ERUPTIVE ROCKS

By the same author

EARTH LORE, GEOLOGY WITHOUT JARGON
THE STUDY OF ROCKS
USEFUL ASPECTS OF GEOLOGY

PLATE I. The Palisades of the Hudson, opposite New York City. The dark band running along the face of the cliff is the "olivine ledge." (For description see page 147.) Photograph by Freeman Foote.
Frontispiece]

ERUPTIVE ROCKS

THEIR GENESIS, COMPOSITION, CLASSIFICATION, AND THEIR RELATION TO ORE-DEPOSITS

WITH

A CHAPTER ON METEORITES

by

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First, of course, there's the things you don't know; then there's the things you do know but can't understand; then there's the things you do understand but which don't matter. Saving your presence, sir, there's a heap of understanding to be done before you're anything but a fool.—A. E. COPPARD, *Simple Simon.*

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PREFACE
TO THE FIRST EDITION

THIS book is the outcome of three insistent wishes which have occupied the writer's mind ever since he came under the spell of the study of eruptive rocks.

There was first the wish to clean up the jungle of rock names. The method of petrographers in giving names, as a great American geologist has pointed out, has been to give an independent name to every rock that is a little different from any one seen before, without reference to any plan; a practice which has led to the multiplication of rock names to a degree far in excess of the needs of the science. The consequence is not merely that one cannot see the wood for the trees, but rather that one cannot see the trees for the brushwood. This state of affairs pleases nobody, yet many would rather perpetuate it than face the revolution of names and terms that a new method involves. Thus a reformer is between the devil and the deep sea, for if he adapts a new system to existing names, with their connotations, it will be still-born, and if he introduces a new nomenclature, it will be stoned to death by those who feel that the burden of rock names is already greater than they can bear. But something must be done; somebody must make a move; and that is the first reason for this book.

A second wish arose out of the first. It was to make the results and conclusions of petrology intelligible to matter-of-fact people like physicists and chemists, who like to do their thinking in quantitative terms, and who are apt to listen with indifference to the assertions of "affinity" and "consanguinity" that underlie so many petrographic contentions. Petrology needs all the help that chemistry can give it; but before they look for help, petrographers must learn to express their problems in terms that a chemist can understand. That means that they must pay more attention than has been their habit to the chemical resemblances and differences between rocks, and above all, that they must give a quantitative meaning to the terms they use. To make a step in this direction is the second reason for this book.

Thirdly, it has been the writer's wish to bring theory and practice together, and to save petrology from the reproach of being an academic study, of little use in life. Petrology and mineralogy owe much to mining, and should give something in return. A rational system of petrology, that is neither too vague nor too difficult to understand, is a gift that the miner and all who are interested in the mineral industries would not fail to appreciate. On the other hand, the petrologist who turns his attention to the study of ore-deposits will find it a rare source of

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information and inspiration. To show that the miner and the petrologist have need of each other's help is the third reason for this book.

. . . .

In the matter of terminology and nomenclature, the writer has gone his own way, guided by the general principles that names should be as few as possible, that the simplest terms are the best, and that when an English term can be used without obscurity, it should be preferred to a Latin or Greek one. In view of the repugnance which a new set of names would arouse, he has used only well-known names of wide implication. These have been redefined, but it will be found that the new definition contains the kernel of the old one in every case. To define afresh all the varietal names that are in use would be an impossible task, even if it were desirable or legitimate. Better to let them all die out; and to introduce new ones from time to time, as the need for further subdivision is felt; so the names will arise out of the classification, as they ought to do, instead of the classification growing round the names.

PREFACE
TO THE THIRD EDITION

In preparing a third edition of *Eruptive Rocks*, the writer has taken the opportunity of reintroducing a discussion of the link between eruptive rocks and ore-deposits. This formed the subject of a chapter in the first edition of the book, but it was omitted from the second edition. Certain aspects of the matter, which seemed quite clear in 1926, no longer seem so in 1946; so the discussion has spread itself over three chapters instead of the original one. The increase does not so much express new knowledge as new doubts and difficulties; but that is the way of science.

The other chapters have been shaken up, and a certain amount of new matter has been introduced. The four chapters of modal analyses of rocks, which are intended to illustrate the writer's system of classification, have been removed to the end of the book, where they form Part II. It is hoped that this rearrangement will make Part I more readable and Part II more accessible.

The writer expresses his thanks to many friends, colleagues and reviewers for criticism which was always welcome and helpful, even when he found himself unable to agree with it. It may not be amiss to refer future readers to a paper entitled "The Species Concept in Petrology"[1] which summarizes the writer's attitude toward the basic problem of our science. A reader who cannot share the viewpoint indicated in that paper will find little to please him in this book.

[1] *American Journal of Science*, vol. 242, 1944, p. 45.

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If a man will begin with certainties, he shall end in doubts; but if he will be content to begin with doubts, he shall end in certainties.—FRANCIS BACON.

PART I

CHAPTER I

ERUPTIVE OR IGNEOUS ROCKS

PETROLOGY begins with the examination, description and classification of the kinds of rock that form what we call "the crust of the earth," but it does not end there. The philosophical aim of the study, which is too often overlooked, is to discover the chemical composition of the matter beneath the crust, to throw light upon the physical conditions that prevail there, and to investigate the physical and chemical changes that take place during the formation of rocks. Physicists may weigh the earth, geodesists find its form and volume, astronomers record its movements, and seismologists study its elasticity, while philosophical geologists build theories of the earth upon these data; but the petrographer alone has tangible evidence of what lies beneath the crust. It is presented to him by his own planet in the form of plugs and tongues of eruptive rock that have come up from the depths, and by mineral veins. It is presented to him by other members of the solar system in the form of meteorites. No theory of the earth or of the solar system can stand that disregards the evidence of eruptive rocks and meteorites. And since the problem of the earth's composition, with which that of its genesis is bound up, surely overshadows all other non-organic problems, so petrology, as the science which contributes most directly to the solution of that problem, may claim a place in the front rank of the natural sciences.

The practical aim of the study is also an ambitious one. Mineralogical and petrological research shows ever more and more clearly that an intimate connection exists between eruptive rocks and mineral deposits of commercial importance. It is already possible to lay down some general principles in the matter, connecting certain minerals and certain special types of ore-deposition with particular groups of eruptive rocks and particular types of intrusion. It is evident that well established generalizations of this kind may be very valuable in the hands of a miner or prospector who has a sufficient knowledge of petrology to apply them correctly. There is much to be learned before the search for mineral deposits can be placed upon a sound scientific basis; but this and nothing less should be the aim of petrology on the practical side.

The science of petrology is generally considered to embrace the study of all rocks, whether eruptive or sedimentary, and whether affected by metamorphism or not. It is well to realize that this arrangement is dictated solely by convenience. The sedimentary rocks, having been formed by the disintegration of older eruptive rocks, have about as much in common with the latter as sawdust has with a living tree; consequently the methods of study and classification which are suited to the two classes of rocks are quite distinct. Biology long ago found it necessary for its health and progress to split itself up into the two sciences of zoology and botany, which have their roots in common ground but otherwise live their lives independently; and for the same reason there is need for petrology to separate its two branches which are concerned with eruptive and with sedimentary rocks respectively, permitting them to meet only in the common ground of metamorphism.

In this volume the eruptive rocks alone are considered, with brief reference to the mineral deposits to which they give rise and the meteorites to which they are nearly related.

We may open the discussion with the remark that the adjectives "igneous" and "eruptive" are currently applied to rocks as if they were synonymous.[1] Are all eruptive rocks igneous, using that word in the fullest sense which is warranted by its derivation from *ignis*, fire? It is true that in the case of lava discharged by an active volcano, we have the evidence of our senses to assure us of its "fiery" or molten origin. But this evidence is lacking in respect of other rocks such as granite and syenite, because these rocks are formed

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[1] The term *roches éruptives* was used by Elie de Beaumont in 1847, and *Eruptivgesteine* by Bernhard von Cotta in 1855. These terms imply a breaking-out, not necessarily to the surface of the earth, but from lower to higher levels in the earth-crust. Von Cotta made a distinction between plutonic eruptive rocks and volcanic eruptive rocks.

deep down in the crust of the earth, and they only become accessible to our observation after ages of denudation. What justification is there for putting deep-seated rocks like granite and syenite in the same category with the really "igneous" lavas? This is the very question that agitated the Neptunists and Vulcanists of last century, and as the matter lies at the root of the science of petrology, we shall devote this chapter to its consideration.

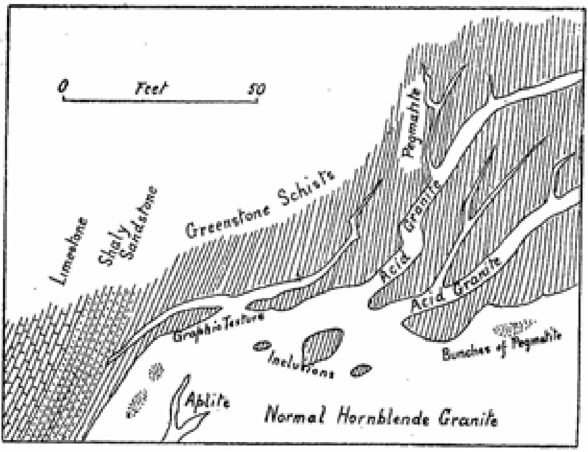

FIG. 1. Part of the margin of the Boulder batholith, Montana. (After P. Billingsley.)

The proof that granite and other deep-seated crystalline rocks were fluid at one time in their history is afforded by the manner of their occurrence in the earth's crust. Wherever the junction between a mass of granite and any other rock is exposed to view, it is found to be a very irregular surface, often approaching the vertical position. In the case of the Marysville granite mass in Montana, the upper surface of the rock was thoroughly explored in mining operations, and Joseph Barrell reports that it was "extremely irregular, steep or vertical surfaces alternating with flat surfaces, giving the whole a rudely pyramidal form with a somewhat step-like surface, from which project dykes, sheets and chimneys."

In the neighbourhood of such a junction, the granite is often seen to enclose blocks of the surrounding rocks, which have been detached from the walls and transported towards the heart of the granite mass. If a series of such enclosed blocks is examined, from the junction inwards, one readily convinces oneself that the foreign rock was actually being corroded, disintegrated, and incorporated in the granitic magma,[1] and that the action was only stopped by the freezing of the mass. (See Pl. III., p. 77.) From such a junction, too, the granite commonly sends offshoots in among the country rocks, either in the form of sheets which follow the bedding planes of the country rocks, or of dykes which cut across these planes, or of thin veins that penetrate the most minute cracks. Or, instead of a regular injection of magma into fissures, the country rock may present the appearance of having soaked up the magma much as a sponge soaks up water, so that the product is charged with feldspar and other minerals apparently derived from the granite magma. The evidence of fluidity on the part of the magma, which is presented by contact phenomena such as these, seems to be incontrovertible.

The proof of fluidity, however, does not carry with it that of high temperature. It is true that no solvent is known which will dissolve and deposit appreciable quantities of

[1] The Greek word "magma" has the meaning of a plastic mass, or a paste of solid and liquid matter. The word has long been used in pharmacy (as in "magnesia magma" or milk of magnesia) and it passed in the 18th century from pharmacy into chemistry. Thorpe's *Dictionary of Applied Chemistry* (1922) gives the definition, "a pasty or semifluid mixture." From chemistry this word passed into petrology, replacing the older term "subterraneous lava." G. Poulett Scrope wrote in 1862 that "lava is not a homogeneous molecular liquid, but . . . a *magma*, or composed of crystalline or granular particles to which a certain mobility is given by an interstitial fluid." With the development of petrology the emphasis has shifted from the presence of crystalline particles to the liquid. Since some lavas are completely free from crystals, and others contain abundant crystals while they are still flowing, we might say that "rock-magma is the partly or wholly liquid material from which eruptive rocks are formed." But there is no evidence that deep-seated rock magma was ever hot enough to be completely liquid. The matter is discussed in Chapter IV.

quartz, feldspar and mica at ordinary temperatures; yet many instances can be named in which crystals of quartz and feldspar have been formed in rocks without the aid of heat. Such are the veins and pockets of rock crystal which one often finds in sandstones and limestones. In certain cases, feldspar and quartz have been formed in unconsolidated deposits on the ocean bottom, under circumstances which seem to preclude the operation of any higher temperature than that of sea water. Many substances which are incapable of true solution may form colloidal suspensions (sols) which serve equally well as media for transport and deposition, and which require no very high temperature for their formation. The hydrothermal solubility of silica has been found by G. C. Kennedy to be only 0.2% at 360°; but silica sols have been prepared in the laboratory which hold as much as ten per cent. of silica; they soon gelatinize, but weaker sols are more stable. From such a sol, gelatinous silica is deposited which may turn into opal, chalcedony, and eventually into quartz.

That water is actually present during the crystallization of granite, and still more of vein-quartz, is abundantly proved. It has long been known that quartz crystals often contain drops of liquid, and gas bubbles, and the investigations of Brewster, Humphrey Davy, Hartley and many later workers have shown that the liquid is just salt water. Although the cavities in which the fluid is found are generally very small, often no more than one ten-thousandth of an inch in diameter, yet their combined volume is considerable. The following calculation is due to A. A. Julien :—

> "In the quartz of Cornish granites the cavities sometimes make up at least five per cent of its volume, and the water in them, on an average, about one per cent of its volume, or four-tenths of one per cent of its weight. As the quartz grains in a granite or in gneiss rarely amount to less than fifty per cent of the whole rock, it is probable that the liquid contents of such a granite commonly reach as much as one-fifth of one per cent.; *i.e.* about four pounds of liquid to a ton of rock."

It is hardly possible to doubt that the water and the dissolved salt were actually present in the magma from which the quartz crystallized. Quartz crystals are almost devoid of cleavage, and except by way of cleavage planes or other cracks the liquid could not have got into the crystals after they were full grown. And if it could enter by way of minute cracks, it could escape by the same way; yet the contents of the cavities are sometimes found to be under pressure; indeed some of them contain, instead of water, liquid carbon dioxide, which at its critical temperature of 31° C. requires a pressure of 73 atmospheres to keep it liquid. Other minerals besides quartz show such enclosures, notably topaz, and in general it is the minerals which were last to crystallize from the magma that are richest in fluid enclosures, indicating that the residual magma was progressively enriched with water and gas as crystallization advanced.

An important investigation of the enclosures in the minerals of eruptive rocks was carried out by Henry Clifton Sorby, who was the first to apply to the study of rocks the method of preparing thin sections by grinding, which had been developed by William Nicol about 1830.[1] Sorby showed that in the minerals of granites the enclosures are mostly of liquid or gas, but that in lavas there are present similar cavities which contain *glass* instead of fluid. In some cases, owing to the glass having become minutely crystalline, the cavities contain an exceedingly fine-grained stony material. Between rocks which showed glass enclosures only, and others which showed liquid enclosures only, a connecting link was found in certain rocks which hold enclosures of both kinds; this condition Sorby observed in the "trachyte" of Ponza, and in some Cornish granites and granitic dykes. F. Zirkel afterwards recorded similar phenomena in leucite crystals, the same cavity sometimes holding both glass and liquid.

The main conclusions to which Sorby was led by his discoveries are given here, in his own words:—

> "On the whole, then, the microscopical structure of the constituent minerals of granite is in every respect analogous to that of those formed at great depth and ejected by modern volcanoes, or that of the quartz in the trachyte of Ponza, as though granite had been formed under similar physical conditions, combining at once both igneous fusion, aqueous solution and gaseous sublimation. The proof of the operation of water is quite as strong as that of heat; and in fact I must admit that in the case of coarse-grained, highly quartzose granites there is so very little evidence of igneous fusion, and such overwhelming proof of the action of water, that it is impossible to draw a line between them and those veins, where, in all probability, mica, felspar, and quartz have been

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[1] Nicol's thin sections were described by Witham in *Observations on Fossil Plants*, London, 1831, p. 45.

deposited from solution in water, without there being any definite, genuine igneous fusion like that in the case of furnace slags or erupted lavas. There is, therefore, in the microscopical structures, a most complete and gradual passage from granite to simple quartz veins."

"With respect, then, to minerals and rocks formed at a high temperature, my chief conclusions are as follows. At one end of the chain are eruptive lavas, indicating as perfect and complete fusion as the slags of furnaces, and, at the other end are simple quartz veins having a structure precisely analogous to that of the crystals deposited from water. Between these there is every connecting link, and the central link is granite."

The evidence, derived from the study of the liquids enclosed in crystals, that water actually played a part in the magma from which granite consolidated, is confirmed by many field observations. Near any great granite intrusion it nearly always happens that dykes and veins of the eruptive rock are thrust out into the country rock, often to considerable distances. Even commoner than these dykes are seams of pure white quartz, which on account of their composition and structure have always been assumed to be of aqueous origin. But it is sometimes possible to follow a single dyke from its point of origin on the edge of the granite mass to a position where, by gradual loss of feldspar and mica, it passes without any lack of continuity into a quartz vein. In illustration of this highly important observation we select the description by C. R. van Hise of the granite dykes of the Black Hills in South Dakota.

"The core rock is a great batholith of granite, 7 miles broad and 11 miles long. . . . From the central mass great quartz-feldspar dikes radiate. In passing away from the core the dikes become smaller and have a less typical form. At the same time the material assumes the appearance which we ordinarily denominate as pegmatitic. These veins might be considered true igneous injections. Passing still further away, the pegmatitic masses begin to have vein-like characters—that is, there is a rough concentration of the material in different layers, parallel to the walls. In going still further away a true banded vein strucrure is found. In passing yet further away, feldspar becomes less and less important in the veins and the quartz more abundant, until remote from the granite the impregnating material is mainly quartz. . . .

If one examined only the outer zone, in which the quartz impregnations occur, he would not doubt that they were ordinary deposits from water; if he examined the masses of the inner zone, adjacent to the great granite mass, he would not doubt that they were true igneous injections; but passing back and forth from the one zone to the other, he observes that there is every gradation between the two."

Similar observations have been made in many parts of the world, but we can allow ourselves only one more quotation, which we take from the memoir by F. D. Adams and A. E. Barlow on the geology of the Haliburton and Bancroft areas in Ontario:—

> "The pegmatites are prevailingly rich in feldspar, this constituent largely predominating over the quartz. In places, however, this latter constituent becomes more abundant, especially at the extremity of the dike where it becomes narrow. In fact, cases can be observed where the feldspar is represented only by a few individuals here and there, and the dike develops into a mass of quartz. Such an occurrence, if seen alone, would undoubtedly be considered as a quartz vein, but in this area no genetic distinction can be drawn between the two occurrences. The quartz veins, in these cases at least, are merely a form or phase of the pegmatite dikes."

The effect of much evidence of this kind is to confirm the opinion that water is a normal component of granitic magma, and to show that the distinction between "igneous" and "aqueous" products is much less definite than it was formerly supposed to be. That H. C. Sorby should have formed a correct opinion on this subject as long ago as 1858 is remarkable and shows him to be one of the greatest of the founders of petrology. The seal of confirmation has been placed upon Sorby's view by R. W. Goranson, who has lately succeeded in measuring the solubility of water in molten granite. At the temperature of 900° C., and under a pressure equal to a depth of fifteen kilometres in the earth-crust, more than nine per cent. by weight of water was taken into solution and was retained when the melt cooled down to the state of glass.

Even lavas such as rhyolite and basalt—the most typical of "igneous" rocks—furnish evidence of the presence of water in the magmas from which they were derived. Many lavas carry minerals such as mica or zeolites, into whose composition water or its elements enter. The more siliceous lavas hold biotite, from which, at a temperature of eight or nine hundred degrees, about five per cent of water can be extracted. Hornblende too gives up a little water when strongly heated. The less siliceous lavas often exhibit cavities lined with crystals of the zeolites, a group of minerals which earned their name (literally *boiling-stone*) by giving up large quantities of water when gently heated. David Forbes produced before the Chemical Society in 1868 a specimen of the lava ejected by Mt. Etna during the eruption of 1865; this specimen, taken while the lava was still flowing, contained crystals of stilbite, a zeolite holding seventeen per cent of water of crystallization. In recent years the work of Day, Shepherd and Jaggar in Hawaii has shown that water is a constant component, and by far the largest component, of the exhalation from the lava lake at Kilauea; and the weight of the evidence indicates that this water is of magmatic rather than meteoric origin. We shall return to this matter in a later chapter.

H. C. Sorby tried to ascertain the actual freezing temperatures of some granites by observing the temperature at which the liquids, enclosed in the cavities of quartz crystals, completely filled these cavities. In his own words:—

> "When crystals are formed at an elevated temperature, evidence of it is afforded by the contraction of the fluids enclosed in the cavities giving rise to a vacuity. . . . Since, of course, the amount of the contraction of the fluid depends upon the height of the temperature from which it has cooled, the relative size of the vacuity must indicate how much the temperature at which the crystals were formed was above that at which they are examined, in the same manner as the sinking of the mercury in a self-registering thermometer shows the difference in the temperature."

By the application of this principle, Sorby obtained results ranging from 89° C., for some granites, to 356° C. for a quartz-bearing lava. More recent work shows that Sorby's estimates were almost certainly too low, but that many of the coarser granite pegmatites completed their crystallization below 575° C., and most of the quartz-bearing rocks probably below 700° C. The truth is, however, that none of the methods yet employed for settling the question is quite beyond criticism, and all that can be said with assurance is that the granites have been formed at temperatures decidedly lower than the freezing temperature of any known lava.

