B. Euboia Island (Lower Malm)	–	–	–	–	–	–	100	–	–
	C. Kiona, Parnassos, Helicon Range (Malm)	–	–	–	–	60	40	–	–	–
	D. Kiona, Parnassos, Helicon Range (Turanian)	–	–	–	–	20	40	40	–	–
	E. Euboia Island (Lower Senonian)	–	–	–	–	–	–	100	–	–
31	Pelagonian Zone (Greece)									
	A. Skopelos Island	–	–	–	–	–	–	100	–	–
	B. Volos	?	?	?	?	?	?	?	?	?
32	Vardar Zone (Greece)									
	A. Chalkidike Peninsula	–	–	–	–	–	–	100	–	–
33	Eleusis, Mandra, Megara (Greece)	–	–	–	–	60	35	5	–	–
34	Gavrovo Zone (Greece)									
	A. Peloponnese Peninsula	–	30	30	–	20	–	20	–	–
	B. Klokova	–	–	–	–	–	10	90	–	–
35	Aegean Islands									
	A. Amorgos	–	–	–	–	–	–	100	–	–
	B. Samos	–	–	–	–	–	–	100	–	–
	C. Naxos	–	–	–	–	–	–	–	–	100
36	Menderes Massif (Turkey)	–	–	–	–	–	–	40	30	30
37	Western Taurus (Turkey)									
	A. Alanya	–	–	–	–	–	–	80	20	–
	B. Akseki, Seydisehir, Sütlegen	–	–	–	–	90	10	–	–	–
	C. Sütlegen	?	?	?	?	?	?	?	?	?

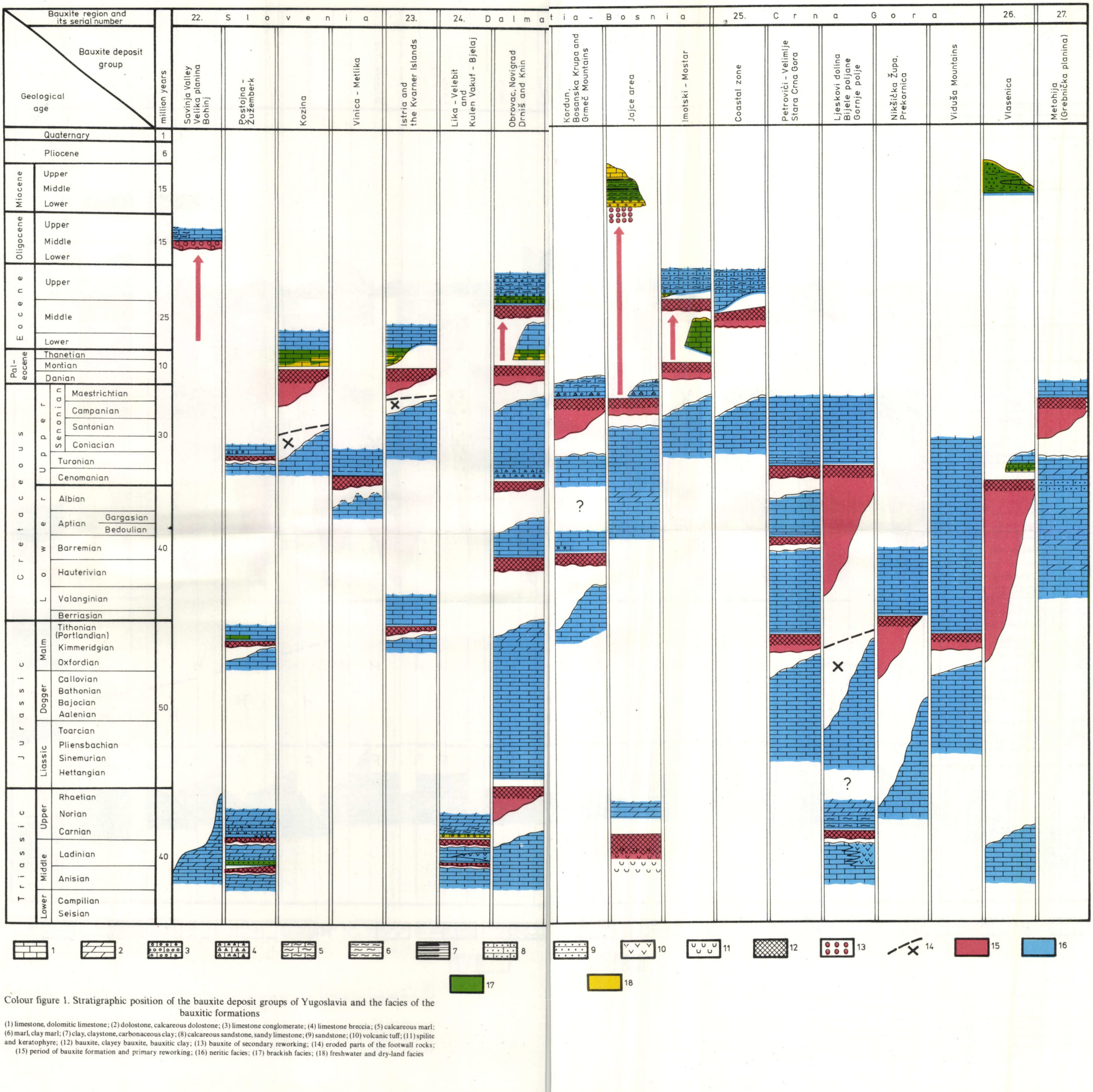
Appendix 3. Percentage abundances of aluminium mineral association in karst bauxites, by deposit groups