J. Clifton Ward, who applied Sorby's methods to the granites of the Lake District of England, came to the conclusion that eruptive magma may exist in two states, which he describes as follows:—

> "(1) That state in which igneous fusion is the most important or conspicuous element. This is seen in the case of lava flowing from a volcanic vent.
>
> (2) A state of aqueo-igneous fusion, occurring at a much greater

> depth than the last, and reaching only a dull red heat (*i.e.* about 400°)[1] as a maximum. In this state we have every reason to suppose the granite magma to be, which may now be solidifying . . . beneath our feet."

The reliability of Sorby's ingenious method was questioned by S. Allport, who pointed out that if the cavities were originally completely filled with liquid, and the bubbles were formed solely by contraction, then the ratio of size of cavity to size of bubble ought to be constant in any one part of the rock. But Allport found that this ratio was not constant either in the same rock or even in the same crystals.

Sorby's method has lately been revived by W. H. Newhouse and used to determine the temperature of formation of ores. Newhouse shows that Sorby failed to distinguish between primary enclosures of liquid or gas and secondary enclosures due to fracturing and subsequent healing of the crystals at a lower temperature than that at which they were originally formed. This may well be the explanation of the discrepancies observed by Allport. If more care were given to making this distinction, it seems possible that Sorby's method might again be put to use to determine the temperature at which the minerals of eruptive rocks crystallized. Newhouse has applied the method to the lead-zinc ores of the Mississippi valley and finds that they were formed between 80° and 135° C.

The temperature of flowing lava has been variously estimated at 1000° to 1200° C. and even higher. The latest and most reliable measurements made by T. A. Jaggar at Kilauea show that such extremely high temperatures are only realized at the surface of the liquid, where escaping gas is in active combustion; but that the temperature of the lava below the surface may be 750° to 850° C. At this relatively low temperature it is likely that crystallization of the lava has already begun, for well-formed crystals of augite and leucite have been observed in perfectly fluid lava at Mount Vesuvius.

It follows, then, that we are justified in placing granite and other plutonic rocks in the same category as volcanic lava

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[1] This is the figure given by Ward, but Dr. R. B. Sosman informs me that the "first visible red" appears between 500° and 550° C., varying with different observers.

and in a different one from sedimentary rocks; but it by no means follows that we are justified in applying the name "igneous rock" to granite and lava alike. The lavas are igneous rocks; so too are many of the rocks that form dykes and sills, for the presence of glass in them shows that they have been molten. But among plutonic rocks, and particularly granite, the evidence for the agency of water as a contributory cause of fluidity is just as strong as the evidence of high temperature. If the term "igneous" suggests that the rocks to which it is applied were once in the state of anhydrous melts, with temperatures of 1000° C. or more, then it is better to throw it over in favour of an alternative adjective such as eruptive or magmatic, which conveys the idea of fluidity without any assertion or implication about temperature.

The foregoing summary has indicated that the long-established practice of placing the "igneous rocks" in one category, in opposition to the "aqueous" mineral veins, requires modification. Some mineral veins are closely akin to eruptive rocks, and ought not to be abruptly separated from them. Some eruptive rocks have more kinship with the contents of mineral veins than they have with lavas. A truer view of the relation of the eruptive rocks to one another, and to a certain class of mineral veins, is given by a continuous linear classification such as the following:—

Conditions	*Products*
Temperature of consolidation high (Above 700° C.?) Water sparingly present	Extrusive rocks (lavas)
Temperature of consolidation high to moderate (700° – 500° C.?) Water present in notable quantity	Most intrusive rocks
Temperature of consolidation moderate (600° – 300° C.?) Water present up to the probable limit of 9%	Coarse-grained rocks with much alkali-feldspar and quartz or nepheline. Some ore bodies
Temperature of consolidation low (Below 300° C.?) Hot aqueous solution	Quartz veins and many mineral deposits

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Advance in science comes by laying brick upon brick, not by the sudden erection of fairy palaces.—J. S. HUXLEY.

CHAPTER II

THE FIXED CONSTITUENTS OF ERUPTIVE ROCKS

A REMARKABLE feature of the eruptive rocks is the small number of mineral species of which they are composed. Out of the hundreds of silicates and aluminosilicates that are known, or that are theoretically possible, only a very few are common constituents of eruptive rocks; and among these the feldspars, pyroxenes, amphiboles and micas, together with quartz, predominate enormously over all others. For this reason the investigation of an eruptive rock usually requires the determination of a few silicates only, together with certain oxides, sulphides, phosphates and titanates which may be present in trifling quantities.

This apparent simplicity is more than neutralized by the very complex constitution of many rock-forming silicates, and by the possibility of atomic substitution, or, in an earlier terminology, of solid solution. The concept of solid solution implies that there may be present, in a single crystal, molecules of two or more substances which are related to each other both chemically and crystallographically. For example, common olivine may be described as a solid solution of forsterite molecules, Mg_2SiO_4, and fayalite molecules, Fe_2SiO_4; and the precise composition of any sample of olivine may be stated in terms of these two end numbers, as Fo_{86} Fa_{14}. This concept and the corresponding terminology remain in general use; but the light thrown upon the structure of crystals by means of X-ray analysis has led to the adoption of a truer concept. As W. L. Bragg and others have shown, the silicates are distinguished from other inorganic substances by a type of atomic structure which has an indefinite extension in space. The structure of a silicate may be represented as an infinitely extended network of linked oxygen atoms, with atoms of silicon and metals occupying the spaces within regular groups of oxygen atoms. In such a network there are no molecules. Any atom in the structure (except oxygen) may be replaced by another atom of similar ionic radius,[1] without causing any serious distortion of the structure; just as a bricklayer, running short of red bricks, may incorporate a yellow brick of the same size here and there in his wall. Thus the atoms of aluminium, chromium, vanadium and ferric iron are mutually replaceable in the structure, and so are those of magnesium, ferrous iron, manganese and nickel. Not only may one element be replaced by another of the same valency, but there may be an interchange of atoms of different valencies, thus Na—Si may be replaced by Ca—Al, K—Si by Ba—Al, Mg—Si by Al—Al, Si by K—Al, or Al_2 by Mg_3. The consequence is that in most of the silicates as we find them in nature there are present not only the particular metallic atoms that are characteristic of each so-called species, but also atoms of other metals which have been able to satisfy the requirements of valency and ionic radius and fit themselves into the structure. For instance, the mineral diopside is theoretically a simple metasilicate of magnesium and calcium; but on account of the partial substitution of certain atoms by others, a chemical analysis of natural diopside will generally show a considerable content of iron and smaller quantities of manganese, nickel, sodium, aluminium and chromium. We may, if we please, describe diopside as a solid solution of the metasilicates of all these metals, but it is more illuminating as well as more correct to discuss the matter in terms of atomic substitution.

We have now to discuss the chemical composition of the principal rock-forming minerals. In doing so we shall sometimes employ one method of description and sometimes the other, partly because most petrologists are accustomed to think

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[1] One may speak either of ionic radius or of atomic radius, as the atom is ionized or not. Most inorganic compounds are ionic, so the use of the former term is preferable; but it is important to realize that the ionic radius is not a fixed quantity but varies with the positive or negative charge on the ion. Positive ions are smaller than the corresponding neutral atom, negative ions are larger. An element has as many ionic radii as it has valencies, for example:—

Fe (neutral atom) 1·24; Fe^{2+} 0·83; Fe^{3+} 0·67;
Mn (neutral atom) 1·18; Mn^{2+} 0·91; Mn^{3+} 0·70; Mn^{4+} 0·52.

in terms of "isomorphous compounds" and partly because the terminology of ionic substitution has not yet provided a simple expression to take the place of "solid solution." Perhaps "ionic substitution complex" might be used, and "ionic substitution series" in place of the familiar "solid solution series.'

THE FELDSPAR FAMILY

It has long been customary to think of this family in terms of the three end-members orthoclase, albite and anorthite, which give rise, by isomorphous crystallization, to several series of solid solutions. A truer picture shows us a continuous network of SiO_4 and AlO_4 tetrahedrons, with atoms of potassium, sodium or calcium fitted into appropriate spaces in the network. When only potassium atoms are present the result is *orthoclase*, $KAlSi_3O_8$, which crystallizes with monoclinic symmetry. The pure sodium feldspar is *albite*, $NaAlSi_3O_8$, and because the atoms of sodium are smaller than those of potassium the angles between the planes of the network are slightly different and the crystals are triclinic. The substitution of Ca-Al for Na-Si in the network gives *anorthite*, $CaAl_2Si_2O_8$, which is again triclinic. But if the large atoms of barium are present instead of the smaller calcium atoms, the result is the rare feldspar *celsian*, $BaAl_2Si_2O_8$, which is almost monoclinic.

Orthoclase forms with albite the very important ionic substitution-series which we call *soda-orthoclase*. The more potassic members of the series are monoclinic; indeed, most of the orthoclase that we identify in rocks is really soda-orthoclase holding about eight times as much potash as soda. The more sodic members of the series are triclinic and are often called *anorthoclase*, while those very rich in soda should be called potash-albite. In addition to these homogeneous varieties there is a group of feldspars composed of sub-parallel lamellæ, alternately sodic and potassic in composition; to these feldspars the names *perthite* (microperthite, cryptoperthite) and *antiperthite* have been given. In the perthites the sodic member forms films, veins and patches within the potassic member; in the antiperthites this relation is reversed. The work of O. Andersen and others has shown conclusively that some perthite is a product of metasomatic replacement, caused by circulating solutions, but that much of it is due to a change in the atomic structure of the feldspar itself. It is believed that when the crystals were formed, presumably at a rather high temperature, the sodium and potassium atoms were evenly distributed through the network; on cooling down to atmospheric temperature the structure became unstable and the large potassium atoms and the small sodium atoms were segregated in different layers. It was established by Dittler and Köhler that microperthite, when heated for a long time at 1000° C. reverts to homogeneous soda-orthoclase. This is confirmed by E. Spencer, who has shown that microperthite can be dissolved and reprecipitated without difficulty. Most of the solution takes place between 350° and 750° C., and exsolution is so easy that in Spencer's opinion "the rate of cooling of quickly chilled lavas would not prevent complete exsolution if the rapid cooling affected only that temperature range."

A further complication in the study of the potash-soda feldspars is introduced by the species microcline, sanidine and adularia. *Microcline* is characteristic of granite-pegmatites and other coarse-grained, low temperature rocks, and *sanidine* of acid lavas, so it is probably correct to regard the former as a low-temperature form and the latter as a high-temperature form of potash-feldspar. Dittler and Köhler heated microcline for 500 hours at 1000°. The cross-hatching disappeared and the extinction angle on (001) changed from 15° (microcline) to 0° (orthoclase). *Adularia* is a very low-temperature form, confined to hydrothermal veins.

The group-name *plagioclase* is applied to the ionic substitution series between albite and anorthite, potassium ions being absent or quite subordinate. To those sections of the plagioclase series which hold respectively from 10 to 30, 30 to 50, 50 to 70, and 70 to 90 per cent by weight of anorthite, the specific names oligoclase, andesine, labradorite and bytownite have been attached. Some analyses of plagioclase feldspar, or of rocks containing it, have shown a deficiency of silica; from this Washington and Wright deduced the presence of *soda-*

anorthite, $Na_2Al_2Si_2O_8$ (at that time supposed to be identical with the synthetic mineral carnegieite) in solid solution. Later investigation by Ernst and Nieland, of the feldspar from Linosa upon which the observations of Washington and Wright were made, have failed to corroborate the supposed deficiency of silica; and Bowen and Greig have proved that carnegieite crystallizes in the isometric system and does not have the structure of anorthite. Nevertheless, the existence of soda-anorthite seems to be theoretically possible—it involves the replacement of one calcium atom by two sodium atoms in the anorthite structure—and a number of independent observers have been constrained to assume its existence in order to explain the composition of certain rocks. T. F. W. Barth (1) pointed out that many basaltic lavas of the Pacific islands have nepheline in the norm (the calculated mineral composition) although that mineral cannot actually be detected in the rocks; he concluded that soda-anorthite must be contained in the plagioclase. The present writer found it necessary to make the same assumption in the case of certain lavas from Mauritius. F. F. Osborne, T. Krokström, and Nockolds and Zies have also favoured this explanation of the abnormal composition of certain feldspars; but D. Beljankin prefers to assume an excess of sodium aluminate rather than a deficiency of silica in such cases. Whichever explanation one adopts, the fact seems to be established that some plagioclase feldspar contains less silica than is appropriate to the simple albite-anorthite substitution series.

Barium ions are present in small quantity in most potash-rich feldspar; also in large quantity in an abnormal plagioclase described by Nockolds and Zies. But barium is not an important constituent of normal eruptive rocks.

Between potassium atoms and calcium atoms there is so great a difference in ionic radius that potassium-calcium feldspar seems to be an impossibility. Potash-rich feldspar in nature carries little more than a trace of lime, and lime-rich feldspar scarcely any potassium; but the more sodic feldspars such as anorthoclase and potash-albite may hold both potassium atoms and calcium atoms in subordinate amount.

The composition of any feldspar is conveniently expressed by the use of the symbols Or, Ab and An for the pure end-members; thus Ab_{32} An_{68} would be a plagioclase falling in the labradorite section of the series; and Or_{26} Ab_{66} An_8 is the composition of a possible anorthoclase.

The feldspars have all been produced synthetically. Plagioclase holding more than 20 per cent of anorthite will crystallize from a simple melt of its constituents, but soda-rich plagioclase, as well as pure albite and pure orthoclase, can only be induced to crystallize by the addition of a flux. The melting point of pure anorthite is 1553±2° C. (Osborn, 1942), that of albite 1115°±5° C. (J. F. Schairer). Orthoclase dissociates at 1150±20° C. into leucite and a siliceous liquid (Bowen and Schairer, 1947). Under hydrothermal conditions, orthoclase has been synthesized at 360°, albite at 450°, and anorthite at 470° (Morey and Ingerson).

THE PYROXENE FAMILY

Structurally, the pyroxenes consist of endless chains of linked silicon-oxygen tetrahedrons, bound together laterally by magnesium and calcium atoms which are joined to oxygen, not directly to silicon. Since each silicon atom is joined to four oxygen atoms, and each oxygen either to another silicon or to a metal atom, the typical formula is either $MgSiO_3$ or $CaMg(SiO_3)_2$. The compound $CaSiO_3$, although it occurs in nature as the mineral wollastonite, has a different atomic structure from the pyroxenes; for this reason the number of calcium atoms in a pyroxene is rarely more than the number of magnesium atoms, though it is often less. The possible kinds of atomic substitution in this family are (i) Fe'', Ni or Mn for Mg; (ii) Al_2 or Fe'''_2 for Mg-Si; (iii) Na-Al, Na-Fe''', or Li-Al for Ca-Mg.

The following are the simple pyroxenes:—

$R''SiO_3$	1.	$MgSiO_3$	Enstatite
	2.	$FeSiO_3$	Ferrosilite (theoretical)
	3.	$MnSiO_3$	Rhodonite
$R''R''(SiO_3)_2$	4.	$CaMg(SiO_3)_2$	Diopside
	5.	$CaFe(SiO_3)_2$	Hedenbergite
$R'R'''(SiO_3)_2$	6.	$NaAl(SiO_3)_2$	Jadeite
	7.	$NaFe(SiO_3)_2$	Ægirine (acmite)
	8.	$LiAl(SiO_3)_2$	Spodumene

The three kinds of ionic substitution mentioned above give rise to four substitution series, as follows:

A. The *Orthopyroxene* or *Hypersthene* series, ranging from enstatite to ferrosilite. A little manganese may be present, but much manganese makes the structure triclinic. The enstatite and the hypersthene ordinarily found in rocks are rhombic in crystallization, but at high temperatures (1140° C. for $MgSiO_3$, below 955° C. for varieties rich in $FeSiO_3$) they transform into monoclinic varieties known as clino-enstatite and clino-hypersthene; these are rarely found in eruptive rocks but are well-known in meteorites. Pure $FeSiO_3$ is unknown in the crystalline state.

B. The *Augite* series. Common augite is merely a moderately aluminous variety of diopside-hedenbergite, formed by the partial substitution of Al_2 for Mg-Si. The proportion of alumina in common augite is from 3 to 9 per cent, and some ferric oxide or chromic oxide may accompany it.

C. The *Enstatite-Augite* series, so named by W. Wahl in 1907, includes pyroxenes which are intermediate between series A and series B. A distinguishing feature of this series is that as the content of enstatite increases the optic axial angle diminishes to zero and then opens out again at right angles to its former position. A pyroxene with a very small axial angle was described by A. N. Winchell in 1900 and named "pigeonite" (from Pigeon Point, Minnesota); it has now become the practice to extend this name to all monoclinic pyroxenes of small axial angle, that is, to the more magnesian members of the enstatite-augite series. H. Kuno (1) put the limit at 45°, all monoclinic pyroxenes of smaller axial angle than this being called pigeonite, all others augite; but H. H. Hess has given reasons for preferring the limit $2V=32°$.

T. F. W. Barth (2) contends that pigeonite is only formed under conditions of rapid cooling, as in basalt, and that under slow cooling hypersthene and augite crystallize separately, as in gabbro and norite; pigeonite is therefore regarded as a metastable mineral. Hess and Phillips have shown that orthopyroxene in plutonic rocks often contains thin lamellæ of clino-pyroxene arranged parallel to the optic plane; they explain this by exsolution of diopside from an originally homogeneous pyroxene (diopside-hypersthene). But H. Kuno maintains that iron-rich pigeonite is stable even under plutonic conditions.

D. The *Soda-Diopside* series. This results from the substitution, in the diopside structure, of Na-Al or Na-Fe''' for Ca-Si or Mg-Si, forming jadeite-diopside and acmite-diopside respectively. Jadeite-diopside is almost restricted to metamorphic rocks, but the acmite-diopside series, often called "ægirine-augite," is well known in the alkaline eruptive rocks.

All the pyroxenes have been prepared synthetically, although Bowen and Schairer found it impossible to induce a pure metasilicate melt to crystallize completely in the rhombic form. By using calcium fluoride as a flux, D. P. Grigoriev (1) succeeded in lowering the freezing-point of a magnesium metasilicate melt so far that rhombic enstatite crystallized out. Clino-enstatite and clino-hypersthene, as well as diopside and hedenbergite, crystallize readily from dry melts. Diopside melts at 1391° C. (Bowen); clino-enstatite melts incongruently at 1557° C. (Bowen and Andersen). Acmite has been synthetized by fusing its constituents along with an excess of NaCl; it melts at 990° C. (Bowen and Schairer, 1).

THE AMPHIBOLE FAMILY

The amphiboles are chain compounds, like the pyroxenes, but with this difference that the amphibole chain is a double one. Chemically, magnesium always exceeds calcium, and hydrogen or hydroxyl ions are always present; thus the formula of an amphibole is much more complex than that of the nearest pyroxene.

Corresponding to enstatite there is *anthophyllite*, which may be written $H_2Mg_7(SiO_3)_8$ or $(OH)_2Mg_7Si_8O_{22}$, and a series of *ortho-amphiboles* analogous to the orthopyroxenes. These seem to be confined to metamorphic rocks.

The place of diopside is taken in the amphibole family by *tremolite*, $H_2Ca_2Mg_5(SiO_3)_8$, or $(OH)_2Ca_2Mg_5Si_8O_{22}$. In this formula calcium is replaceable by sodium; magnesium by ferrous iron or manganese and also in part by aluminium;

silicon in part by aluminium; and aluminium by ferric iron. Titanium and fluorine may also enter the structure. The product of all these substitutions is the *hornblende series,* analogous to the augite series but considerably more complex. The mineral that we call common hornblende has really a very wide range of composition. Most hornblende is green, but there is a dark brown variety that has long been known as "basaltic hornblende." It has been shown by Kozu and Yoshiki that this brown variety is formed when common green hornblende is heated above 750° C. The transformation is accompanied by oxidation of iron and loss of water.

From the tremolite formula, by the usual substitutions, we may deduce a series of *soda-amphiboles.* The best known of these are glaucophane, $(OH)_2Na_2Mg_3Al_2Si_8O_{22}$, and the related riebeckite and arfvedsonite in which the place of magnesium and aluminium is taken by ferrous and ferric iron. Ænigmatite, another sodium-iron silicate, is usually placed among the amphiboles but is now known to have a different atomic structure.

V. von Chrustchoff is reported to have obtained amphibole crystals by heating silica, alumina, and various hydroxides to 550° C. in presence of water; but a supposed amphibole obtained by Allen and Clement under similar conditions has since been shown to be olivine. D. P. Grigoriev (2) has however prepared a fluorine hornblende, that is, one in which the place of hydroxyl is taken by fluorine, from a dry melt.

THE MICA FAMILY

X-ray study has shown that the unit cell of the mica structure contains twelve oxygen atoms, two of which belong to hydroxyl groups. The formula of *muscovite* is accordingly $KAl_3Si_3O_{10}(OH)_2$. The substitution of sodium for potassium gives *paragonite,* and the partial substitution of lithium for aluminium gives *lepidolite* and other lithia micas. In *phlogopite* two atoms of aluminium are replaced by three atoms of magnesium, giving the formula $KMg_3AlSi_3O_{10}(OH)_2$; and in *biotite* there is a further substitution of Fe'' for Mg and Fe''' for Al. Lepidomelane is a variety of biotite very rich in both ferrous and ferric iron. Fluorine is a fairly constant minor constituent of most micas; it replaces hydroxyl and may make up as much as 7 per cent of some lithia-micas.

Both muscovite and biotite have been prepared in the laboratory with the aid of fluxes containing fluorine. The presence of water is not essential since the place of hydroxyl can be taken by fluorine; in this way D. P. Grigoriev (3) has prepared phlogopite from a dry melt. W. Noll has lately succeeded in synthesizing muscovite from a gel at the surprisingly low temperature of 225° C.; it is thought that it may form at even lower temperature in nature. When heated under atmospheric pressure the micas do not melt but dissociate, forming various anhydrous silicates of simpler constitution.

THE TOURMALINE FAMILY

The range of composition exhibited by members of the tourmaline family is exceptionally wide and the atomic structure is not yet known. M. J. Buerger gives the following formula for the colourless tourmaline of De Kalb, New York: $NaMg_3B_3Al_6Si_6O_{27}(OH)_4$. Tourmaline is the only common rock-forming mineral that contains boron, always amounting to about 10 per cent. of B_2O_3. Tourmaline has been synthesized by heating the necessary oxides together with water in a high pressure autoclave at 400° to 450°. F. G. Smith finds that tourmaline is stable in aqueous solutions of low alkalinity, but unstable in either strongly acid or strongly alkaline solutions. Iron-rich tourmaline melts incongruently about 1050°.

THE OLIVINE FAMILY

Common olivine is a substitution complex ranging in composition from *forsterite*, Mg_2SiO_4 to *fayalite*, Fe_2SiO_4. A small proportion of manganese is usually present. A lime-bearing olivine, *monticellite*, is of very rare occurrence in eruptive rocks. Olivine crystallizes readily from magnesium silicate melts which lack the silica necessary for the formation of enstatite. If excess of silica is present, forsterite may appear as a temporary phase, but it is later transformed into enstatite (the Bowen-Andersen reaction).

Forsterite melts at 1890° C. and fayalite at 1205° C. (N. L. Bowen and J. W. Schairer, 2). Greig, Merwin and Shepherd have shown experimentally that fayalite may be formed by reaction between iron oxide and hot vapours carrying silica; they consider that the fayalite that so often occupies cavities in acid lavas may have been formed in this way.

THE GARNET FAMILY

This family includes a number of isomorphous species which conform to the empirical formula $R_3X_2(SiO_4)_3$, where R stands for magnesium, calcium, ferrous iron or manganese, and X for aluminium, ferric iron, titanium or chromium. An artificial soda-garnet has been prepared, and soda has been detected in some varieties of garnet. Some of the simpler species are as follows :—

Grossularite	$Ca_3Al_2(SiO_4)_3$
Pyrope	$Mg_3Al_2(SiO_4)_3$
Spessartite	$Mn_3Al_2(SiO_4)_3$
Almandine	$Fe_3Al_2(SiO_4)_3$
Andradite	$Ca_3Fe_2(SiO_4)_3$

Common garnets contain the ions of two or three of these species at once. Grossularite is confined to metamorphic rocks, but the other species may occur scantily in eruptives. Pyrope and *melanite*—a titaniferous variety of andradite—are only known in rocks with a very low content of silica, the former in association with olivine, the latter with nepheline or leucite; almandine and spessartite on the other hand are found in very acid granites and pegmatites. Most garnets do not recrystallize as such after melting, but melanite and spessartite have been prepared from melts. Spessartite melts congruently at 1200° (R. B. Snow). Manganese and lime garnets were prepared by A. Gorgeu, by heating pipeclay to redness, along with manganese chloride or calcium chloride, in a stream of water vapour.

THE EPIDOTE FAMILY

This family appears, from its empirical formula, to be closely related to the garnet family. The simple lime-epidote (zoisite) has the formula $Ca_2(AlOH)Al_2(SiO_4)_3$. Common epidote differs from this only by having part of the aluminium replaced by iron. In allanite, or cerium epidote, aluminium is partly replaced by cerium, lanthanum, and didymium. Although epidote occurs quite abundantly in some granites, it is not strictly a magmatic mineral but a product of hydrothermal action in the late stages of cooling. The synthesis of epidote has not yet been effected.

THE MELILITE FAMILY

The following simple molecules enter into the composition of natural melilite:

Akermanite	$Ca_2MgSi_2O_7$
Iron-akermanite	$Ca_2FeSi_2O_7$
Gehlenite	$Ca_2Al_2SiO_7$
Soda-melilite	$Na_2Si_2O_7$

The general formula established by C. Warren is $(Ca,Na)_2(Mg,Al)(Si,Al)_2O_7$. J. R. Goldsmith finds that soda-melilite is soluble in gehlenite up to 15 per cent, but is not soluble in akermanite. Melilite crystallizes easily from dry fusion of its constituents; it is a common product in furnace slags. Melilite is only formed in eruptive rocks of extremely low silica content. It is nearly always accompanied by olivine and perovskite and frequently by nepheline or leucite.

THE NEPHELINE FAMILY

This family contains the species *nepheline,* to which the formula $NaAlSiO_4$ has generally been assigned, and *kaliophilite* or potash-nepheline, as well as *kalsilite,* a rare polymorph of the latter. It has long been known that natural nepheline contains some potassium, and Washgen and Buerger have lately found that the crystal structure of nepheline requires one alkali-atom in four to be potassium, so the formula of nepheline should be written $Na_3K(AlSiO_4)_4$. Some analyses of nepheline show more silica than this formula requires, and N. L. Bowen has proved that potential albite (as well as potential anorthite) may be present in natural nepheline. The composition of two examples of this mineral from East Africa is expressed as follows (Bowen and Ellestad),

	(1)	(2)
Soda-nepheline	61·8	57·0
Potash-nepheline	12·9	37·6
Albite	20·8	1·0
Anorthite	4·5	4·4

This rather surprising result is more easily understood if one writes the formula of nepheline thus, $Na_8Al_8Si_8O_{32}$; then the substitution of four atoms of calcium for eight of sodium gives $Ca_4Al_8Si_8O_{32}$, or anorthite, and the replacement of 4 NaAl by 4 Si gives $Na_4Al_4Si_{12}O_{32}$, or albite.

Nepheline has been synthesized by fusing its constituents together, with the addition of sodium tungstate as a flux. It has also been produced by various hydrothermal reactions; G. Friedel obtained it by treating muscovite with a ten per cent solution of sodium hydroxide at 280° C. Kaliophilite was prepared by G. W. Morey by heating potassium aluminate and silica with water at 600° C. under a pressure of about 1,000 atmospheres.

Nepheline transforms at 1248° C. into a cubic modification called carnegieite which, on cooling, does not revert to nepheline but develops a complicated twinning like that of plagioclase (Bowen and Greig).

Cancrinite is a substitution product of nepheline containing CO_3 groups, on occasion also SO_4 groups. Its composition may be expressed in an easily remembered form by the empirical formula $6\ NaAlSiO_4 + NaHCO_3$. Cancrinite was synthesized by J. Lemberg by heating kaolin in a solution of sodium carbonate.

In spite of a difference in crystalline form (isometric instead of hexagonal) the *sodalite group* of minerals should also be considered here. T. F. W. Barth (3) has shown that the formula of sodalite is $Na_8Al_6Si_6O_{24}Cl_2$ (more easily remembered as 6 nepheline + Na_2Cl_2) and that of nosean is similar, with SO_4 in place of Cl_2 (or 6 nepheline + Na_2SO_4). Haüyne, formerly regarded as a distinct species, was shown to be a solid solution between nosean and a calcium silicate—that is to say, some of the sodium atoms in nosean are replaced by calcium atoms and extra SO_4 groups are simultaneously built into the structure. Both sodalite and nosean were prepared synthetically by G. Friedel, by heating muscovite with sodium chloride or sodium sulphate solutions. T. F. W. Barth has prepared nosean by fusing a glass of the composition of nepheline along with sodium sulphate at 800° C.

LEUCITE AND ANALCIME

The empirical formula of leucite is $KAlSi_2O_6$. The formula of analcime is similar to that of leucite, with soda in place of potash, and with water of crystallization added; it is generally written $NaAlSi_2O_6 + H_2O$. J. Lemberg showed that leucite can be transformed into analcime, or analcime into leucite, by treatment of the powdered minerals with solutions of soda and potash salts respectively at 180° C. In spite of this simple relationship, natural leucite contains very little soda and analcime seldom more than a trace of potash. But Larsen and Buie have described a high-temperature analcime in which potash forms a quarter of the total alkali present. Leucite crystallizes readily from dry fusion, but analcime requires the presence of water under pressure, and was prepared by A. de Schulten by heating alumina with a sodium silicate solution at 180° C. Leucite melts at 1686° C. (Bowen and Schairer, 3).

Pseudo-leucite and *epi-leucite* are formed in certain rocks by the transformation of leucite (and perhaps analcime?). Pseudo-leucite is an aggregate of orthoclase and nepheline; in a further stage of transformation the nepheline may be replaced by white mica or zeolites. Epi-leucite, as described by A. Zavarickij, is an intergrowth of orthoclase and muscovite, both of which have been formed directly from leucite without the intervention of nepheline.

THE SILICA GROUP

Quartz is the stable form of silica under atmospheric pressure, at all temperatures up to 870° C.; tridymite between 870° and 1470° C.; and cristobalite from 1470° C. to its melting point at 1713° C. With slow cooling, and in presence of fluxes which ensure free movement of the molecules, both tridymite and cristobalite transform into quartz when the temperature falls below 870° C.; hence in deep-seated eruptive rocks only quartz is found. But the transformation is sluggish, and rapid cooling of the melt may delay it to such an extent that one or other of the high-temperature forms is perpetuated under conditions that are theoretically unfavourable to it. Thus one can understand why tridymite and cristobalite are often found in lavas, but never in deep-seated rocks. C. N. Fenner writes that :—

> "The presence of cristobalite or tridymite in a rock does not necessarily imply that at the time of formation of these minerals the temperature was above the respective inversion points (1470° and 870°). Any set of conditions which will bring together quantities of ungrouped SiO_2 molecules in such a manner as to favor their rapid assemblage in definite groupings without giving time for perfect equilibrium to be established (as in the reactions of vapors); or which will suddenly bring a system in which equilibrium prevails into new conditions, at the same time introducing obstacles to the establishment of a new equilibrium (as in the sudden chilling of a melt), will favor the deposition of unstable forms."

A. Lacroix claims that tridymite was formed, in the lavas of Mt. Pelée, by crystallization of the glassy base of the rock under the influence of heated water vapour; and A. F. Rogers has described an instance in California of the production of a feldspar-tridymite rock by devitrification of obsidian under the action of hot vapours. D. S. Beljankin and V. P. Petrov find cristobalite to be not at all uncommon in young dacitic and andesitic lavas of the Caucasus and Transcaucasia. E. K. Ustiev describes a dacite of the middle Caucasus, containing 7 to 10 per cent of tridymite and cristobalite; both minerals are believed to have been formed by recrystallization of the glassy base of the rock under the influence of magmatic gases. Van Valkenburg and Buie record octahedral crystals of cristobalite, deposited on fibres of mordenite in cavities in basalt. They conclude that the crystals must have formed "not far from 300° or 400° C." But H. Kuno (2) has observed tridymite and cristobalite in the groundmass of many Japanese lavas and considers these minerals to be of primary crystallization. The same opinion is expressed by A. G. Macgregor regarding tridymite in the lavas of Montserrat; and by E. S. Larsen, who has described lavas containing 20-30 per cent of tridymite in Colorado.

Quartz itself is known in two forms which differ slightly in appearance and crystallographic character, and have a transformation point at 573° C. The quartz in nearly all eruptive rocks has the crystal habit of high-quartz, although at atmospheric temperature it is actually low-quartz. In some coarse-grained granites and pegmatites low-quartz was present from the beginning. (See p. 55.)

Any form of silica can be converted into cristobalite by heating nearly to the melting point, without the aid of a flux. Both tridymite and cristobalite make their appearance in the silica bricks which are used for lining furnaces. To obtain quartz crystals in the laboratory, the agency of water as a flux is generally required, but Morey and Bowen (2) have obtained quartz bipyramids from an anhydrous sodium silicate melt at 750° C.

Under hydrothermal conditions quartz has been synthesized at 250°, tridymite at 350°, and cristobalite at 650° (Morey and Ingerson).

OTHER MINERALS

Oxides. The commonest of these is *magnetite*, Fe_3O_4, or FeO,Fe_2O_3. The ferrous iron in this formula is replaceable by magnesium or manganese, and the ferric iron by aluminium or chromium, giving a series of isomorphous compounds which are known collectively as *spinel*. Ferric iron is also replaceable to a considerable extent by titanium, giving titanomagnetite, which forms a connecting link between pure magnetite and the highly titaniferous *ilmenite*, $FeTiO_3$, or $(Fe,Ti)_2O_3$.

Hematite, Fe_2O_3, is a scanty constituent of many eruptive rocks and is perhaps in most cases an alteration product. The same remarks apply to *rutile*, TiO_2, and other oxides of titanium. *Corundum*, Al_2O_3, is an uncommon constituent of eruptive rocks, and its appearance, in certain cases at least, is due to solution of aluminous sediments by the magma.

Magnetite crystallizes directly from silicate melts and furnace slags. It has been melted in the laboratory at 1591° C., under a small pressure of oxygen (J. W. Greig and colleagues). Hematite is produced by the action of steam on iron salts at red heat; it is often observed in the slaggy parts of lava-flows, where the iron silicates of the magma have been decomposed by steam. The spinels have been crystallized by various processes involving the use of fluxes. Corundum crystallizes from dry silicate melts, and from direct fusion of alumina in the oxy-hydrogen blowpipe: it has also been obtained at temperatures as low as 400° C. by heating alumina with sodium hydroxide. The following melting points are known :—

Corundum	2040° C.
Chromite	2180° C.
Hematite	1565 °C.
Ilmenite	1450° C.

Zircon and Zirconosilicates. The principal minerals which contain zirconium, in eruptive rocks, are :—

Zircon, $ZrSiO_4$
Eucolite (eudialyte), perhaps
$(Na,Ca,Fe)_6Zr(OH,Cl)(SiO_3)_6$
Catapleiite, $Na_2Zr(SiO_3)_3+2H_2O$

also *wöhlerite, rosenbuschite, lavenite,* and others of doubtful composition. Zircon alone is of common occurrence, but eudialyte and others are important in certain classes of rocks. The melting point of zircon is 2550°.

Titanosilicates and Titanates. Many minerals have already been mentioned in which titania takes part. The following should be added :—

Sphene	$CaSiTiO_5$
Perovskite	$CaTiO_3$
Astrophyllite	$Fe_4Na_4Ti(SiO_4)_4$

Ænigmatite and *rhönite* are rare titanosilicates which were formerly referred to the amphibole group.

Phosphates. In addition to *apatite,* $3Ca_3P_2O_8,CaF_2$, which is found in traces in nearly all eruptive rocks, it appears that *xenotime,* YPO_4, is not uncommon in granites and syenites, where it is often mistaken for zircon. *Monasite,* a phosphate of the cerium group of metals, occurs in some granites. W. Mackie has found it to be widely distributed in the granites of the North of Scotland. The melting point of fluor-apatite is 1630° C.

Sulphides. *Pyrite,* FeS_2, and *pyrrhotite* (magnetic pyrites) Fe_5S_6 are common. Other sulphides, such as chalcopyrite, $CuFeS_2$; bornite, Cu_3FeS_3; pentlandite (Fe,Ni)S; and mispickel, FeAsS, are only occasionally present in significant quantities. The sulphides are slightly more fusible than the silicates. Pyrrhotite melts at 1183° C. (E. T. Allen and others), and chalcopyrite probably below 1100°. J. H. L. Vogt gives the following melting points:—Cu_2S, 1130° C.; FeS, 1170° C.

Fluorides and Fluosilicates. Among these are:—

Fluorite	CaF_2
Topaz	$Al(AlF_2)SiO_4$
Cryolite	Na_3AlF_6

Topaz and fluorite are often found in the marginal facies of granite masses,, where they replace feldspar. Cryolite is only known in some granite-pegmatites. Fluorite melts at 1378° (Kracek) and cryolite at 977°.

THE RELATIVE ABUNDANCE OF THE VARIOUS MINERALS AND ELEMENTS

F. W. Clarke has made an approximate estimate of the proportions in which the commoner minerals occur in eruptive rocks, with the following result:—

> Feldspars make up 59 per cent of all rocks examined. Metasilicates (pyroxenes and amphiboles) make up 17 per cent of all rocks examined. Quartz makes up 12 per cent of all rocks examined. Micas make up 4 per cent of all rocks examined. All other minerals together make up 8 per cent of all rocks examined.

If we compare intrusive with extrusive rocks, in respect of mineralogical composition, it will be found that such differences as appear relate only to the less common minerals. Thus tridymite and cristobalite are restricted to lavas, and leucite and melilite are nearly so restricted; while muscovite, tourmaline, and cancrinite are very seldom seen in lavas, whatever their composition. All other minerals occur indifferently in major intrusions, minor intrusions, and surface flows.

We owe to F. W. Clarke and H. S. Washington an elaborate computation of the average chemical composition of eruptive rocks. The first table gives the proportions of the commoner oxides; the second table that of the individual elements. In the case of those elements that amount to less than ·001 per cent, the approximate proportion is indicated by a cross in the appropriate position after the decimal point.

TABLE I

AVERAGE COMPOSITION IN TERMS OF OXIDES

(Clarke and Washington)

SiO_2	59·12	CO_2	0·101	Cr_2O_3	0·055
Al_2O_3	15·34	TiO_2	1·050	V_2O_3	0·026
Fe_2O_3	3·08	ZrO_2	0·039	MnO	0·124
FeO	3·80	P_2O_5	0·299	NiO	0·025
MgO	3·49	Cl	0·048	BaO	0·055
CaO	5·08	F	0·030	SrO	0·022
Na_2O	3·84	S	0·052	Li_2O	0·008
K_2O	3·13	$(Ce,Y)_2O_3$	0·020		
H_2O	1·15				99·984

TABLE II

AVERAGE COMPOSITION IN TERMS OF ELEMENTS

(Clarke and Washington)

1.	Oxygen	46·41	Rubidium	·000,X
2.	Silicon	27·58	Arsenic	,,
3.	Aluminium	8·08	Molybdenum	,,
4.	Iron	5·08	Tin	,,
5.	Calcium	3·61	Bromine	,,
6.	Sodium	2·83	Caesium	·000,0X
7.	Potassium	2·58	Scandium	,,
8.	Magnesium	2·09	Antimony	,,
9.	Titanium	0·720	Cadmium	,,
10.	Phosphorus	0·157	Mercury	,,
11.	Hydrogen	0·129	Iodine	,,
12.	Manganese	0·124	Bismuth	·000,00X
13.	Chlorine	0·096	Tungsten	,,
14.	Barium	0·081	Thorium	,,
15.	Sulphur	0·080	Columbium }	,,
16.	Chromium	0·068	Tantalum }	
17.	Zirconium	0·052	Uranium	,,
18.	Carbon	0·051	Silver	,,
19.	Vanadium	0·041	Selenium	,,
20.	Strontium	0·034	Platinum	·000,000,X
21.	Nickel	0·031	Tellurium	,,
22.	Fluorine	0·030	Gold	,,
23.	Cerium, yttrium	0·020	Thallium	·000,000,0X
24.	Copper	0·010	Indium	000,000,00X
25.	Lithium	0·005	Gallium	,,
26.	Zinc	0·004	Germanium	·000,000,000,X
27.	Cobalt	0·003	Radium	·000,000,000,00X
28.	Lead	0·002		
29.	Boron	0·001		
30.	Beryllium	0·001		

Later computations, which have been summarized by E. Tröger, suggest that some of the minor elements—particularly beryllium, lithium and boron—may be less scarce than these figures suggest. An alternative estimation has been made by T. F. W. Barth (4).

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> *Knowledge is always being born and dying; it is a tree which pushes new branches while others wither and drop. We have to be on our guard against dead and petrified knowledge bequeathed to us by former generations.*—Sir Walter Raleigh.

CHAPTER III

THE FUGITIVE CONSTITUENTS OF THE MAGMA

By fugitive constituents we mean substances which were present in the magma before freezing set in, but were for the greater part lost during the process of crystallization, so that they do not commonly appear as rock constituents. These substances may be gases and vapours which escape from the magma when the external pressure is reduced, or soluble salts which are carried away in solution by magmatic or even meteoric water.[1]

The evidence relating to the fugitive constituents of the magma is drawn from four sources; first, from observation of active volcanoes and fumaroles in the field; second, from the chemical and microscopical study of the liquids and gases enclosed or occluded within eruptive rocks and rock minerals; third, from the study of contact metamorphism; fourth, from experiments upon the conditions of formation of crystalline silicates.

The most satisfactory investigations of volcanic gases that have ever been made are those of A. L. Day, E. S. Shepherd and T. A. Jaggar at Kilauea. The gases were collected by thrusting iron pipes, or vacuum tubes with soft glass plugs, into cracks in the domes which formed above lava fountains. All possible care was taken to prevent contamination of the

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[1] N. L. Bowen has introduced the term *hyperfusible components,* and this, in the abbreviated form "hyperfusibles," is coming into popularity. It does not, however, have such a wide connotation as "fugitive constituents" since compounds such as sodium metasilicate, sodium aluminate, and the chlorides and carbonates of the alkalis cannot fairly be said to be "excessively fusible." Both terms are useful, but they should not be confused.

gases by air, but the difficulty of collecting flaming gas is so great that some degree of contamination seems to have been unavoidable. In the first collections by Day and Shepherd, water was not estimated, owing to the observers being unprepared for its appearance; nevertheless a large proportion of water condensed in the collecting tubes. In later collections by Shepherd and Jaggar, arrangements were made for determining water. A few representative analyses from the 1917 and 1919 collections are quoted below. The figures refer to volumes per cent at 1200° C.