Serial number	Bauxite region Bauxite deposit group	Gibbsitic	Gibbsito Boehmito	Gibbsito–boehmito–diasporic	Gibbsito–corundous	Boehmitic	Boehmito diasporic	Diasporic	Diasporo–corundous	Corundous
1	2	3	4	5	6	7	8	9	10	11
38	Eastern Taurus (Turkey)									
	A. Adama, Bolkardag, Saimbeyli	–	–	–	–	–	–	60	40	–
	B. Kücük Koras, Sebilköy	–	–	–	–	70	20	10	–	–
39	Payas, Islahye	–	–	–	–	–	10	90	–	–
40	Zonguldak (Turkey)	–	–	–	–	–	100	–	–	–
41	Crimea (USSR)	–	–	–	–	–	80	20	–	–
42	Maktesh Ramon (Israel)	–	–	–	–	–	100	–	–	–
43	Little Caucasus (Julfa) (USSR)	–	–	–	–	–	–	50	50	–
44	Bukan (Iran)	–	–	–	–	–	–	100	–	–
45	Alborz Range (Iran)									
	A. Jajarm (Permo–Triassic)	–	–	–	–	–	–	100	–	–
	B. Jajarm (Upper Triassic)	–	–	–	–	–	–	100	–	–
46	Central Iranian Plateau									
	A. Yazd	–	–	–	–	–	–	100	–	–
	B. Kerman	–	–	–	–	20	60	20	–	–
47	Zagros Range (Iran)									
	A. Do-Polan, Do-Rag	–	–	–	–	20	40	40	–	–
	B. Behbahan (Sar-e-Fariab)	–	–	–	–	90	10	–	–	–
	C. Shīraz (Daryush Dam)	–	–	–	–	–	–	100	–	–
48	Kandahar	–	–	–	–	–	–	100	–	–
49	Pamirs (USSR)	–	–	–	–	–	–	70	30	–
50	Hindukush Range (Pakistan)	–	–	–	–	–	100	–	–	–
51	Northeast Pakistan									
	A. Salt Range, Sargodha	–	10	–	–	90	–	–	–	–
	B. Campbellpore, Muzaffabad	–	–	–	–	–	–	100	–	–
52	Jammu (Kashmir-India)	–	–	–	–	40	50	10	–	–
53	Samar Island (Philippines)	70	30	–	–	–	–	–	–	–
54	Solomon Island									
	A. Wagina and other islands	100	–	–	–	–	–	–	–	–
	B. Rennell Islands	100	–	–	–	–	–	–	–	–
55	Loyalty Group									
	A. Lifou Island	–	80	–	–	20	–	–	–	–
	B. Maré Island	0	90	–	–	–	–	–	–	–
56	Fiji-Tonga Group									
	A. Lau Islands (Fiji)	100	–	–	–	–	–	–	–	–
	B. Niué Island (Tonga)	100	–	–	–	–	–	–	–	–
57	Sisophon, Battambang (Campuchea)	–	–	–	–	40	20	40	–	–
58	Socialist Republic of Vietnam									
	A. Lang-Son (Upper Permian)	–	–	–	–	20	30	50	–	–
	B. Lang-Son (Neogene)	–	–	40	–	–	10	50	–	–
	C. Cao-Bang (Upper Permian)	–	–	–	–	–	20	70	10	–
	D. Cao-Bang (Neogene)	–	–	30	–	–	10	50	10	–
59	Kunming (Yünnan Province) (China)	–	–	–	–	–	30	70	–	–
60	Kweichow Province (China)									
	A. Kveyang (Carboniferous)	–	–	–	–	–	30	70	–	–
	B. Kveyang (Permian)	–	–	–	–	–	30	70	–	–
61	Gun District (Honan Province) (China)	–	–	–	–	–	–	100	–	–
62	Nanting, Po-San (Shantung Province) (China)	–	–	–	–	–	40	60	–	–
63	Penhsihu (Liaoning Province) (China)	–	–	–	–	–	30	70	–	–
64	Sikhote-Aliny Range (USSR)	–	–	–	–	100	–	–	–	–
65	Boksonsk (Eastern Sayan) (USSR)	–	–	–	–	–	50	50	–	–
66	Chadobetsk Area (USSR)	80	–	–	20	–	–	–	–	–
67	Region along Yenisei (USSR)									