CO_2	CO	H_2	N_2	A	SO_2	SO_3	S	Cl_2	H_2O
47·63	1·46	0·48	2·41	0·14	11·15	0·42	0·04	0·04	36·18
11·12	3·92	1·42	—	0·51	—	—	8·61	0·02	77·60
2·65	1·04	4·22	23·22	n.d.	0·16	—	0·70	n.d.	67·99
17·95	0·36	1·35	37·84	n.d.	3·51	—	0·49	n.d.	38·48
33·48	1·42	1·56	12·88	0·45	29·83	—	1·79	0·17	17·97
6·63	0·22	0·15	2·37	0·56	3·23	5·51	—	1·11	80·31
5·79	—	—	7·92	n.d.	4·76	2·41	—	4·08	75·09
1·42	0·05	0·08	0·68	0·05	0·51	—	0·07	0·03	97·09

In summarizing the results obtained between 1912 and 1919, Shepherd points out that the principal emanation from Kilauea is water, which averages about 70 per cent of the total gas evolved. Next in abundance is carbon dioxide, with sulphur dioxide in the third place. Chlorine is relatively scarce, and hydrocarbons are absent or inappreciable. The ratio of argon to nitrogen is about three times as great in the volcanic gases as it is in air.

All the work that has been done on the subject goes to show that the composition of the emanation varies greatly from place to place and from time to time. This is plainly shown even by the small number of analyses quoted above. That this must be the case is evident when one considers that such gases as hydrogen and carbon dioxide, hydrogen and sulphur dioxide, react with one another at high temperatures; and further, that some of these gases are capable of reacting even with the silicates and oxides in lava; that, in fact, "whatever may have been the previous opportunities for chemical readjustment among the gases as they rose in solution with the magma and were gradually set free with diminishing pressure, they are still in process of active reaction when discharged into the air." (Day and Shepherd.)

The question must then arise, whether the water which is the chief and in many respects the most interesting component of the emanation was present as such in the lava, or was formed by oxidation of hydrogen at the surface of the lava lake. T. A. Jaggar has shown that air may be carried down into the lava by the foundering of blocks of the frozen crust, which carry air in their pores; thus oxidation may begin even before the gases reach the surface of the lake. But only a small proportion of the water in the emanation could have been formed by atmospheric oxidation, for if much water were formed in this way then the ratio of nitrogen to oxygen in the escaping gases would be much greater than it actually is. In a later paper, Jaggar admits that he formerly attached too much importance to atmospheric burning, and says "there is no slightest question that hydrogen monoxide is a magmatic product."

Water may also be formed by reaction between hydrogen and carbon dioxide, with production of carbon monoxide. But the latter gas is quite a minor constituent of the emanation, generally forming less than one per cent of the whole, so little water can have been produced in this way. In short, we seem to be compelled to believe, with E. S. Shepherd, that "at Kilauea the major constituent is water, and the question as to its meteoric or plutonic origin has been fairly well established in favor of the plutonic."

This conclusion receives strong support from the presence of firmly held water (as distinct from moisture given off at 100° C.) in all fresh eruptive rocks. From data collected by H. S. Washington, it appears that the average proportion of combined water in deep-seated rocks is about 0·8 per cent of the weight of the rock, and in lavas 1·6 per cent, while in certain glassy rocks (pitchstones) there may be as much as ten per cent of water by weight. J. D. Dana showed that the amount of water required to produce the beautiful "thread lace" scoria at Kilauea was at least 1·1 per cent by weight of the active lava. E. T. Allen calculated that if the water in average granite were entirely expelled at 100° C., under atmospheric pressure, it would occupy a volume thirty times that of the rock, and the water in average basalt under the same conditions would amount to eighty-seven times the volume of the rock from which it was expelled.

E. S. Shepherd and H. E. Merwin have studied the composition of the gases yielded by specimens of cold, solid lava when heated in vacuo. Lavas of Mont Pelée and of Hawaii were examined in this way, and Shepherd reports that the gases given off are almost identical with gases collected directly from the lava lake at Kilauea. He shows, furthermore, that the gases obtained from lavas, regardless of origin, all agree in this respect that water forms about eighty per cent of the total gas. As regards the source of this water, Shepherd establishes the very significant fact that the ratio of nitrogen to water in the gases is about 8 : 68, whereas if the nitrogen and oxygen had been derived from the atmosphere this ratio would have been 70 : 80.

It has long been known that pitchstones and basaltic glasses, which represent rapidly chilled magma, hold from five to ten per cent of water; and this has been regarded as strong evidence of the hydrous character of natural magma. But C. N. Fenner (1) has thrown some doubt upon this conclusion by showing that between the finely-crystalline Watchung basalt of New Jersey and its glassy selvedge there is not only a difference in water content, but also a marked difference in the ratio of soda to potash, and in the proportion of silica. L. Hawkes and H. F. Harwood have made a similar comparison between the central stony portion and the glassy marginal facies of a dyke near Sandfell, Iceland, with the result shown in the following table:—

	Glassy Margin.	*Stony Centre.*
	Percentage Composition.	
SiO_2	70·55	77·77
Na_2O	5·08	3·91
K_2O	2·58	4·34
H_2O	7·28	·62
	Normative Composition.	
Quartz	29·26	38·35
Orthoclase	15·23	25·68
Albite	42·92	33·06
Anorthite	1·78	1·39

Both Fenner and Hawkes attribute these differences to changes that have taken place in the composition of the glassy rocks since their formation. If they are right, then it would no longer be permissible to argue from the water-content of the glassy rock to that of the magma, and one of the oldest arguments in favour of the hydrous nature of the magma would be destroyed. But the question need not concern us greatly in view of the evidence derived from other sources.

An extraordinarily important series of experiments by R. W. Goranson has at last given us a direct measurement of the solubility of water in granitic magma. Goranson prepared a granitic glass by fusing granite from Stone Mountain, Georgia, and cooling it rapidly. Samples of this glass were sealed up together with water in platinum capsules, which were strongly heated in bombs containing carbon dioxide (to serve as the pressure-medium). After two to four hours the temperature was lowered quickly, and the water content of the samples (again in the glassy state) was determined by ignition in air. The solubility of water in the granitic glass at 900° C. was found to be as follows:—

Pressure (Metric atmospheres)	500	1000	2000	3000	4000
Corresponding depth (kilometres)	2	3·8	7·6	11·4	15
Solubility (weight per cent)	3·75	5·75	8·15	8·90	9·35

When the pressure was fixed at 980 metric atmospheres and the temperature was varied, the following data were obtained:—

Temperature	600	750	800	900	1000	1200
Solubility	5·96	6·30	5·94	5·70	5·50	4·83

At this pressure, the solubility of water in molten granite decreases at the rate of about 0·3 per cent for each rise of 100° C.

It is known from stratigraphic data that some bodies of granite crystallized at depths of 4,000 to 12,000 feet (say 1 to 4 kilometres); and geological thermometer studies (see chapter IV) have shown that the temperature of deep magma probably never exceeds 870° C. or thereabouts. It follows from Goranson's experiments that natural granitic magma, at such depths and pressures, may hold between six and nine per cent of water. This is a conclusion of the greatest importance to all students of rocks and ore-deposits.

The solubility of water in basaltic magma has not yet been determined. T. C. Phemister has discussed the matter, basing his argument on the study of the artificial system water—potassium-metasilicate—silica, which was examined experimentally by G. W. Morey and C. N. Fenner. Phemister concludes that the water-content of deep-seated basalt may be in the neighbourhood of five per cent. This has only the validity of a reasonable extrapolation, but in view of Goranson's data and Sir Charles Parsons' demonstration of the solubility of water in molten olivine, there is no reason to question Phemister's estimate.

Besides the gases that we have discussed, various solid substances condense and form incrustations about volcanic vents and fumaroles. These may be direct sublimates from the magma, or they may result from the decomposition of gases or vapours, or even from reaction between the volcanic gases and the country rocks. Several analyses of such deposits are given by A. Brun.

Vesuvius, eruption of 1906		*Chinyero, eruption of 1909*		*Vesuvius, from cinders of cone*	
Al_2Cl_6 (with Fe,Mn)	9·80	NH_4Cl	79·35	$FeCl_3$ (with Mn)	1·99
NH_4F	6·12	$FeCl_3$	1·86	$MgCl_2$	4·25
NH_4Cl	85·02	NH_4F	9·46	$MgSO_4$	21·96
HCl	0·09	CaF_2	1·21	NaCl	31·50
SO_3	0·15	SiF_4	3·67	KCl	37·39
$CaCl_2$	trace	HCl	3·72	NH_4Cl	1·50
SiF_4	trace	Al_2O_3	trace	H_2O and free acid	1·41
	101·18		99·27		100·00

The incrustations formed about the fumaroles in the Valley of Ten Thousand Smokes, in Alaska, since the eruption of June, 1912, were studied by J. W. Shipley and E. G. Zies. The gases that issue from these fumaroles, at temperatures up to 645° C., are often strongly acid; sulphur trioxide, hydrochloric acid and hydrofluoric acid have all been detected in them. The most characteristic deposit of the fumaroles, according to Shipley, is amorphous silica, which reacts for fluorine and is believed to have been formed by decomposition of silicon tetrafluoride. Another common type of incrustation is a mixture of gypsum or anhydrite with salt and alum. Ammonium chloride is abundant; it constituted 98 per cent of the incrustation in one case. Potash alum is also common. Magnetite and red ochre (hematite) are found at almost every vent. Some crystals of magnetite were examined spectroscopically and chemically by E. G. Zies, and were found to contain the following metals in addition to iron:—

Lead	·005%	Molybdenum	·04
Zinc	·47	Manganese	·1
Copper	·23	Nickel	·01
Tin	·004	Cobalt	·02

Zies states that lead, copper, tin, and zinc are common constituents of the incrustations obtained from many of the fumaroles in the Valley.

Further evidence regarding the composition of the fugitive constituents of the magma has been obtained by studying the solutions and gases which are sometimes trapped within crystals of quartz, topaz, and other rock-forming minerals. The pioneer work of Humphrey Davy, Brewster, and other early workers in this field has lately been supplemented by quantitative data. H. Faber has determined the proportions of water and dissolved salts enclosed in crystals of pegmatitic feldspar and quartz with such results as these:

In feldspar—water 0·16% containing 18·2% of dissolved salts.
In quartz —water 0·15% containing 18·8% of dissolved salts.

The dissolved salts were chiefly chlorides and sulphates of sodium and potassium, with some silica and a considerable excess of alkalis.

W. H. Newhouse has studied the liquids enclosed in crystals of galena and zinc-blende in the Mississippi valley, and finds them to be concentrated solutions of sodium chloride and calcium chloride. These observations bring to our notice the presence of a class of fugitive substances in the magma which are not appreciably volatile at magmatic temperatures but are readily soluble in water. But for the accidental trapping of small quantities of these substances within growing crystals, we might have remained in ignorance of the presence of simple salts of soda and potash in the magma, since their great solubility must cause them to be leached out of the rocks and carried away by circulating ground-water. One might expect to find traces of these simple salts retained in glassy rocks. The matter has not been specially investigated, but J. B. Ferguson found in an obsidian from Iceland quantities of combined chlorine and sulphuric anhydride equivalent to 0·23 per cent of sodium chloride and 0·14 per cent of sodium sulphate. A slightly soluble fluoride of sodium (the mineral villiaumite) was found by A. Lacroix in the foyaite of Los Islands (Islas de los Idolos), West Africa. It may be added that Zünckel and Hempel have shown experimentally that silicate melts can dissolve as much as 0·86 per cent. of sodium chloride.

The possible importance of sodium carbonate in magmatic emanations is suggested by the presence of soda-springs and soda-lakes in some volcanic regions, notably East Africa. Lake Natron and Lake Magadi yield the mineral trona ($Na_2CO_3.NaHCO_3$) in marketable quantities; and C. W. Hobley reported that after the eruption of Oldoinyo N'Gai in 1917 "the country for miles around was covered with a pall of grey volcanic ash and soda," which extended as far as twenty-five or thirty miles from the volcano.

Some magmas seem to hold soda in the form of sodium metasilicate, because analyses of soda-amphiboles and soda-pyroxenes often show an excess of soda and potash molecules over those of alumina and other sesquioxides. For example, an analysis of obsidian from Lake Naivasha, Kenya, by G. T. Prior shows 7·08 per cent of sodium metasilicate in the norm. The mineral pectolite, which is found in some highly alkaline rocks, is a double metasilicate of sodium and calcium with the formula $HNaCa_2(SiO_3)_3$. It seems, then, that the simple metasilicate Na_2SiO_3, which is readily soluble in water, may be present in the magma under certain conditions.

Sodium borate (borax) and boric acid (HBO_2, passing over into H_3BO_3) are other water-soluble compounds found in volcanic emanations, especially in Tuscany and California.

Additional evidence relating to the fugitive constituents of the magma is to be found in the contact zones surrounding large bodies of eruptive rock. Studies of contact metamorphism often demonstrate a wholesale transference of silica, alkalies and iron, as well as chlorine, fluorine and sulphur, from the magma to the country rock. Limestone contacts are more instructive than those which involve siliceous or aluminous sediments, because chemical reactions take place between the escaping substances and the limestone, by which the former are retained in the zone of contact alteration.

In the metamorphic limestone blocks of Monte Somma, Italy, the following minerals containing chlorine, fluorine, sulphur or phosphorus are found :—

Anhydrite (S)
Apatite (P,F,Cl)
Chondrodite (F)
Clinohumite (F)
Cuspidine (F)
Fluorspar (F)
Galena (S)
Haüyne (S)
Humite (F)
Lazurite (S)
Microsommite (Cl,S)
Molybdenite (S)
Pyrrhotite (S)
Scapolite (Cl)
Sodalite (Cl)
Vesuvianite (F)
Zinc Blende (S)

Other well-known minerals produced in contact metamorphism, which contain elements foreign to ordinary sedimentary rocks, are :—

Tourmaline (B,F)
Axinite (B)
Sphene (Ti)
Perovskite (Ti)
Koppite (F)
Scheelite (W)
Datolite (B)
Topaz (F)

It may seem that undue stress is laid upon minerals that are, after all, of comparatively rare occurrence even in contact zones. But the significance of these minerals does not depend on their abundance, but on the unusual elements which they contain and which can only be traced to emanations from eruptive magma. As regards actual abundance, other minerals, such as garnets and pyroxenes, usually predominate greatly, in contact zones, over any of the minerals named above; but then the proof that they owe any of their components to emanations from eruptive magma is not so obvious as it is in the above cases. Nevertheless, the transference of silica, iron and alkalies from magma to its country rock can often be demonstrated on an enormous scale, and to a considerable distance. A noteworthy case was studied by W. Lindgren (1) at Clifton, Arizona, where a very pure limestone has been altered into a garnet-pyroxene rock through a contact zone two thousand feet wide. The transformation involves large additions of silica and iron to the limestone. It is interesting to note that the quartz crystals, in the granite which was the cause of the metamorphism, are full of microscopic enclosures of a strong solution of common salt containing an iron compound. Another eruptive rock in the same neighbourhood, which has no such enclosures, has effected very little metamorphism of the limestone. One is forced to believe that the cause of the metamorphism was the escape from the magma of aqueous solutions containing silica and iron.

In a similar case at Bingham, Utah, Lindgren (2) has shown that the limestone has gained silica, sulphur, iron, magnesium, aluminium, and sodium, the additions amounting locally to twenty-nine per cent of the volume of the limestone.

Striking evidence of the same kind was brought by J. F. Kemp from Tamaulipas, Mexico. In this mining district, highly acid quartz porphyries with a low content of iron cut very pure limestone. In the neighbourhood of the intrusions the limestone develops large quantities of aluminium-iron garnet. Kemp is driven to the conclusion that

> "from the intrusive rock has come either highly heated water-gas, or highly heated water itself in the closing stages, and that one or both of these have brought to the limestone the silica, iron oxide, and alumina for production of the silicates. After the production of the garnet and its associates was well under way, they brought in also the copper and iron sulphides which are the commonest ores."

In the Haliburton-Bancroft area of Canada, the contact alteration of limestone by granite is seen on an enormous scale. Great intrusions of gneissose granite invade the Grenville limestone series, and are surrounded by contact zones from a quarter to half a mile wide, in which the limestone has largely been altered into amphibolite (hornblende-rock). In an area of some 3,400 square miles, about one-fourth is underlaid by limestone and amphibolite, and the remainder by gneiss and amphibolite, the amphibolite making up about twenty per cent of the whole. The changes that have taken place in the contact zone are thus described by F. D. Adams and A. E. Barlow :—

> "When the granite-gneiss invades bodies of the limestone, the changes produced and the phenomena resulting from the intrusion are varied. The products of alteration may be divided into two classes :—(1) Alteration of the limestone into masses of granular pyroxene rock, usually containing scapolite, or into bodies of a fine-grained aggregate of scales of a dark-brown mica. (2) The intense alteration of the limestone along the immediate contact into a pyroxene gneiss or an amphibolite. The alterations of the first class may be considered as due to the heated waters or vapours given off by the cooling magma, that is, to be of pneumatolytic origin ; while the alteration products of the second class probably result from the more immediate action of the molten magma itself."
>
> "A comparison of the analyses shows that the granite at first transfuses into the limestone silica, alumina, oxides of iron, and magnesia, with some alkalies and a small amount of titanic acid. As the alteration progresses, all these constituents continue to increase in amount. But in these later stages of the alteration the alumina, oxides of iron, and alkalies are added in relatively greater proportion than the other constituents, while no further addition of magnesia or lime takes place, the proportions of these constituents remaining essentially the same, the carbonic acid escaping and carrying the rest of the lime with it. This means, speaking generally, that pyroxene and some scapolite were first developed in the limestone, and that later, felspathic constituents increased in amount, the calcite present being removed in solution."

W. G. Foye, studying the same area, showed by analysis of the amphibolite and the unaltered limestone that the material gained by the limestone from the granite magma had the following composition :—

SiO_2	60·4%
Al_2O_3	22·8
Fe_2O_3,FeO	8·8
MgO	1·7
Na_2O	5·0
K_2O	1·3
	100·0

When the rocks in a contact zone are shale and sandstone, the evidence of the addition of silica and alkalis is less conspicuous but equally real. V. M. Goldschmidt, in his great work on contact metamorphism in Norway, described the conversion of a shale containing less than one-and-a-half per cent of soda into a hornstone with five-and-a-half per cent of soda, in the contact zone of a small syenite stock of Aarvoldaas. More recently it has been shown that the shales around the Dartmoor granite in England have been enriched with both soda and potash. The normal shale is dominantly potassic, but bodies of shale which are enclosed in the granite are dominantly sodic. (A. Brammall and H. F. Harwood.) This matter is further discussed in chapter V.

F. F. Grout has studied the contact metamorphism of the slates of Minnesota by the Duluth gabbro. From the following table, which shows the composition of the average shale and of bodies of slate which were enveloped in the gabbroic magma, it is evident that not only soda but lime, magnesia and titania were added to the slate during metamorphism.

	Average Shale.	*Enclosures of slate in gabbro.*
SiO_2	58	50
Al_2O_3	19	20
Fe_2O_3	2	2
FeO	7	6·4
MgO	3	5·1
CaO	1	8·8
Na_2O	1·5	2·4
K_2O	3·3	1·3
TiO_2	·2	·8

Grout holds that "emanations from the magma passed through the sediments, adding some elements and carrying others away into the zone of groundwater circulation."

Remarkable evidence of the feldspathization of quartzite by magmatic emanations has been furnished by several observers. W. B. Wright, in his report on the geology of the island of Colonsay, described and photographed boulders of quartzite enclosed in the marginal hornblendic phase of a small syenite intrusion. The boulders, which are seen in all stages of corrosion, are surrounded by mantles of alkali-feldspar; in extreme cases the whole block of quartz has been replaced by feldspar without losing its original angular shape. This occurrence has been studied afresh by D. L. Reynolds, who has shown with the aid of chemical analysis that there has undoubtedly been a diffusion of alumina, potash and soda from the magma into the quartzite. Lime and iron oxides were found to be less diffusible and magnesia least of all.

The feldspathization of quartzite enclosures in the Bushveld Complex of the Transvaal was described by R. A. Daly. In this instance alumina and potash were the oxides mainly introduced into the quartzite.

FIG. 2. A siliceous inclusion in gabbro, surrounded by zones of plagioclase and hornblende. Sudbury, Ontario. (W. A. Jones.)

W. H. Collins presented striking field evidence of the feldspathization of large enclosures of quartzite, from one to six feet in diameter, by a diabasic magma at Blind River, Lake Huron. Some of the blocks have been converted entirely into a "light-colored, coarse-grained diabase" which either merges insensibly into the normal diabase or may be separated from it by a zone of hornblende. Other blocks retain cores of quartzite in every stage of transformation. Chemical data are lacking, but the feldspar of the enclosures is oligoclase or andesine and that of the normal diabase is labradorite, so it is clear that there must have been a supply of soda, at least, to the reaction zone.

W. A. Jones has studied bodies of plagioclase rock, with mantles of hornblende and cores of quartz, which are found in the gabbro of Sudbury, Ontario. These are clearly similar in nature and origin to the rocks described by Collins. (*Fig.* 2.)

A. Holmes has described xenoliths of transfused quartz in lavas of Uganda. The product of transfusion is a glass which is shown to have the composition of a silica-rich potassic obsidian. The constituents introduced were mainly alumina, potash and water, with small quantities or silica and other oxides.

THE IMPORTANCE OF THE FUGITIVE CONSTITUENTS

In reviewing the facts set out in these pages, we realize that the visible constituents of a rock represent only a part—the non-volatile and insoluble part—of the original complex mixture of substances that constitutes a magma: no doubt the larger part, and possibly ninety-nine per cent of the whole, but still only a part. The special importance of the remaining part, consisting of what we have called the fugitive constituents of the magma, lies in the influence that these constituents have upon the behaviour of the magma as a whole. This influence is seen especially in a lowering of the viscosity of the magma and a depression of the range of temperature within which crystallization begins and ends. The former effect depends upon the low melting-points of many of the fugitive constituents; the latter effect is the consequence especially of their small molecular weight.

An anhydrous silicate melt, especially if it contains feldspar, is extremely viscous. The presence of gases or easily fusible substances in the melt greatly reduces its viscosity.

This is an observation that has been made repeatedly, not only in the laboratory, but also in connection with the flowing lavas of active volcanoes. If the lava, after cooling and losing its gases, is reheated to the temperature that prevailed in the crater or lava-sheet, it does not regain its former degree of fluidity. For example, A. L. Day observed that the lava in the crater of Kilauea is still capable of slight movement at temperatures of about 600° C.

A particularly neat demonstration of this fact has emerged from the studies of R. B. Sosman and H. E. Merwin on the intrusion temperature of the Palisade diabase in New Jersey. This rock has been intruded in the form of a sill through a series of shales and feldspathic sandstones, and it contains numerous enclosures of sandstone which show no signs of fusion. Yet experiment shows that the sandstone melts at about 1150° C., while the diabase is not readily fluid until the temperature rises above 1200° C. Clearly the diabase must have been fluid at the time of intrusion, yet its temperature was well below 1150° C.

The second function performed by the fugitive constituents is that of lowering the freezing range of the magma. It is well known that the effect of dissolving a salt in water is to lower the freezing point of the solution, the degree of lowering being proportional to the quantity of salt added. This is sometimes known as Blagden's Law, but it is better stated in the more generalized form given to it by Raoult:—

> "Every substance, solid, liquid or gaseous, when dissolved in a definite liquid compound capable of solidifying, lowers its freezing point." (Raoult-Blagden Law.)

To this may be added the further extension known as Raoult's Law:—

> "In a constant weight of a given solvent, all the physical molecules, whatever their nature, produce the same lowering of the freezing point."

What is true of other liquids in this respect is true of a silicate magma. The presence in it of the volatile substances which we have described has the effect of depressing the freezing point of the magma below that of a pure silicate melt. The amount of depression depends entirely on the mass of the substances in question: a quantity which, as we have seen, we possess no means of estimating. But since water, hydrochloric and hydrofluoric acids, and other fugitive constituents of the magma are compounds of low molecular weight, and since the effect depends only on the *number* of molecules present, it is clear that the depression of the freezing point may be much larger than the absolute quantity of these substances in the magma would lead us to expect.

G. W. Morey (1) has heated silicate glasses along with water in gas-tight bombs, and has obtained mobile liquids at temperatures considerably below the melting point of the dry glass. Many other illustrations of this effect will be given in a later chapter.

We have already discussed the fact that, on the whole, deep-seated rocks seem to have completed their crystallization at lower temperatures than lavas. It is not unreasonable to connect this with the greater concentration of fugitive substances in deep-seated magma.

A third function which the fugitive constituents perform is that of determining which of several possible minerals shall actually crystallize from the magma at a given stage. About half of the rock-forming minerals have the power of crystallizing from dry melts in the absence of any flux. The other half, either because they build very complex molecules or because the elements of water or another fugitive substance enter into their constitution, can only be induced to crystallize in the presence of water or a suitable flux. These two groups are commonly described as *pyrogenetic* and *hydatogenetic* (fire-formed, water-formed) respectively. The latter term must not be interpreted too literally, but only as meaning that the minerals to which it is applied crystallize more readily in the presence of water. The chief members of the two groups are named below.

Pyrogenetic.

Anorthite; all lime-rich feldspar
All pyroxenes except ægirine
Olivine
Nepheline
Leucite
Melilite
Tridymite
Magnetite
Ilmenite
Pyrite

Hydatogenetic.

Quartz
Orthoclase, Microcline
Albite, all soda-rich feldspar
All amphiboles
All micas
Most garnets
Tourmaline
Ægirine
Sodalite, cancrinite
Analcime

Whether pyrogenetic or hydatogenetic minerals are formed by the crystallization of a particular magma must depend to some extent on the concentration of the fugitive constituents in that magma. If freezing takes place under such conditions that these bodies are not permitted to escape, and therefore attain a high concentration in the residual magma, then the last minerals to be formed will be hydatogenetic, that is, a stage of pyrogenetic mineralization will be followed by one of hydatogenetic mineralization. But if freezing takes place with rapid loss of fugitive constituents, as it does in lava, then crystallization will be mainly of the pyrogenetic type, ending sometimes with the formation of an uncrystallizable glassy residue.

WATER AND VOLCANIC ACTION

G. W. Morey (3) has discussed the effect of the rise of vapour pressure that accompanies the crystallization of a melt containing volatile and non-volatile components. He shows the reality of the phenomenon by giving experimental data relating to the systems water—potassium-nitrate and water—potassium-metasilicate—silica. For instance, if potassium metasilicate at its melting-point is saturated with water at a pressure of one atmosphere, it takes up about one per cent of water, which is enough to lower its melting-point by 35° C. If the saturated liquid is cooled quickly it becomes supersaturated; the molten aqueous glass remains liquid until cooled several degrees below its melting-point. Then bubbles begin to form within the melt, which quickly swells into a pumiceous mass, increasing many times in volume and overflowing the crucible that contained it. Morey considers that these observations afford a complete analogy to the paroxysmal discharge of gases from volcanic vents.

R. W. Goranson (2) has since studied the systems albite-water and orthoclase-water, and shows that tremendous pressures may be produced by the crystallization of these systems. The system albite-water, at 1100° C. and 606 bars (equal to a depth of 2¼ kilometres) holds 4·2 per cent of water. On cooling the system, albite begins to crystallize at 960° C., and when the temperature has fallen to 819° C. more than half of the albite has crystallized, leaving a liquid that contains 9·5 per cent of water. The pressure is now 3,000 bars, and continued crystallization must raise it to five or six thousand bars. "There is here," Goranson writes, "a mechanism for developing all the pressure a volcanologist may desire."

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According as henceforward I shall have the means of making more or fewer experiments, I shall in the same proportion make greater or less progress in the knowledge of nature.—DESCARTES.

CHAPTER IV

TEMPERATURE AND PRESSURE IN THE MAGMA

WE have discussed the composition of eruptive rocks, both qualitatively and quantitatively; we have described the fugitive constituents of the magma, chiefly in a qualitative way; from the data so gathered we must form the best idea we can of the composition of a body of magma before freezing sets in. The magma, at this stage, is a solution of silicates of eight or more common metals, together with smaller quantities of other compounds such as titanates, phosphates, and sulphides, and with an unknown content of certain volatile and water-soluble compounds, including water itself and either free or combined carbonic acid.

After the composition of the magma, we are concerned to know its temperature, and the pressure under which it is confined within the earth-crust.

The temperature of the magma, when it finds its way to the surface and issues as lava, can be measured directly, although the process is attended by some difficulty and danger. Direct methods are not available for deep magma, so we fall back upon indirect methods depending upon certain mineral transformations. Before we can apply these satisfactorily, it is necessary to know the pressure under which the magma crystallized. Now the pressure that prevails in a magma-reservoir has three distinct components. The chief of these, unless the magma has climbed very nearly to the surface, is the weight of the overlying layer of rock. This may be estimated at about 400 atmospheres per mile below the surface. It is open to question whether any body of rock, now visible at the surface of the earth, crystallized at a greater depth than—say—three miles. Clearly, the greater the depth of consolidation, the more erosion must have taken place before the rock was laid bare, consequently the more incomplete the stratigraphical evidence by which alone the depth of consolidation can be judged. For this reason J. C. Ward was unable to decide between quantities so different as 12,000 feet and 30,000 feet, for the depth of consolidation of the granites of the Lake District of England. The roof of the Finmarken granite laccolite, in Norway, was about 1,600 metres thick, according to W. C. Brögger, who also says it is "absolutely certain" that the laccolithic granite of Drammen crystallized within 2000 feet from the surface. J. Barrell showed that the Marysville batholith of Montana lay originally within 4,000 feet from the surface; and Leith and Harder gave the same figure for the Iron Springs laccolite in Utah. D. L. Scholtz finds that the gabbro of East Griqualand, South Africa, crystallized beneath a load of 10,000 feet of sediments. It is likely, therefore, that in the case of all intrusions except perhaps the largest and most ancient batholiths, the pressure due to the weight of the roof did not exceed 1200 atmospheres in the upper part of the reservoir at the time freezing took place, and that it was generally a good deal less than that.

It must be said, however, that F. F. Grout has lately brought forward evidence that parts of the Saganaga granite batholith in Minnesota, now exposed at the surface of the earth, were at one time as much as twenty-five miles below the surface. The evidence is deduced from the structural features of the batholith and may perhaps be open to a different interpretation. P. Eskola (1) has contended that the greatest depth at which rocks of Archean age, now exposed at the surface, can formerly have lain is about 20 kilometres (12½ miles); for most rocks he believes the depths to have been much less.

The second component of the total pressure is the vapour-pressure of the dissolved gases (water vapour, carbon dioxide, etc.). So little is known of the concentration of gases in natural magmas that it is impossible to form a satisfactory measure of this component. If the concentration is low enough, the vapour pressure may be negligible in comparison with the pressure of the overlying rock; but when the magma begins to crystallize, the whole of the gases become concentrated in the diminishing residue of liquid, and the vapour pressure must rise steadily, unless the gas is able to escape by way of openings in the roof and walls of the reservoir.

The third component of the total pressure only comes into operation if the viscosity of the magma is so high as to prevent it from accommodating itself at once, by change of shape, to lateral pressure such as that which accompanies mountain-building movements. Since all available evidence indicates a high degree of fluidity in the magma, this factor will not be effective until the magma has nearly completed its crystallization.

If the pressure prevailing within a magma reservoir is difficult to estimate, the temperature is still more uncertain. Sorby's method, already referred to, was based on measurement of contraction undergone by the fluid enclosures in quartz crystals; it yielded values ranging from 300° to 350° C. This, as well as Lehmann's estimate of 500° C., refers to highly siliceous rocks such as granites and trachytes. The method of F. E. Wright and E. S. Larsen, which depends on the use of quartz as a "geological thermometer," points to the majority of the quartz-bearing rocks having crystallized between 573° and 870° C. The principle of the method may be explained briefly.

The mineral which we know as quartz varies slightly in crystalline form according to whether it was formed at a temperature above or below 573° C. The kind of quartz formed below 573° C. is known as α-quartz, and the high temperature variety as β-quartz.

> "If at any time in its history a particular piece of quartz has passed the inversion point and been heated above 575°, it bears ever afterwards marks potentially present which on proper treatment can be made to appear, just as an exposed photographic plate can be distinguished at once from an unexposed plate on immersion in a proper developer, although before development both plates may be identical in appearance." (Wright and Larsen.)

The critical points in the discrimination of the α and β forms are as follows:—

For the α form (low quartz): (1) presence of faces of the trigonal trapezohedron; (2) unequal development of the rhombohedral faces; (3) twinning lines regular and sharply marked; (4) intergrowths of right- and left-handed forms frequent and regular; (5) habit of crystals is prismatic; (6) absence of cracks.

For the β form (originally high-quartz, now transformed to low-quartz): (1) absence of faces of trigonal trapezohedron; (2) equally developed pyramid faces; (3) twinning lines very irregular; (4) intergrowths of right- and left-handed forms very irregular; (5) habit is pyramidal; (6) crystals are much cracked.

The granites and granite-porphyries which were tested by Wright and Larsen showed as a rule the presence of β-quartz; the only specimens of α-quartz were derived from pegmatite veins.

A. E. Fersmann has shown that in the microcline-quartz pegmatites of Mursinka, Ural Mts., the stems of the quartz crystals are (or were) β-quartz, but the heads of the crystals, where they project into cavities, were α-quartz from the beginning.

E. S. Bastin used the method of Wright and Larsen to determine the crystallization temperatures of the gem-bearing pegmatites of Maine; he reached the following conclusions:—

> " . . . . Although many of the finer-grained pegmatite masses and most of the graphic intergrowths of the coarser pegmatites crystallised at temperatures above 575°, the coarser and more siliceous portions—the portions characterised by the cavities, and hence presumably richer in gaseous or fluid constituents—crystallised at temperatures below 575°. The portions characterised by high and low temperature quartz are commonly so intimately associated in the same pegmatite mass that it seems unreasonable to assume great differences in the temperature of crystallisation of different portions. It is probable, therefore, that the whole mass of many of the coarser pegmatites crystallised not far from the inversion point of quartz; that is, not far from 575° C."

R. E. Gibson has studied the effect of pressure on this inversion point. At a pressure equal to a depth of 10 km. the inversion temperature was raised to 644°.

V. B. Meen has found both high-quartz and low-quartz in auriferous quartz veins in Canada.

E. F. Holden has shown that rose-quartz, a variety which occurs in some pegmatites and which owes its colour to manganese, is decolorized by heating to about 573° C., and that quartz formed above that temperature is colourless, even when it holds enough manganese to have given it a rose colour.

Quartz transforms at 870° C. into tridymite. Observation shows that while tridymite is a fairly common constituent of acid lavas, it never appears in plutonic rocks. It is found, too, that quartzose rock fragments which have been picked up by flowing lava often have their quartz changed to tridymite. Similar enclosures are very common in plutonic rocks, but in these the transformation is not observed. Thus it seems that we are confined to 870° C. as the upper limit of the temperature experienced, during the period of crystallization, by the more siliceous deep-seated magmas. But since there is a considerable difference in density between quartz and tridymite, the transition point ought to be raised by pressure. C. N. Fenner has discussed this point, but finds it impossible to form any reliable measure of the effect. V. M. Goldschmidt suggests that an increase of pressure by 400 atmospheres may raise the transition point by 100° C. or more, but Wright and Larsen think it probable that "pressure has but a slight effect in raising or lowering such an inversion point, and that whenever quartz appears in nature it was formed at a temperature below 870°." (800° in the original paper, but afterwards corrected to 870°.)

Larsen has since suggested that the inversion temperature may be varied about 100° C. by a pressure of 1,000 atmospheres, but this is no more than guesswork.

A further application of the "geological thermometer" principle is possible when limestone occurs in contact with an eruptive rock or among the fragments enclosed in the latter. Under moderate pressure the reaction

$$CaCO_3 + SiO_2 \Longleftrightarrow CaSiO_3\text{(wollastonite)} + CO_2$$

moves towards the right above 500°. The effect of higher pressures upon the temperature of reaction is shown in *Fig.* 3, after V. M. Goldschmidt. It will be seen that if the pressure under which a given rock crystallized can be estimated from stratigraphic data, then the presence or absence of wollastonite in the contact zone will enable one to set either a lower or an upper limit to the temperature of the magma at the place of contact. P. Eskola (2) points out that the association of quartz with calcite at such a contact proves conclusively that the transformation temperature was not reached, but the appearance of wollastonite does not prove the contrary, for it may be formed at temperatures below the transformation curve if silica meets lime in other form than the carbonate, and no free carbon dioxide is present. Eskola finds that wollastonite is rare in the contact-altered Archaean limestones of Finland, but quartz is often present. It should be noticed that the temperature indicated in this way is not the freezing temperature of the magma, or that of any of its constituents, but is the maximum temperature which prevailed at any time during which the magma and the limestone were in contact.

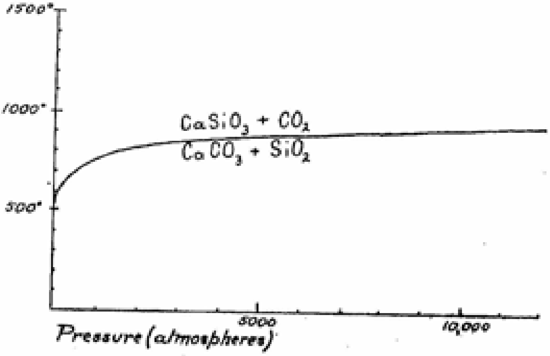

FIG. 3. The stability-curve of calcium carbonate and silica. (V. M. Goldschmidt.)

R. B. Sosman has described a peridotite dyke which cuts a coal seam in Pennsylvania and contains fragments of coal. From the behaviour of these fragments when heated, compared with that of coal from the same seam where it has not been affected by the intrusion, it is possible to deduce that the maximum temperature of the peridotite "magma" can hardly have exceeded 500° C. It is clear that the injected fluid was not melted olivine, and Sosman thinks it may have been "a mass of crystals lubricated by very thin films of a liquid which is not a silicate melt but is simply water, or a water-solution at a temperature far below the melting temperature of any rock."

C. W. Wright studied the production of garnets in the contact zones between granitic rocks and limestone. These garnets are often birefringent, and Wright reports that "experiments made with this garnet by Dr. Merwin show that it loses its birefringence after it has been heated for a few hours at 800° C., and does not regain it after several hours heating at 600°. This fact indicates that most of the metamorphism took place at temperatures lower than 800° C."

A further point on the geological thermometer scale is fixed by the irreversible transformation of wollastonite to pseudo-wollastonite, which takes place, according to the Geophysical Laboratory, at 1150° C. and is practically unaffected by pressure. Pseudo-wollastonite is unknown in nature, although it is commonly formed in furnace slags. Only wollastonite is formed by the contact metamorphism of limestones, from which we may conclude with a close approach to certainty, that the various magmas which have effected this kind of metamorphism in limestones were intruded at temperatures lower than 1150° C.

A similar conclusion is indicated by the study of the transformation nepheline-carnegieite, which takes place at 1248° C. N. L. Bowen writes that "in the event of the discovery of a natural carnegieite, its crystallisation above 1200° could be asserted with but little fear of error." But carnegieite has not been identified in nature, and Bowen concludes that most natural nepheline crystallized as such.

The dissociation of orthoclase into leucite takes place at 1170° C. (Morey and Bowen). This is a useful maximum for orthoclase-bearing rocks; and the discovery that natural leucite inverts and becomes birefringent at 603° C. affords a minimum for rocks containing birefringent leucite. Bowen and Schairer (2) have also shown that acmite (ægirine) melts incongruently at 990° C. This temperature must be lowered by the presence of feldspars, nepheline, and other minerals that are commonly associated with acmite in rocks. Later work by the same scientists (3) demonstrates that the rhombic pyroxenes crystallize between 1140° (pure $MgSiO_3$) and 955° C. (pure $FeSiO_3$). Since the rhombic pyroxenes are minerals of early crystallization, this observation sets an upper limit to the freezing range of a large class of rocks.

An upper limit to the range of magmatic temperatures is also indicated by observations made on the diabase (dolerite) sill of the Palisades of the Hudson, New Jersey, which is intrusive in a series of feldspathic sandstones and shales and holds many fragments of these. The melting temperature of these rocks was determined by R. B. Sosman and H. E. Merwin, who found that both the sandstone and the shale were more than half-melted after heating for seventy-five minutes at 1150° C. Yet fragments of these rocks which were completely enveloped by the magma retain their original structures and show no sign of having been fused, thus proving that the intrusion temperature of the diabasic magma was below 1150° C.

A. L. Day and E. T. Allen have ascertained that the biotite in the lava of Lassen Peak, California, begins to decompose at 840° C., and melts at 1050° C., under atmospheric pressure. The crystallization of the biotite must therefore have taken place at a temperature lower than 840° C., unless this transformation point is greatly affected by pressure.

A new and very valuable fixed point has been added to the geological thermometer by the work of three Japanese petrologists, S. Kozu, B. Yoshiki, and K. Kani, who have shown that common green hornblende transforms at 750° C. into a dark brown variety which is identical in its optical properties with the so-called "basaltic-hornblende," well-known in many basic lavas. This temperature is therefore the maximum at which the crystallization of green hornblende can have taken place in such rocks as rhyolite, trachyte, andesite. The further discovery that brown hornblende dissociates at about 1050° C. establishes a maximum temperature for all hornblendic rocks.

Very high estimates have sometimes been made of the temperature of flowing lava. J. Friedländer, for example, heated pieces of lava in the laboratory until they showed the same degree of luminosity as the active lava previously observed in the craters of Kilauea and Savaii, and concluded that the temperature of the latter must have been above 1600° C. The method is clearly a very gross one, but besides that, Day and Shepherd showed that combustion of gas is taking place all the time at the surface of the lava lake in Kilauea, so that the surface layer must be expected to show a considerably higher temperature than the lava underneath. T. A. Jaggar has since made actual measurements of the temperature of the lava at Kilauea, both at the surface and beneath it. He found that in lava fountains and domes where active combustion of gas was going on, the temperature rose as high as 1350° C.; but three feet below the surface of the lava the temperature was only 750° to 850° C. Jaggar's procedure was to put Seger cones of known melting point into steel pipes, and thrust these down into the active lava. His measurements, although still rather rough, are far more reliable than any of the earlier estimates.