	A. Angara area	70	–	–	30	–	–	–	–	–
	B. Ivanovsk area	70	–	–	30	–	–	–	–	–
	C. Tatarsk area	80	–	–	20	–	–	–	–	–
	D. Sulakshinsk area	100	–	–	–	–	–	–	–	–
68	Altay–Sayan Region (USSR)									
	A. Alatau of Kuznetsk, Gornaya Shoriya	–	–	–	–	–	–	10	30	60
	B. Berdsk, Obukhovsk	–	–	–	–	–	–	50	40	10
	C. Salair, Barzassk	100	–	–	–	–	–	–	–	–
69	Soviet Central Asia									
	A. Bukantau, Tamdytau, Nuratau	–	–	–	–	–	20	50	20	10
	B. Turkestan and Aliysk ranges	–	–	–	–	10	40	40	10	–
	C. Southern Ferghana	–	–	–	–	–	100	–	–	–
	D. Chimkent area (Liassic)	–	100	–	–	–	–	–	–	–
	E. Northern Ferghana	–	–	–	–	30	50	20	–	–
	F. Kansaysk area	100	–	–	–	–	–	–	–	–
	G. Chimkent area (Valanginian)	10	50	–	–	40	–	–	–	–
70	Central Kazakhstan (USSR)									
	A. Tselinograd area	–	–	–	–	–	–	60	40	–
	B. Tselinograd, Ekibastuz	80	–	–	20	–	–	–	–	–
	C. Arkalyk	70	–	–	30	–	–	–	–	–
71	Turgay Plateau (USSR)									
	A. Naurzumsk	100	–	–	–	–	–	–	–	–
	B. Western Turgay area	70	–	–	30	–	–	–	–	–
	C. Mugojar area	80	–	–	20	–	–	–	–	–
72	Southern Urals (USSR)									
	A. Novo-Pristansk	–	–	–	–	20	60	20	–	–
	B. Orsk, Magnitogorsk	100	–	–	–	–	–	–	–	–
	C. Chelyabinsk	100	–	–	–	–	–	–	–	–
73	Central Urals (USSR)									
	A. Nizhne-Serginsk, Pashiya	–	–	–	–	20	50	30	–	–
	B. Kosikovsk	–	–	–	–	–	30	70	–	–
	C. Kamensk-Uralsk	100	–	–	–	–	–	–	–	–
	D. Alapayevsk, Mugaysk	100	–	–	–	–	–	–	–	–
74	Northern Urals (USSR)									
	A. Subrovsk bauxite horizon	–	–	–	–	10	40	50	–	–
	B. Bogoslovsk bauxite horizon	–	–	–	–	10	40	50	–	–
	C. Pripolyarny, Polyarny Ural	–	–	–	–	–	–	100	–	–
	D. Bogoslovsk, Volchansk	80	–	20	–	–	–	–	–	–
75	Taimyr Peninsula (USSR)	–	–	–	–	100	–	–	–	–
76	Timan Range (USSR)									
	A. Central Timan	–	–	–	–	80	20	–	–	–
	B. Tobyssk, Kedvinsk	80	20	–	–	–	–	–	–	–
	C. Southern Timan	–	40	–	–	60	–	–	–	–
77	Tulsk, Ryazansk (USSR)	100	–	–	–	–	–	–	–	–
78	Puerto Rico	–	–	–	–	100	–	–	–	–
79	Dominican Republic	40	60	–	–	–	–	–	–	–
80	Haiti	20	80	–	–	–	–	–	–	–
81	Jamaica									
	A. St Ann Plateau	70	30	–	–	–	–	–	–	–
	B. Cockpit Country	100	–	–	–	–	–	–	–	–
	C. Essex Valley	60	40	–	–	–	–	–	–	–
	D. Manchester Plateau	80	20	–	–	–	–	–	–	–
	E. Clarendon Plateau	80	20	–	–	–	–	–	–	–
	F. St Catherine Plateau	70	30	–	–	–	–	–	–	–
82	Cayman Islands	–	100	–	–	–	–	–	–	–
83	Bahamas	–	100	–	–	–	–	–	–	–
84	Missouri (Gasconade County) (US)	–	–	–	–	10	20	70	–	–
85	Western rim of Coastal Plain (US)	100	–	–	–	–	–	–	–	–
86	SE rim of Coastal Plain (US)	100	–	–	–	–	–	–	–	–
87	SW Valley and Ridge Prov. (US)	100	–	–	–	–	–	–	–	–
88	NE Valley and Ridge Prov. (US)	100	–	–	–	–	–	–	–	–