G. W. Tyrrell has tried to estimate the intrusion-temperature of a basalt dyke on the island of Arran, Scotland, by the effect it has had on enclosed fragments of pitchstone. The latter rock contains well-formed crystals of quartz, orthoclase and andesine, in a glassy base. The orthoclase crystals have been partly melted; the andesines are only slightly affected; the quartz crystals are cracked but not melted. Pure orthoclase dissociates at about 1170° C.; andesine (Ab_3An) melts about 1375° C., so Tyrrell concludes that the temperature of the basalt was between these limits. The argument is scarcely conclusive, since the product of melting was not analysed—it might, for example, be a quartz-orthoclase eutectic, with a melting point considerably lower than that of pure orthoclase—so no great value is to be attached to these figures; yet the observations certainly indicate that the temperature of the basalt was higher than that of any known plutonic magma.

H. A. Brouwer has observed the melting of quartzite enclosures in the andesitic lavas of Merapi, Sumatra, and the conversion of quartz into tridymite. The melting point of the quartzite is not known, and it cannot be assumed to be the same as that of pure quartz, so the observations only prove that the temperature of the lava exceeded 870° C. A. Sigmund has recorded the melting of quartz grains in a magma-basalt at Furstenfeld, Austria. Each grain is surrounded by a mantle of yellowish glass, but clearly this is explained more readily by solution than by melting of the quartz, and it is not clear that the temperature was above the tridymite transformation-point. On the other hand, H. H. Thomas describes dykes in Mull, Scotland, which have attacked the quartz grains of invaded sandstones, and have given rise to tridymite fringes round the undissolved portions of the grains. With cooling, the tridymite reverted to quartz, but the shape of the tridymite crystals was retained, proving that the transformation point was passed.

An observation which seems to demonstrate an unusually high temperature in lava was made by A. Amstutz. An olivine-basalt at Dorgali, in Sardinia, encloses fragments of biotite-granite, which have been recrystallized with formation of leucite. If the leucite was formed at the expense of biotite, the temperature need not have been higher than about 1000° C.; but if it was formed from orthoclase, as is likely, then the temperature must have been in the neighbourhood of 1170° C.

V. M. Goldschmidt (2) has described enclosures of granite in lavas of Spitzbergen, in which the biotite has been melted, producing a pale brown glass full of needles of ortho-pyroxene. These lavas were accordingly hotter than the transformation point of biotite (840° C. or more), but not so hot as the dissociation point of orthoclase (1170° C.). A somewhat similar case was recorded by A. Knopf in Owens Valley, California. Granodiorite was invaded by basalt which was hot enough to dissociate biotite but not to dissociate orthoclase.

P. Ramdohr (1) has described enclosures of zinc-blende in basalt and has shown that the transformation temperature of zinc-blende into wurtzite, lying between 880° and 1020° C., was exceeded by the basalt. In another paper, Ramdohr (2) describes drop-shaped particles of metallic iron which are enclosed in basalt from Greenland; it is concluded that this basalt must have attained the temperature of 1145° C. at least.

Summarizing the matters discussed in this chapter, we may say that *there is an entire lack of evidence that any body of deep magma ever had a temperature higher than* 1170° C.*, or that it even approached that temperature; in very many cases the temperature does not seem to have exceeded* 870° C. In the case of surface magma (lava) the top layer may be heated up, by gas reactions and oxidation, to 1300° C. or more, but there is no reason to think that the body temperature of lava ever reached such a figure. It may be granted, however, that surface magma, and small bodies of magma in dykes near the surface, may show appreciably higher temperatures than deep magma.

The Geological Thermometer

Degrees C.	
1340	Calcite melts (under pressure of 1000 atmospheres of CO_2).
1300-	
1248	Transformation of nepheline→carnegieite.
1200-	
1175	Incongruent melting of orthoclase.
1150	Transformation of wollastonite→pseudo-wollastonite.
1140	Transformation of enstatite→clino-enstatite.
1100-	
1080	Dissociation of brown hornblende.
1000-	
990	Dissociation of acmite.
900-	
870	Quartz→tridymite inversion.
840	Dissociation of biotite (at Lassen Peak).
	Talc→enstatite + liquid under 1 atmosphere.
800-	
750	Transformation of green into brown hornblende.
700-	
600-	Minimum for leucite showing polysynthetic twinning.
575	Transformation α-quartz → β-quartz. Maximum for rose-quartz.
	Magnetite→hematite (in presence of oxygen).
500-	Wollastonite formed from $CaCO_3$ and SiO_2 under moderate pressure (12 atmospheres).
400-	Transformation marcasite→pyrite.

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They are ill discoverers who think there is no land, when they can see nothing but sea.—FRANCIS BACON.

CHAPTER V

THE MAGMA AND ITS WALLS

WE may now turn to the consideration of a body of deep magma which is exposed to such conditions of temperature and pressure as we have discussed in these pages. It is contained within a chamber whose walls are formed of rocks of familiar character; that is, of eruptive rocks such as granite and diabase, and of siliceous, aluminous and calcareous sediments, or the metamorphic equivalents of these. A question of fundamental importance now arises: to what extent is the composition of the magma modified by assimilation of the containing walls?

The answer to this question depends in the first place upon what we mean by assimilation. If we mean dissolution—that is, the passage of foreign solids into a liquid system which remains wholly liquid—then it seems that this can only take place to a small extent in nature, for it depends on the presence of superheat in the magma; that is, on the difference between the initial temperature of the magma and the temperature at which freezing sets in. From the data given in Chapter IV we are justified in concluding that deep-seated magma ordinarily has no great store of superheat.

It is not likely that any experienced petrologist who has discussed the assimilation of country rock by deep magma has ever thought of the process as one of simple solution or melting. One gathers that the prevailing idea, though often vaguely expressed, has been that of *reciprocal* action between the magma and the solid rock, by which the magma gained certain constituents while giving up others to the solid. The expression "osmotic reciprocal reaction" was applied to this process by H. J. Johnston-Lavis in 1894.

In an account of the Glen Dubh granite in the island of Arran, written in 1903, A. Harker (1) described dark enclosures in the granite which he declared to have been "totally reconstituted, not merely by metamorphism but by interchange of material between them and the surrounding magma."

J. J. Sederholm, too, when describing the assimilation of diabase by granite at Wiborg, Finland, showed that the granite was enriched with plagioclase and hornblende while the diabase enclosures were acidified by the granite magma. He summarized the matter in the following words: "The composition and structure of the rock show that its material has been derived from the neighboring basic rock by refusion or resolution, and that there has been a continuous adjustment of its composition, whereby reactions have also taken place between the neighboring minerals." It should be observed that although Sederholm uses the words refusion and resolution he clearly does not imply the disappearance of the solid phase.

A. Lacroix, in his account of the granite of the Pyrenees and its contact-phenomena (1898) appears to use the nouns dissolution and assimilation in the same sense; but then Lacroix, following in the footsteps of Michel-Lévy, links the modification of the granite so closely with that of the surrounding sediments that his idea seems to be very similar to the "reciprocal reaction" of Johnston-Lavis.

Petrologists are certainly to blame for having used such terms as solution, fusion, assimilation, and incorporation without fully explaining their meaning; but the references just made show that among the leaders of petrology the idea of a simple "melting in" of the solid phase was not generally upheld.

N. L. Bowen has clarified the idea expressed by Lavis, Harker and Sederholm and placed it on a sound chemical basis. He has shown that in many chemical systems, equilibrium between the solid and liquid phases can only be maintained by reaction between solid and liquid throughout the period of crystallization. Thus in the system albite-anorthite (page 93) the early formed crystals are rich in anorthite, but as cooling progresses these crystals react with the remaining liquid and become progressively richer and richer in albite until freezing is completed. Thus the plagioclase feldspars constitute a *continuous reaction series.* But if, with falling temperature, a phase ceases to be stable and transforms into something entirely different, as forsterite into enstatite or pyroxene into amphibole, then we have a *discontinuous reaction series.* The most complete example of such a series is given by the magnesian minerals

olivine→rhombic pyroxene→monoclinic pyroxene→
hornblende→biotite.

Each member of this series tends, as crystallization progresses, to transform into a later member which is in equilibrium with the changed conditions of temperature and concentration in the residual liquid. It follows that a liquid which is supersaturated with regard to one member of the series cannot *dissolve* any of that member but can *react* with it in such a way as to convert it into a later member of the series, with which, at the moment, the liquid is in equilibrium. In this process the solid phase will not disappear but it will be made over into a different solid phase; in short it will be metamorphosed; and the liquid phase will also suffer a change in composition. Thus a granitic magma, saturated with regard to biotite, cannot dissolve olivine, pyroxene or hornblende, but can and will react with these minerals, converting them into biotite. To effect this change *no heat is required,* for the crystallizing phase gives up as much heat as the disappearing phase takes in. Bowen calls this process "reactive solution and precipitation," and he considers it possible for a magma that lacks superheat to take up considerable quantities of foreign matter in this way.

It is clear that there are three processes by which foreign matter may be taken up by a liquid magma, namely

(1) Mechanical incorporation of solid particles without any chemical change;
(2) Reactive solution and precipitation; involving the replacement of one solid phase by another solid phase;
(3) Total dissolution, the solid phase disappearing altogether.

It is very desirable that these three processes should receive distinctive and unequivocal names. Loosely speaking, the term "incorporation" might be applied to all three of them, and "assimilation" to two; but much would be gained if we could agree to restrict "incorporation" to the first process, "assimilation" or perhaps "cross-assimilation" to the second, and "dissolution" to the third.

Instances of *mechanical incorporation* are numerous; they are to be found on a larger or smaller scale at almost any intrusive junction. The dolerite dyke shown in *Plate* II is crowded with angular lumps of quartz often several inches in diameter. In rising through the granite which forms the country rock, the dolerite appears to have followed a channel provided by a reef of quartz or pegmatite; and lumps of quartz and feldspar, breaking away from the walls, were enveloped in the rapidly cooling magma with little or no alteration. In other instances the enclosed grains are smaller and then their foreign origin is less obvious. Some basalts contain quartz grains which have been regarded as crystallizations from the magma, and indeed it is sometimes difficult to prove that they are not so. The kimberlite of the South African diamond mines contains rounded grains of garnet, enstatite, diopside, and other minerals which were at one time regarded as crystallizations from the magma that formed the kimberlite. Later studies have shown conclusively that these are foreign bodies picked up by the rising magma from the country rocks through which it broke its way.

A remarkable rock which was described by A. G. Högbom (1) in Angermanland, Sweden, consists of large anorthite crystals lying in a base of quartz and orthoclase. Högbom shows that this rock was formed by the invasion of a coarse-grained anorthosite (plagioclase-rock) by a granitic magma which insinuated itself between the large feldspars of the anorthosite and crystallized there. R. S. Cannon has described large crystals of andesine in a quartz-syenite of the Adirondack Mts., New York, and advances the same explanation.

Although mechanical incorporation has been the dominant process in these instances, and in many others that could be cited, yet some degree of reactive solution and precipitation can generally be demonstrated in addition, if only on a microscopic scale. The purely mechanical process possesses little interest, except in so far as it provides us with petrographical puzzles like those just described.

Evidence of *simple dissolution* of country rock is scanty, and it is confined to rocks formed at or near the surface of the earth. Although we may not assume much superheat in deep magma, yet the observations of T. A. Jaggar at Kilauea show that there may be considerable super-heating of surface lava in consequence of gas reactions. The temperature of the gases escaping from domes on the surface of the lava lake at Kilauea ranged from 1100° to 1350° C., and it was repeatedly observed that any increase in the quantity of gas discharged was accompanied by an increase in the surface temperature of the lake. This must be the explanation of the often-repeated observation that fragments of rock enclosed in active lava have sometimes been melted. A. Lacroix recorded the melting of granite by andesite on Santorin, forming a perlitic glass. The melting of granite in contact with basalt or diabase has been described by J. S. Flett, L. Hawkes (1), and A. Laitakari; in each case the action was on little more than a laboratory scale. Melting on a larger scale has lately been recorded by Larsen and Switzer in San Diego county, California. A mass of granodiorite measuring about 50 x 40 feet is enclosed in a plug of andesite and has been almost half melted. Alkali-feldspar and biotite have disappeared; hornblende has been transformed into augite; and the product is a rock that looks like obsidian and consists of crystals of quartz, plagioclase and augite set in a black glass. In this case there is no evidence of mixing between magma and melt; but in Owens Valley, California, A. Knopf was able to show that a body of olivine-basalt, intruded into granodiorite, has developed a base of clear brown glass by melting and dissolving the minerals of the granodiorite.

J. P. Iddings observed the converse process, melting and dissolution of solid basalt by liquid rhyolite, in the Yellowstone Park. C. N. Fenner (1) has made a close study of the latter phenomenon and has described a series of hybrid rocks resulting from it. One of these had a composition corresponding to 30 parts of basalt in 70 parts of rhyolite; another corresponded to 69 parts of basalt and 31 parts of rhyolite. Under the microscope the hybrid rocks showed corroded crystals of olivine, pyroxene and labradorite derived from the basalt, together with insets of quartz and orthoclase belonging to the rhyolite, all of these being scattered through a fine-grained groundmass of hybrid origin, composed of pyroxene, ore, oligoclase and orthoclase. It should be added that R. E. Wilcox believes the effect to be due to the interaction of two hot liquids rather than the solution of solid basalt by liquid rhyolite; but Fenner (2) gives strong field evidence in support of his contention. The difficulty of arriving at the true interpretation is brought out by L. Hawkes (2) who has tried to reconcile the two views. Similar observations were made by Fenner at Novarupta volcano in Alaska, but here the evidence of dissolution of basalt is even more complete for the product is largely a frothy glass having the chemical composition of an acid andesite. Fenner can offer no indication of the source of heat beyond the suggestion of exothermic reactions connected with the escape of gases from the rhyolitic lava.

It must be admitted, then, that simple melting and dissolution of foreign rocks is not impossible, although it is probably restricted to magma that has risen nearly or quite to the surface.

Evidence of *assimilation* of country rock by magma, in the sense defined on page 67, is widespread, although the recognition of this fact has been long delayed by the very character of the reaction. The tendency of cross-assimilation is to produce the same phases both in the magma and in the country-rock; thus it may be impossible to distinguish the products of reactive solution and precipitation from the products of normal crystallization. A further difficulty which arises out of this one is that in certain cases one does not know what the composition of the original intrusive material was, for the rock exposed at the surface, which passes for the unmodified eruptive, may itself be a composite thing. C. N. Fenner (3) has described some beautiful cases of admixture of granitic and metamorphic rocks in New Jersey. The purest variety of the granite contains little but microcline, albite and quartz; and the dark coloured metamorphic rocks, which form bands and indefinite streaks within the granite, consist of biotite, hornblende, quartz, microcline and albite. Fenner remarks that "in such occurrences as show bands of basic rock in largest amount, the adjacent granite contains the largest quantity of dark silicates, and where inclusions are rare, the granite is very light coloured and nearly free from ferro-magnesian minerals, and it seems probable that the invading magma was composed essentially of quartz and felspars, and that the dark minerals, even in massive granite, were derived to a large extent from the rocks which were invaded."

W. J. Miller has assumed a very extensive assimilation of country rock (metamorphic rocks of the Grenville Series) in order to explain the phenomena shown by the granite masses of the Adirondack Mountains, New York. He has, however, fallen into the common error of confusing "assimilation" with fusion, and he repeatedly speaks of "melting" of the gneiss when it would be more in accordance with modern views to speak of "assimilation." He tells us that the top of Prospect Mountain is formed of fine-grained, grey biotite-granite, which is very homogeneous except for occasional patches and stringers of Grenville gneiss. Passing southward down the mountain side, this rock shows fewer and fewer inclusions, passing into medium-grained biotite-granite and eventually into biotite-granite porphyry. Passing westward down the mountain side, however, the fine-grained granitic rock becomes coarser and holds more and more numerous and clearly-defined inclusions of Grenville gneiss, and passes eventually into biotite-garnet gneiss and quartzitic gneiss at the base of the mountain. As there is a perfect transition from the gray, granitic rock into the granite porphyry, on the one hand, and into Grenville gneiss on the other, it is held that the gray rock was formed by "fusion and incorporation" of gneiss by the granite-porphyry magma.

It seems clear that there is no necessity for the assumption of "fusion" in this case. The minerals of the gneiss are to a large extent the same as those of the granite, and the conjunction of cross-assimilation with mechanical incorporation should be adequate to explain all the phenomena. But Miller has certainly demonstrated a very extensive intermingling of magmatic with non-magmatic material in this region.

W. H. Collins makes out a strong case for extensive assimilation of sedimentary material by granite in the Onaping area, Canada. Batholiths of biotite-granite have invaded a

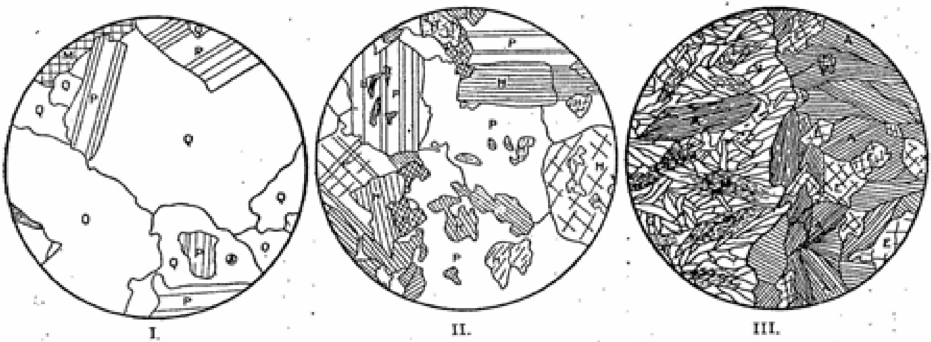

FIG. 4. Thin sections showing assimilation of ferruginous slate ("iron formation") by granite; Onaping, Canada. (W. H. Collins.) Magnification, 15 diameters. I, normal biotite granite; II, hornblende-plagioclase rock, two inches away from an enclosure of ironstone; III, actinolite-grünerite-epidote rock at the margin of the enclosure. [A=actinolite; B=biotite; E=epidote; G=grünerite; H=hornblende; M=magnetite; Mi=microcline; O=orthoclase; P=plagioclase; Q=quartz; T=titanite (sphene).]

complex of ancient schists which represent, in a highly metamorphic state, a system of slates, conglomerates, arkoses, tuffs and lava beds. The more usual types are chlorite schists, sericite schists, and "iron formation," or banded slates containing much magnetite and hematite. Where the biotite granite makes contact with the schists it is found to be full of blocks of the schist, showing all possible stages of disintegration, the contact itself "resembling in miniature an irregular, deeply embayed coast-line fringed with a myriad of islands and islets." The iron-rich inclusions are surrounded by dark areas of rock so rich in hornblende as to be nearly black, fading away outwards into normal biotite granite; these areas are clearly due to assimilation of fragments of iron-formation, as a result of which the biotite granite has transformed itself locally into hornblende rock. Near Burwash Lake many bodies of iron formation are enclosed in the eruptive, but the schists which normally lie above and below the iron formation have all disappeared except for swarms of partly digested fragments that lie in the granite. It is concluded that over an area of at least 50 square miles a thickness of schists measurable in hundreds, if not thousands, of feet was absorbed by the magma, and that the iron formation survived by virtue of being the most refractory material in the schist complex encountered by the granite. (*Fig.* 4.)

Many descriptions of interaction between magma and country rock fail to carry conviction on account of inadequate chemical study. It is therefore a pleasure to turn to the admirable description, given by A. Brammall and H. F. Harwood, of the Dartmoor granites and their contact phenomena, and to find that the precise petrographical work of one of the authors is supplemented by a large number of the accurate chemical analyses for which the other author is celebrated. The Dartmoor granite is intrusive in Paleozoic shales and grits, and the shale has been highly metamorphosed near the contacts. The following analyses clearly show that an exchange of certain constituents (cross-assimilation) has taken place between the granite and the shale. The numbers are molecular proportions.

	Average Granite	*Contact Granite*	*Contact Shale*	*Average Shale*
SiO_2	782	782	661	654
Al_2O_3	92	98	162	173
FeMn oxides	27	23	36	54
MgO	12	11	36	44
CaO	17	12	4	5
Na_2O	31	27	15	13
K_2O	35	42	58	51
TiO_2	4·8	3·6	6	7
P_2O_5	·9	2·3	3	2

To bring the contact-granite back to the composition of the normal granite it would be necessary to subtract from it a fraction amounting to 32 per cent of its mass and having the following (normative) composition:—quartz 34, orthoclase 45, albite 7, corundum 7, magnetite 2, enstatite 1 per cent.

The figures show that passage from shale to granite is characterized by a selective abstraction of the siliceous-aluminous constituents of the shale, which caused the masses of shale to disintegrate. The products of thermal metamorphism (andalusite, tourmaline and pinite) were incorporated in the granite magma, where they tended to be "made over" into mica. The granite itself shows a wide range of composition, as follows:—

Quartz	19 to	40 per cent
Biotite	2	49
Microperthite	19	42
Plagioclase	8	21
Anorthite in plagioclase	2	25
Colour index of rock	5	53

The general conclusion drawn by Brammall is that "the variation observed is attributable to the basification of initially acid magma while emplacement and differentiation were proceeding. The phenomena as a whole are consistent with the view that the initial magma was sodi-potassic (or even dominantly sodic), and that the potassic products . . . are differentiates from magma which assimilated shale at an early stage."

Among recent studies of reaction between magma and its wall rock, the description of H. H. Thomas and W. Campbell Smith of the xenoliths in the Trégastel granite, in Brittany, is of outstanding importance. The Trégastel granite is made up of microcline-microperthite and quartz, with subordinate oligoclase and biotite. It holds a great number of enclosures, some of which are obviously of sedimentary origin while others have an igneous aspect. The source of the latter is shown to be a body of olivine-norite which was invaded and cut up by the granitic magma. The least-altered norite consists of olivine and hypersthene, with subordinate biotite and plagioclase (bytownite) and a little apatite and ore. In the slightly altered norite hypersthene occurs to the exclusion of olivine; some monoclinic pyroxene makes its appearance, and the feldspar is labradorite. Other examples are found in which monoclinic pyroxene predominates and the feldspar is a still more sodic variety. In addition to these noritic and gabbroic enclosures there are others which might be described as dioritic or monzonitic in composition, consisting of orthoclase, quartz, zoned plagioclase of the mean composition of andesine, hornblende, biotite and sphene. The most acid type of all consists of quartz, orthoclase, acid plagioclase and abundant sphene. The granite itself, near the enclosures, is enriched with oligoclase, biotite, and hornblende. The evidence seems conclusive that the olivine-norite reacted with the crystallizing granitic magma and that the olivine was transformed into hypersthene and clinopyroxene, and the hypersthene again into biotite, just as the reaction principle requires.

A remarkable example of assimilation of gabbro by granophyre-magma at Carrock Fell, Cumberland, was described by A. Harker (2) in 1895, but still awaits chemical study. Passing outwards from the heart of the granophyre area, the rock becomes less and less siliceous and develops increasing quantities of augite, apatite and iron ores. The marginal facies of the granophyre, which is in contact with the gabbro, is a compact, dark rock with a specific gravity of 2·925, that of the normal rock being 2·670. The facts lead to the conclusion that portions of the gabbro were assimilated by the granophyre magma. The augite seems to have been made over completely, since the augite of the granophyre is not identical with that of the gabbro. The feldspar seems also to have been made over in great part, though it is considered possible that some of the relatively large crystals of plagioclase seen in the modified granophyre have been directly derived from the older rock, for they frequently show a clear border surrounding a turbid interior of rounded shape and corroded appearance. The iron ore and the apatite of the gabbro have been simply incorporated by the granophyre without undergoing solution. (*Fig.* 5.)

A very similar case was studied by W. J. Sollas in the district of Carlingford, Ireland. The top of the hill of Barnavave is gabbro, the base is granite and granophyre (granite-porphyry), the junction being well exposed. Dykes of granophyre cut the gabbro, and lumps of gabbro are enclosed in the granite, sometimes forming intrusion-breccia. Detached crystals from the gabbro lie in the granophyre, and in the absence of field evidence they would be mistaken for true constituents of the latter. From the large plates of augite (diallage) in the gabbro smaller augite grains have been formed, and occasionally some biotite or green hornblende. The basic plagioclase crystals of the gabbro have developed mantles of orthoclase, and the hybrid rock so produced might be called quartz-biotite diorite (Sollas).

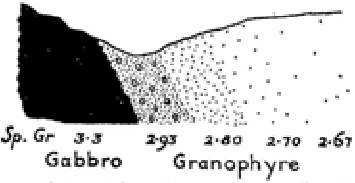

FIG. 5. Section illustrating the modification of the Carrock Fell granophyre by incorporation and assimilation of gabbro. (After A. Harker.)

A careful study of the normal and the contaminated granite, the xenoliths, and the hybrid rocks developed in this area has lately been made by S. R. Nockolds, who has supplied the chemical evidence required for the study of the reciprocal action described by Sollas.

Instances of reaction between granite and diabase-magma, or diabase (dolerite) and granitic magma, are frequently mentioned in geological literature, though details are often lacking. F. P. Mennell has described dolerite dykes which cut granite in Southern Rhodesia and incorporate masses of the granite itself or of quartz and feldspar. (See *Plate* II.) Mennell shows that there has been some solution of the enclosed material and some crystallization of micropegmatite in the dolerite. T. Krokström gives a full account of the Hällefors dolerite dyke in Sweden. Near its contacts with granite, this rock holds inclusions of granite and is itself greatly enriched with quartz and alkali-feldspar, which form micropegmatite intergrowths. The reaction is sometimes so complete that the dyke-rock is said to merge continuously with the wall rock, without any distinct boundary.

The converse occurrence, solution of diabase in granite, has been observed by A. G. Högbom (2) at Ragunda, Sweden. The granite sends a network of veins and dykes into the diabase, giving rise to an intrusion-breccia which consists of fragments of diabase cemented and more or less altered by the granite. (*Plate* III.) Högbom claims that intermediate rocks, characterized by chemical and structural inconstancy, have been produced by the assimilation of diabase by granite. J. J. Sederholm has given an account of reaction between diabase and intrusive granite near Wiborg, Finland. The granite encloses large masses of the diabase, several hundred metres in diameter, and in the neighbourhood of these the granite is so enriched with plagioclase that it acquires the character of quartz-monzonite. At Pecos, New Mexico, the injection of granitic magma into solid diabase has produced a remarkable banded rock which has been described and figured by P. Krieger.

The contamination of gabbro or norite magma by incorporation of sedimentary and metamorphic material has been studied closely by H. H. Read in Aberdeenshire. Magma which normally gives rise to norite (plagioclase-hypersthene rock) has, owing to assimilation of its country rock, produced rocks which contain cordierite, sillimanite, spinel, garnet, and hornblende, together with orthoclase and quartz. It is shown that in the process of assimilation there is an interchange of material between the magma and the enclosures, the magma

PLATE II.—The outcrop of a dolerite dyke which cuts granite near Bulawayo, Southern Rhodesia. The dolerite is crowded with fragments of quartz and feldspar, probably derived from a pegmatite reef in the granite. (By courtesy of Mr. A Frost, Bulawayo.)

[*To face p.* 76.

PLATE III.—Assimilation of dolerite (diabase) fragments by granite at Hammarforsen, Sweden. A. G Högbom. (Two-sevenths of natural size.)

[*To face p.* 77.

yielding lime, magnesia, and possibly soda to the enclosures and receiving alumina and potash in exchange. Read believes that after this interchange has taken place, the modified enclosures sink, leaving the magma lighter and more acid than before, so the final product of contamination may be a fairly normal-looking granitic or dioritic rock composed of quartz, biotite, hornblende and feldspar.

Even such highly resistant material as sandstone and quartzite has undergone assimilation by basic magma. W. A. Jones has studied the assimilation of quartzite enclosures by gabbro at Sudbury, Ontario. The irregular or lens-shaped enclosures are of all sizes up to a hundred feet long and forty feet wide. About every enclosure there is a marginal zone of hornblende and plagioclase, followed inside by a zone of plagioclase and quartz and sometimes there is a core of pure quartz. (See *Fig.* 2, p. 46.)

W. Q. Kennedy and H. H. Read have furnished an admirable study of cross-assimilation between an acid dioritic pegmatite-magma and greywacke in Dumfries, Scotland. The pegmatite loses alumina, soda and potash, and gains iron, magnesia and lime, resulting in basification of the magma and feldspathization of the greywacke.

An astonishing case of assimilation of sandstone was described at Julianehaab, Greenland, by N. V. Ussing. An augite-syenite consisting of feldspar, augite and olivine holds immense included blocks of sandstone and countless small fragments. One of the great blocks has a volume exceeding 200,000 cubic metres. Round each of these inclusions there is a zone of soda-granite, formed by solution of sandstone in the syenite-magma. The soda-granite has been completely fluid, since it sends small veins into the sandstone and the surrounding augite syenite.

The clearest evidence of solution at eruptive contacts will naturally be found in cases where the country rock differs very widely in composition from the invading magma, for then no amount of contact metamorphism can completely obscure the nature of the country rock, and any appreciable assimilation of it is bound to be evident in the changed composition of the eruptive near the junction. Such a case would be, for example, the contact of a granitic magma with a limestone. We shall see that such contacts have in point of fact furnished some of the most convincing evidence we possess of magmatic assimilation. As a first illustration we shall consider the classic work of A. Lacroix on the granites of the Pyrenees.

The granite massif of Quérigut is in contact with schists, quartzites and limestones of Paleozoic age, through which it is believed to have risen by dissolving them away, the schists yielding easily to dissolution while the limestones were more resistant. Consequently great masses of more or less altered limestone are completely enclosed in granite, and as a result of assimilation of limestone the granite loses its quartz and is locally converted into hornblende-mica-diorite, or where reaction has gone further, into norite and hornblende-peridotite. The argillaceous rocks all around the granite have been changed to mica-schist and mica-gneiss, and the limestone near the contact is silicified and full of wollastonite, epidote, plagioclase, garnet and hornblende. The proof of the modification of the granite by assimilation of limestone is complete. Wherever one finds a sheet of limestone enveloped by granite, it has a margin of hornblendic rocks which do not occur in any other situation. (*Fig.* 6.)

Lacroix holds that the granite rose into place in a slow, progressive manner by assimilation of the sediments which it replaced. Constantly, as the granite rose to higher and higher levels, it was preceded and accompanied by a "cortège d'émanations transformatrices," which reacted chemically with the sediments, so that the material which eventually passed into the magma was no longer normal sedimentary rock, but rock which had already undergone extreme metamorphism under the influence of the emanation. The elements which were added to the contact rocks were especially alkalies and silica. Thus in the case of the mica schists, these were first "feldspathized" by the emanation and converted into mica-gneiss, with a mineralogical composition not greatly different from that of the granite itself; consequently when the gneiss was incorporated in the granite, the only visible change in the latter was the development of some excess of biotite. It is otherwise with the limestone; no addition of alkalies and silica

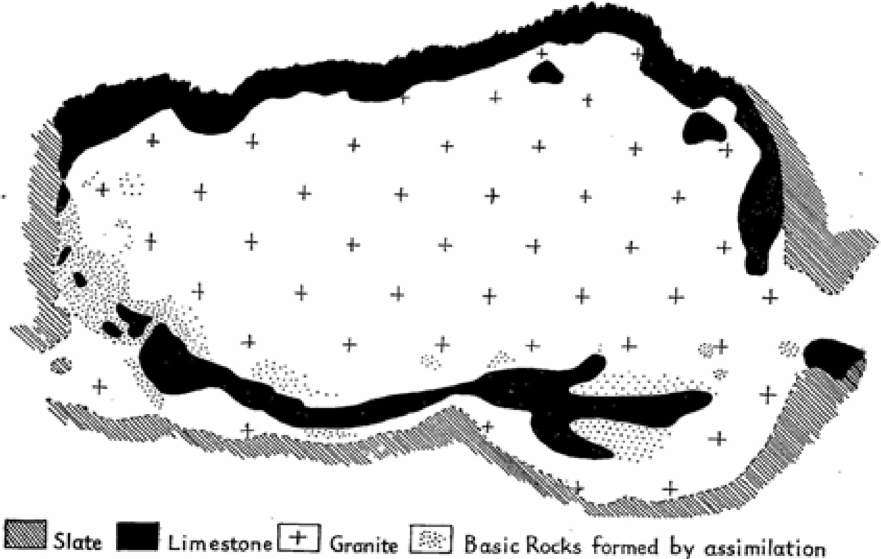

FIG. 6. The granite massif of Quérigut, Pyrenees. (After A. Lacroix.)

can transform this into anything resembling a normal granite, and the most characteristic minerals produced by its alteration are hornblende and lime-feldspar. It follows that assimilation of limestone by the granite magma must profoundly modify the composition of the latter, the tendency being towards the production of hornblende diorite or gabbro.

There is an abundance of evidence connecting the amount of assimilation performed with the degree of metamorphism in the country rock. A good illustration is supplied by the granite massif of Bordères. On its north side this massif has exerted little metamorphic action; the schists are not feldspathized and the limestone is not silicified. In this part the granite contains no hornblende and shows no basic modification. On the south side the contact metamorphism is intense; the granite is rich in hornblende and passes into a dioritic facies.

Another area of the greatest importance for the study of assimilation at granite contacts is the Haliburton-Bancroft district of Ontario, which has been described by F. D. Adams and A. E. Barlow. In an area of 4,200 square miles, granite and granite-gneiss occupy 2,596 square miles, or about 62 per cent. The eruptive complex involves three distinct elements, as follows:—

(a) red granite or gneiss,
(b) gray granite or gneiss,
(c) amphibolite inclusions.

The first of these is essentially a rock composed of quartz, orthoclase and oligoclase, with very little biotite. The second is richer in dark minerals, and contains both ortho- and clinopyroxenes as well as biotite and hornblende. The amphibolite inclusions form large masses of irregular shape, and also long narrow bands with distinct foliation; they range from masses occupying acres of ground to fragments of a few square inches. The constituent minerals are hornblende, which makes up about half of the rock, clino- and orthopyroxene, labradorite, and a little pyrite and calcite. Other basic inclusions are rich in olivine and spinel. These amphibolite masses vary much in number from place to place, although they are seldom entirely absent. On the southern margin of the Glamorgan batholith they are reckoned to constitute about half of the entire stock. There is no sharp line between granite with inclusions of amphibolite, and amphibolite cut up by veins of granite. All stages of the process can be observed whereby an immense solid mass of black amphibolite, absolutely free from any granitic admixture, is resolved into a swarm of amphibolite fragments distributed in a long stream through a body of mixed rock. While the detached masses often preserve their angular form, they sometimes seem to have been softened and rendered more or less plastic; and when this is the case, movements of the magma draw them out into long, narrow ribbon-like forms which, when pulled out very thin, pass into ill-defined gray streaks in the granite.

The amphibolites result in part from the intense alteration of limestone, every intermediate stage being represented between almost unaltered limestone, partly silicified limestone, and amphibolite with a little residual calcite. W. G. Foye shows that the conversion of limestone into amphibolite, as described by Adams and Barlow, involves a loss of 39 per cent of lime and 42 per cent. of carbon dioxide from the limestone, and a gain of 48 per cent of silica, 19 per cent of alumina, 6·8 per cent of ferrous oxide, 1·4 per cent of magnesia, 4·5 per cent of soda and 1·2 per cent of potash by the limestone from the granite.

The present writer has studied the reaction between granitic magma and a moderately magnesian limestone at Palabora, eastern Transvaal. (*Fig.* 7.) The intrusion of granite into the ancient schists of this region took place in several stages. At each stage there was an injection of pegmatite sheets and dikes which intersect those of earlier stages, forming an intrusive complex of a highly involved character. The most acid type of granite has the following composition:—quartz 27, microcline and microperthite 35, oligoclase 35, biotite, sphene and apatite 3 per cent. The products of reaction with limestone are:

(a) diopside-arfvedsonite syenites, containing microcline and only a little quartz;

(b) shonkinites (melanocratic syenites) consisting roughly of one-third microcline, two-thirds diopside with some apatite;

(c) pyroxenites, almost wholly composed of diopside or diopside and apatite; also apatite rocks.

All of these rocks are cut by younger granite. In the limestone, however, no granite veins were observed but only veins of syenite and shonkinite. The variation diagram (*Fig.* 8) shows the chemical changes involved in the reaction. The horizontal co-ordinate is the colour index, or the proportion of heavy minerals in the rock. The vertical co-ordinate gives the percentage of the various constituents, namely, quartz (Qz), albite (Ab), orthoclase (Or), apatite (Apat.) and $CaSiO_3$, $MgSiO_3$, $FeSiO_3$; also the sum of Or+Ab

Fig. 7. Map of the reaction complex between granite and limestone at Palabora, eastern Transvaal. (From *Trans. Geol. Soc. S. Africa*, vol. 51, *plate* 27.)

and the ratio Or/Ab. From the diagram one can deduce the operation of three distinct processes. The decrease of quartz and the concurrent increase of the three meta-

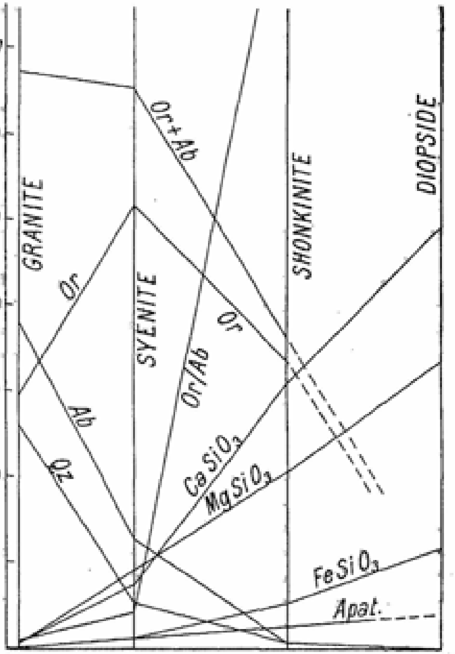

FIG. 8. Variation-diagram for the rocks of Palabora, Transvaal. The horizontal co-ordinate is the colour index. Explanation on pp. 82-83.

silicate molecules which constitute diopside can hardly be explained by any other process than assimilation of limestone, which went on until most of the free silica in the bordering magma was neutralized. The further large increase of lime and magnesia that appears in passing from syenite to shonkinite cannot be explained in this way, but seems to be due to the accumulation of diopside crystals in certain parts of the body of contaminated magma. The third process that is shown by the variation-diagram, namely, the progressive elimination of albite, does not concern us at present.

The syenites, shonkinites and pyroxenites of this region occupy several square miles around the central limestone mass, which has been charged with magnetite and apatite and has also developed diopside and olivine by thermal metamorphism. (*Fig.* 7.)

Although the most convincing examples of assimilation of limestone or dolomite have been found in connection with acid magmas, there is reason to believe that even very basic magmas can assimilate carbonate rocks. The curious rocks called "sagvandite," which have been studied by T. F. W. Barth, are found in Norway, Finland, and New Zealand in association with peridotites and dolomite-rich sediments. Typical sagvandite consists of large crystals of enstatite, or enstatite and olivine, in a groundmass of enstatite and magnesite (magnesium carbonate). Accessory minerals are talc, picotite, and anthophyllite. The rocks occur in small pipes or stocks from ten to five hundred metres in diameter, and their igneous character is shown by their texture and by the thermal metamorphism of the surrounding sediments. Barth considers sagvandite to have resulted from the assimilation of dolomite by an ultrabasic magma, and points out that the high concentration of CO_2 which must result from this reaction would enable the contaminated magma to make its way explosively into the surrounding rocks.

It has been maintained by R. A. Daly and others that a magma may lose so much silica, by assimilation of limestone, that it must form nepheline or leucite instead of the customary alkali-feldspars, and Daly holds that all or most of the feldspathoidal rocks have been generated in this way. There is a good deal of field evidence in support of Daly's view, which will be discussed in a later chapter. If an acid magma may be desilicated by reaction with limestone, it is also true that a magma of such low acidity that nepheline or olivine would normally be formed from it, may be so acidified by contact with siliceous country rocks that these minerals fail to develop in the neighbourhood of the contact. The nepheline-syenite of the island of Alnö, Sweden, is intrusive in granitic gneiss, and towards the junction the syenite loses its nepheline and acquires biotite, while the gneiss loses its quartz and develops ægirine. There is thus a contact-zone three hundred metres wide which is divisible, according to A. G. Högbom (3) into an outer belt of modified gneiss and an inner belt of modified syenite, with no sharp division between them. The nepheline syenite of Iivaara, Finland, also shows an acidified marginal zone, and so does that of Mariupol, Russia; while several cases have been described in Greenland. At Korok, for example, a syenitic rock containing nepheline and olivine loses both of these minerals and acquires quartz where it comes into contact with sandstone (N. V. Ussing). Similar evidence is available in the case of many olivine-bearing rocks; thus A. G. Högbom (4) describes sills of olivine-dolerite at Nordingra, Sweden, which lose their olivine and become quartz-bearing near their contacts with granite and sandstone.

Further discussion is hardly necessary. We are already in a position to meet the question, whether deep magma exerts any solvent action on its containing walls, with the assurance that a certain amount of assimilation of country rock has unquestionably taken place in many cases. What it is that determines whether reaction shall take place or not it is hard to say, but one may guess at the operation of several factors. The temperature of the magma at the time of intrusion is important, as well as the presence or absence of any notable degree of superheat. The composition of the wall-rock in relation to that of the magma is also significant. F. Zirkel enunciated the rule that in acid magmas, acid enclosures are not assimilated but basic ones are; in basic magmas, basic enclosures are not assimilated but acid ones are. This rule certainly expresses in a rough way the result of many field observations. It should be remembered, however, that in the event of the assimilation of an acid rock by an acid magma, or a basic rock by a basic magma, the composition of the magma would undergo so little change that evidence of the reaction might be hard to find in the field. H. H. Read reports that in Aberdeenshire the norite magma "appears to swallow up with avidity material of phyllitic nature, but to reject quartzose or calcareous rocks." The granites of the Onaping area, described by W. H. Collins, have found the siliceous and argillaceous schists much more digestible than the ferruginous slates.