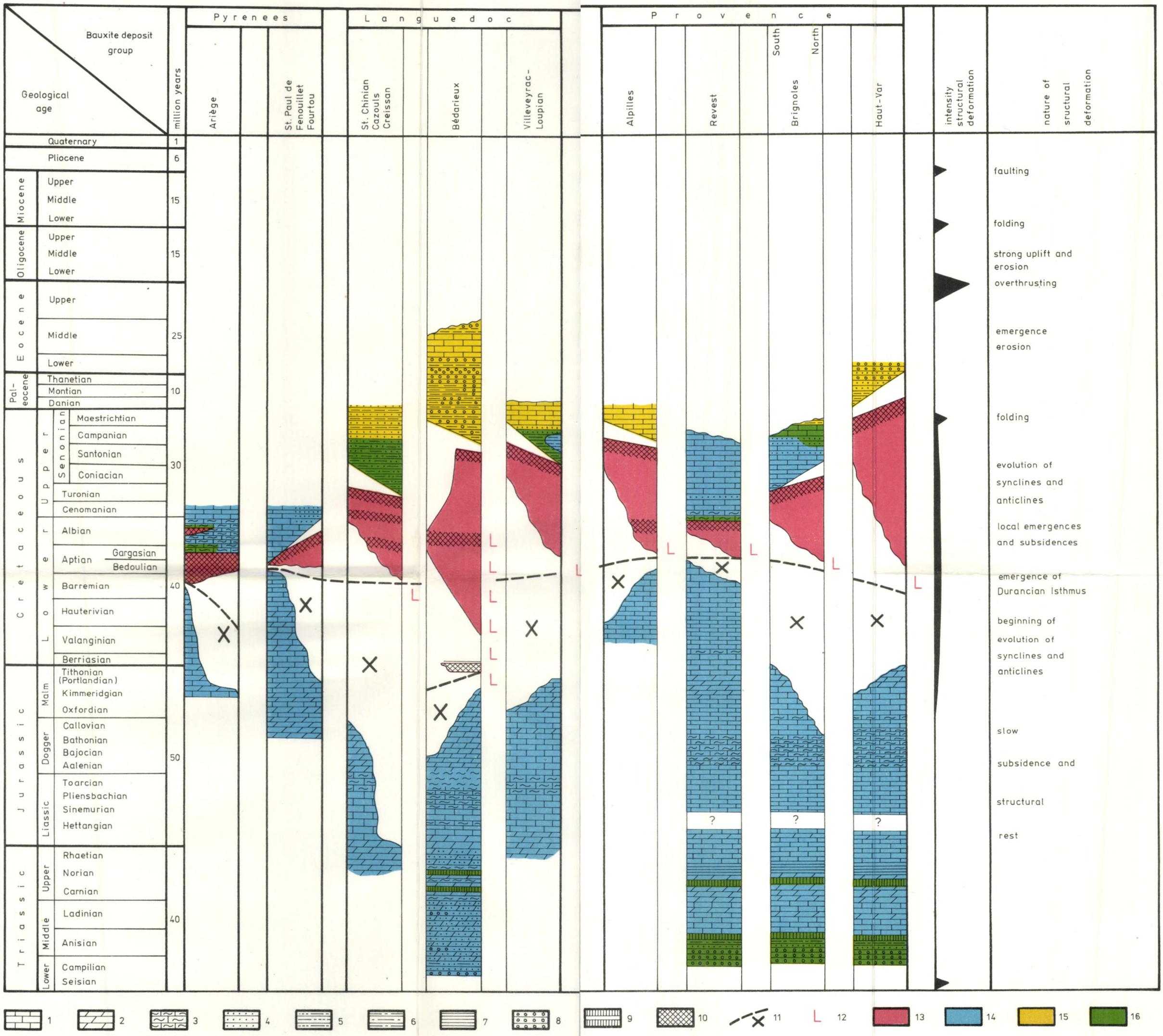
Colour figure 2. Stratigraphic position of the principal bauxite deposit groups of Hungary and the facies of the bauxitic formations

(1) limestone, dolomitic limestone; (2) dolostone; (3) calcareous marl, marl; (4) sandstone; (5) clay, sandy clay; (6) clay with detritus of manganese ore; (7) carbonaceous clay and lignite; (8) limestone- and dolostone-conglomerate; (9) limestone- and dolostone-breccia; (10) period of accumulation of bauxite deposits; (11) secondary reworking of bauxite deposits; (12) eroded cover; (13) an eroded, incomplete succession under the bauxite deposits; (14) an eroded, complete succession under the bauxite deposits; (15) period of bauxite formation and primary reworking; (16) neritic facies; (17) bracking and lagoon facies; (18) swamp and freshwater facies



Colour figure 1. Stratigraphic position of the bauxite deposit groups of Yugoslavia and the facies of the bauxitic formations

(1) limestone, dolomitic limestone; (2) dolostone, calcareous dolostone; (3) limestone conglomerate; (4) limestone breccia; (5) calcareous marl; (6) marl, clay marl; (7) clay, claystone, carbonaceous clay; (8) calcareous sandstone, sandy limestone; (9) sandstone; (10) volcanic tuff; (11) spilite and keratophyre; (12) bauxite, clayey bauxite, bauxitic clay; (13) bauxite of secondary reworking; (14) eroded parts of the footwall rocks; (15) period of bauxite formation and primary reworking; (16) neritic facies; (17) brackish facies; (18) freshwater and dry-land facies



Colour figure 23. Comparison of the principal bauxite deposit groups of France as to stratigraphic position, facies and relationship to the structural evolution of the Pyreneo–Provencal intracontinental orogen

(1) limestone, dolomitic limestone; (2) dolostone; (3) calcareous marl, marl; (4) sandstone; (5) clayey sandstone; (6) sandy claystone; (7) claystone; (8) conglomerate; (9) gypsum; (10) period of accumulation of bauxite deposits; (11) eroded footwall succession; (12) laterisation in nearby regions; (13) period of bauxite formation and primary reworking; (14) neritic facies; (15) freshwater and swamp facies; (16) brackish and lagoon facies

ISBN 963 05 2480 5