The concentration of the fugitive constituents in the magma, and the occurrence of conditions which facilitate or retard their escape into the country rock, must have great influence in the matter. For instance, the reaction

$$CaCO_3 + SiO_2 \Longleftrightarrow CaSiO_3 + CO_2$$

is reversible, and if the concentration of CO_2 in the magma is high enough the silicification of a limestone in contact with that magma may be prevented. V. M. Goldschmidt ascribes the stability of calcite at the granite contacts of Rien, Norway, to the high partial pressure of CO_2. The granite of Bordères, in the Pyrenees, has reacted with limestone and schists on its south side, but not on the north. It seems likely that the facilities for the escape of fugitive constituents from the magma were greater on one side than the other, owing either to the structure of the country or to the inclination of the intrusive mass.

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> *A scientific argument which is wholly lacking in quantitative support cannot be satisfactory; but on the other hand a highly developed mathematical treatment which rests upon unsound premises is usually much worse than none.*—T. W. Richards.

CHAPTER VI

THE FREEZING OF SILICATE MELTS

When a liquid freezes, it may give either a mass of crystals or a glass. In the first case there is an abrupt change of all the physical properties of the system, such as viscosity, density, refractive index, and specific heat, at the freezing point. In the second case there is only a gradual increase in the viscosity of the liquid, until it becomes so great that we think of the system as a solid, accompanied by small progressive changes in the other properties; there is no definite freezing point at all. The factors which determine whether any liquid shall crystallize or become a glass are principally the fluidity of the liquid and the rate of cooling; other things being equal, slow cooling favours crystallization and high viscosity retards it. Since the gases and simple salts dissolved in a magma increase its fluidity, a high concentration of fugitive constituents will favour crystallization; and there is reason to believe that the glassy character of certain lavas is due more to the loss of their gases than to rapid cooling.

Freezing with production of glass introduces nothing new, for glass is just a more viscous liquid; but crystallization involves the appearance of new solid phases, and the course of crystallization and the nature of the products are subject to laws which we must study.

A pure substance such as mercury or water has a perfectly definite freezing point (at a given pressure); above the freezing temperature only the liquid form can exist, below it only the solid form is stable. Under a fixed pressure, there is only one temperature at which the solid and the liquid can exist indefinitely in contact with each other, and that is the freezing point. If we ever had to deal with a magma consisting of molecules of one kind only, say a pure anorthite magma, we should expect it to crystallize at 1550° C. precisely (under atmospheric pressure), while a pure diopside magma should crystallize at 1390° C. Such a magma is probably never realized in nature. Many rocks are known which consist almost entirely of one mineral, but small amounts of other minerals are seldom absent, and in addition we have seen that certain fugitive constituents are probably present in every magma. Most rocks contain from four to ten or more distinct minerals, and when some of these are complex solid solutions, like hornblende and augite, the number of simple silicate molecules in the magma must be large.

Now a liquid mixture with two or more components behaves in general quite differently, in the act of freezing, from a liquid of simple composition. The difference is shown in the following ways:—

(1) The temperature at which freezing begins is not the same as the freezing point of any one of the components, and is different even in different mixtures of the same components.

(2) Freezing does not take place entirely at one temperature, but continues through a certain range of temperature. That is, instead of a freezing point there is a freezing interval.

(3) The crystals that form first do not have the same composition as the liquid out of which they crystallized. In different mixtures of the same components, the crystals that form first may be different.

The first point is just a re-statement of the Raoult-Blagden Law (page 48) that every substance when dissolved in a liquid capable of crystallizing, lowers its freezing point. Sea water has a lower freezing point than fresh water on account of the small quantity of salt dissolved in it, and by increasing the concentration of salt the freezing point can be brought down to —23° C. In a mixture of molten silicates there is no distinction of solvent and dissolved substance, but each component may be considered to be dissolved in the others.

The freezing of the system anorthite-diopside has been studied by N. L. Bowen (1). The melting or freezing point of pure anorthite is 1550° C., that of diopside is 1390° C., but a mixture containing anorthite 80 parts, diopside 20 parts, only begins to freeze at 1485° C. and is not completely solid till 1270° C. is reached. A mixture of diopside 80 parts, anorthite 20 parts, begins to freeze at 1350° C., and the last trace of liquid only disappears at 1270° C. Each of these mixtures has therefore a *freezing interval* instead of a single freezing point, which illustrates the second of the above statements. But the mixture which contains 58 parts of diopside to 42 parts of anorthite does not even begin to freeze until 1270° C. is reached, and then crystallizes completely at that temperature. This particular mixture of diopside and anorthite has the lowest possible freezing point, and is called the *eutectic mixture* of these two components. Observe that a eutectic mixture resembles a pure substance in having a definite freezing point.

The third point is easily verified by referring again to salt and water. It is well-known that when sea-water freezes, the ice that forms is entirely free from salt; on the other hand, that if concentrated brine is cooled down, it deposits pure salt at first. The solution that contains exactly 23·5 per cent of salt (the eutectic mixture) deposits ice and salt simultaneously. A silicate melt behaves in a similar way, as the following table shows:—

Mixtures of diopside and anorthite.

I.	II.	III.
Mixtures with more than 58% of diopside.	*Mixture of 58 parts diopside, 42 parts anorthite.*	*Mixtures with more than 42% of anorthite.*
Diopside crystallizes first and continues to form until the eutectic temperature (1270° C.) is reached; then diopside and anorthite crystallize together in eutectic proportions.	This mixture remains fluid until the eutectic temperature (1270° C.) is reached; then it gives crystals of both minerals in the above proportion of 58 to 42.	Anorthite crystallizes first and continues to form until the eutectic temperature (1270° C.) is reached; then anorthite and diopside crystallize together in eutectic proportions.

We may put these observations in the form of a rule of general application, as follows: Any mixture which contains one component in excess of the eutectic proportion will deposit that excess as it cools, until the remaining liquid has the eutectic composition, when the mixture will freeze as a whole.

The precise composition of the eutectic mixture of any two substances can only be found by experiment, but it is roughly true that the eutectic mixture holds more of the component of lower melting point, and the greater the difference between the melting points of the two components, the more of the more fusible component appears in the eutectic.

It is much easier to demonstrate these relations graphically than verbally. In *Fig.* 9 the composition of all possible mixtures of diopside and anorthite is plotted on the base line, while temperature is recorded vertically. When the initial freezing points of various mixtures are inserted in the diagram, they are found to lie on two curves, which start from the melting points of the pure components and meet at the eutectic point, corresponding to the composition diopside 58, anorthite 42, and to the temperature 1270° C. This diagram therefore contains complete information about the crystallization of all possible mixtures of diopside and anorthite.

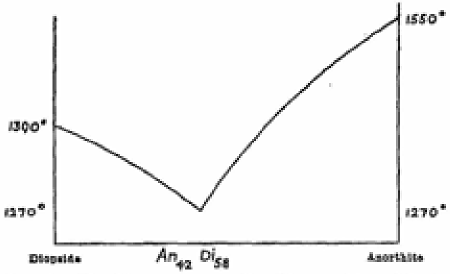

FIG. 9. The binary system diopside-anorthite. (N. L. Bowen.)

Every extra component that is added to such a mixture (provided that it does not enter into chemical reaction or isomorphous substitution with the orginal components) causes a further lowering of the freezing point, the new minimum corresponding to a ternary, quaternary, or higher eutectic mixture. Thus in the system silica-anorthite-wollastonite, which has been studied by G. A. Rankin (*Fig.* 10) there are three binary eutectics crystallizing at 1436°, 1359°, and 1299° C. respectively, and a ternary eutectic, composed of 40 parts of anorthite, 32 of wollastonite, and 28 of tridymite, which freezes at 1165° C. In the freezing of a ternary mixture, we may say loosely that the component which is in greatest excess, as compared with the composition of the ternary eutectic, will crystallize first; then the component which is next in excess, along with the former one; then the ternary eutectic of all three components.

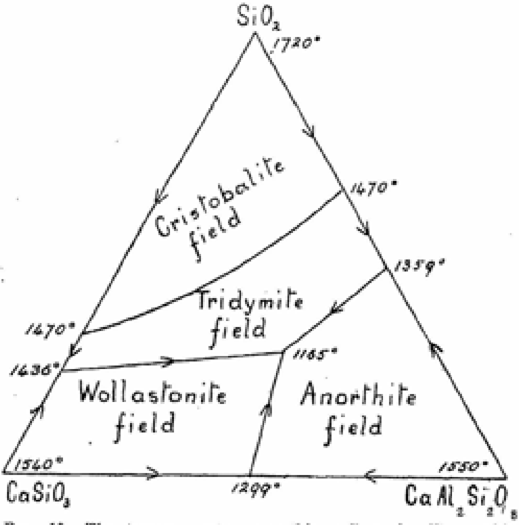

FIG. 10. The ternary system anorthite-wollastonite-silica. (After G. A. Rankin.)

The graphical representation of the freezing of such a system becomes very difficult as the number of components increases. For a ternary system one uses a triangular diagram in which the percentage of each component is measured from 100, at an apex of the triangle, to zero at the opposite side. Thus in *Fig.* 10 the point marked 1165° C. corresponds to the composition anorthite 40, wollastonite 32, tridymite 28, which is that of the ternary eutectic of these components. A point on one of the sides of the triangle obviously represents a mixture containing only two of the components. The lines that join the ternary eutectic point to the three binary points cut the triangle up into three fields, one of which encloses all mixtures from which anorthite crystallizes first, another those from which wollastonite crystallizes first, and a third those from which silica crystallizes first. It is to be noted that the latter field is itself divided by a line which corresponds to the temperature of 1470° C. Mixtures which are very rich in silica begin their crystallization above 1470° C., so the silica crystallizes as cristobalite; when the temperature of crystallization is below 1470° C., tridymite appears instead. The directions of falling temperature are indicated by arrows, the temperature co-ordinate being supposed to stand at right angles to the plane of the paper.

In short, from a eutectic system of any number of components, the course of crystallization will be ideally of the following type:—

$$A;\ A+B;\ A+B+C;\ A+B+C+D;\ \text{etc.},$$

always ending with a multiple eutectic of all the components. The eutectic relation only holds between substances which are not isomorphous—that is, which are not related to each other by ionic substitution. If two substances in mutual solution belong to an ionic substitution series, as albite and anorthite, or forsterite and fayalite, then the crystals formed at any temperature always hold some of *each* component—in fact the composition of the crystals may be only a little different from that of the liquid phase—and, furthermore, the composition of the first-formed crystals varies with that of the liquid.

The behaviour of such a system may be learned from *Fig*. 11, which is N. L. Bowen's (2) well-known diagram for the plagioclase feldspars. There are two curves extending from the melting point of anorthite (1550° C.) to that of albite (1100° C.). The upper curve (called the liquidus) contains the initial freezing points of all possible liquid mixtures of albite and anorthite; and the lower curve (the solidus) expresses the composition of the feldspar crystals in equilibrium with these liquids. From this diagram one learns, for example, that a melt of the composition Ab_1An_2 begins to crystallize at 1490° C., and that the crystals formed at that temperature have the composition $Ab_{12}An_{88}$. Thus the remaining liquid is enriched with albite, and as cooling proceeds the composition of the liquid changes steadily, following the liquidus towards the albite end. The crystals meanwhile continue to grow. Under ideal conditions of cooling the crystals should react continuously with the liquid so as to remain always in equilibrium with it; that is, the composition of the solid phase should follow the solidus while the liquid phase follows the liquidus, until the crystals have the composition Ab_1An_2, by which time all the liquid will have been used up. But if it should happen that crystallization is too rapid for the crystals to adjust their composition in this way, then growth will take place by the addition of new layers of feldspar which are successively richer and richer in albite, ending up with almost pure albite.

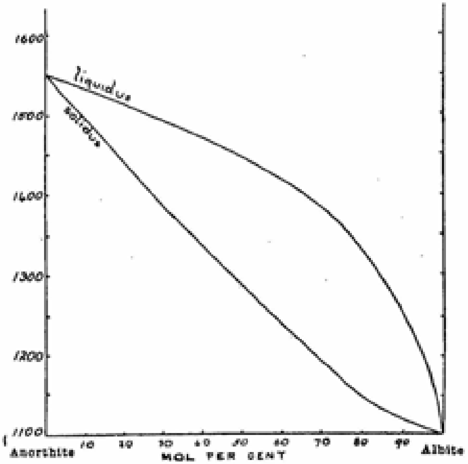

FIG. 11. The melting and freezing curves of the plagioclase feldspars. (N. L. Bowen.)

The crystals will then exhibit the zonal structure with which every petrographer is familiar.

In this type of crystallization there is no eutectic, and the lowest temperature that can be reached by the liquid is the melting point of the more fusible component—in the above case, albite. This is the commonest type of crystallization among such of the rock-forming minerals as have been studied experimentally. When isomorphism is not so complete as in the above instance, two distinct solid phases may be formed in the system, one richer in component A, the other richer in component B; and between these two there may be a low-point somewhat similar to a eutectic point. This possibility is believed to be realized in the case of orthoclase and albite.

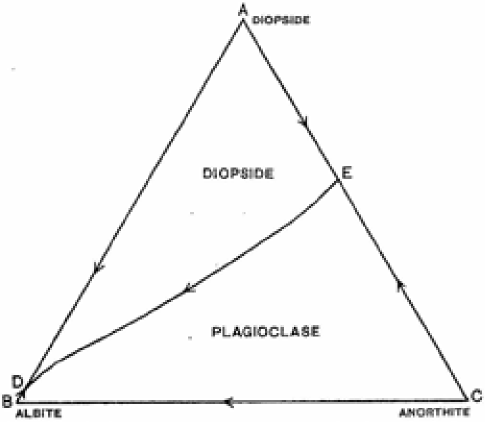

FIG. 12. Field diagram of the system: diopside, plagioclase. (N. L. Bowen.) Arrows indicate directions of decreasing temperature.

The important point to observe, in connection with ionic substitution systems, is that where there is no binary eutectic point there cannot be a ternary, a quaternary, or a multiple eutectic; and since the feldspars, the micas, the pyroxenes, and most of the other important rock-forming minerals are ionic substitution complexes, it follows that the crystallization of a granitic magma cannot end with the formation of a "granite eutectic," or a gabbro magma with a "gabbro eutectic," as was at one time conjectured. In short, eruptive magmas are not simple eutectic systems.

We may inquire next what happens when an additional component is added to an ionic substitution system to which it is not related. *Fig.* 12 represents the system diopside-albite-anorthite according to N. L. Bowen (8). There are only two fields, one from which diopside crystallizes first and one from which a plagioclase of variable composition crystallizes first. Between these fields there is a boundary which Bowen calls a reaction-curve, Vogt a cotectic line.

From any melt whose composition throws it into the plagioclase field, plagioclase will crystallize first until the composition of the liquid reaches a point on the reaction curve (cotectic line); thereafter plagioclase and diopside will crystallize together, but all the time the plagioclase will be reacting with the liquid and changing its composition in the manner already explained in connection with the binary system. Crystallization will end at a point which is *not fixed,* as a eutectic point is fixed, but varies according to the initial composition of the melt and the extent to which reaction takes place between feldspar and liquid.

A still more complex case is presented by a system containing *two pairs of isomorphous compounds,* such as albite-anorthite and diopside-hedenbergite (or simply plagioclase and pyroxene). A lucid account of the crystallization of this system has been given by W. Wahl. In *Fig.* 13 we may suppose A to represent albite, B anorthite, C diopside, D hedenbergite, these four points forming the corners of a tetrahedron. Then the point e_1 represents the eutectic diopside-albite, e_2 the eutectic diopside-anorthite, while e_3 and e_4 are the eutectics hedenbergite-anorthite and hedenbergite-albite. We have therefore in the system two binary isomorphous series, A-B (plagioclase) and C-D (pyroxene); four binary eutectic systems, C-A, C-B, D-B, D-A, with the binary eutectic points e_1, e_2, e_3, e_4; four ternary systems with the cotectic curves e_1e_2, e_2e_3, e_3e_4, e_4e_1; one quaternary system with the cotectic surface $e_1e_2e_3e_4$.

A quaternary mixture represented by any point in the upper part of the figure will solidify in such a way that mix-crystals AB (plagioclase) separate first. The liquid will thereby be enriched with the components C and D until a point on the cotectic surface is reached. Now AB (plagioclase) and CD (pyroxene) will begin to separate simultaneously, and as cooling proceeds the crystals will react with the melt and adjust their composition continuously until the last of the liquid is used up. If the rate of cooling is too rapid to permit of complete reaction, then both kinds of crystals will be zonally built.

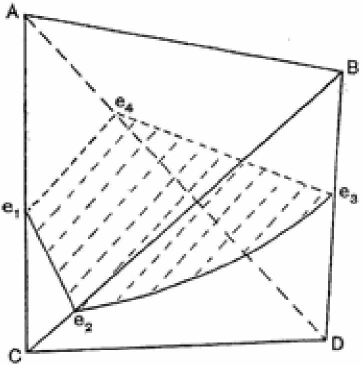

FIG. 13. Schematic equilibrium-diagram of a quaternary system consisting of two pairs of isomorphous substances with a cotectic surface between them. (After W. Wahl.) Explanation in text.

Although there is no quaternary eutectic there will yet be a definite order of crystallization just as in a binary system, plagioclase crystallizing first, followed by plagioclase and pyroxene. If the initial mixture had a composition represented by a point in the lower part of the figure the order of crystallization would be first pyroxene, then pyroxene + plagioclase. Furthermore, the behaviour of the solution on the cotectic surface depends on the shape of that surface, which may have a maximum or a minimum point or may be saddle-shaped. The problem is further complicated by the possibility that the extent of ionic substitution in one or both series may be limited, as in the case of orthoclase and albite; this might lead to the simultaneous crystallization of three different kinds of solid solution (as AB, BA, and CD) at the cotectic surface.

The difference between crystallization at a eutectic point and crystallization at a cotectic surface may be summarized as follows :—

Eutectic Point.	*Cotectic Surface.*
(1) Crystallization takes place at constant temperature.	(1) Crystallization takes place with falling temperature.
(2) Crystallization gives a mixture of constant composition.	(2) Crystallization gives a mixture of variable composition.
(3) The phases separating have a definite composition.	(3) The phases separating have a considerable range of composition.

It will now be realized that the study of the course of crystallization in a melt or magma becomes enormously complicated when cotectic surfaces take the place of the comparatively simple eutectic points.

T. F. W. Barth has tried to establish the position of the cotectic surface for basalt in terms of the components albite, anorthite, diopside, hypersthene. Reasoning that basalts which show evidence of simultaneous crystallization of plagioclase and pyroxene throughout their cooling-history should lie on the cotectic surface, he has plotted the positions of nine such basalts in a tetrahedral diagram like that of *Fig.* 11 (hypersthene taking the place of hedenbergite) and finds that they lie almost exactly in one plane. Since a plane is completely defined by three co-ordinates, Barth finds it possible to define the position of the cotectic surface by means of the following equation (in which ab stands for albite, di for diopside, hy for hypersthene);

$$ab' + 2\ di' + 2{\cdot}3\ hy' = 123$$

To apply this formula one must first calculate the norm of the rock, then recalculate ab, an, di, hy, to 100, which gives ab′, an′, hy′. If the sum of the three terms is greater than 123 the basalt lies in the pyroxene field, if less, in the plagioclase field.

Another kind of complication arises when two of the components of a system are capable of reacting together as the temperature falls. Some of the rock-forming silicates show the phenomenon of *incongruent melting;* that is, they decompose as they melt. Thus enstatite decomposes at its melting point into forsterite and silica (Bowen and Andersen); and orthoclase decomposes into leucite and silica (Morey and Bowen). Thus from a melt rich in the constituents of enstatite, olivine will generally crystallize first, and later, when the temperature has fallen sufficiently, it will react with the remaining liquid to form enstatite. Or from a melt rich in the constituents of orthoclase, leucite may crystallize first, afterwards reacting with the liquid to form orthoclase. But the reaction may be prevented, either by rapid cooling or by the separation of the early crystals from the residual liquid. We shall see in later chapters that all these possibilities are realized in rock magma.

SILICATE MELTS CONTAINING WATER

The experimental study of silicate systems containing water presents great difficulty on account of the high pressure required to retain water in solution. The system $H_2O - K_2SiO_3 - SiO_2$ was studied by G. W. Morey (1, 2) in 1917. Although this system bears little resemblance to rock magma, some results of great importance to petrology and volcanism came out of the investigation. It was shown (1) that water vapour under atmospheric pressure is appreciably soluble in liquid silicates at their melting points, and that the amount of water dissolved is sufficient to cause an appreciable lowering of the freezing point; (2) that in the binary system $H_2O - SiO_2$, at the critical point of pure water, the amount of silica in both liquid and vapour phases is approximately zero; (3) that a silicate magma containing only a small amount of water, cooled in such a manner that the water cannot escape, will develop very high pressure as crystallization advances.

R. W. Goranson (1) investigated the solubility of water in granite magma, with results which we have already summarized (p. 38). He also studied (2) the binary systems albite—water and orthoclase—water, and found the maximum amount of water soluble in melted albite to be in the neighbourhood of 9 per cent at 900°. He discussed the pressure generated by the cooling and crystallization of albite-water melts, and showed that it might reach 4000 and perhaps even 5000 bars. He also demonstrated (3) that at 700°, under a vapour pressure of 980 bars, the granite of Stone Mtn., Georgia, became completely liquid (except for a little hematite), the melt holding 6·5 per cent of water in solution. He calculated that the liquidus temperature of the dry granite must be about 1050° ± 50°, but later corrected this figure to "about 950°."

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Confusion is merely the unknown intermixture of laws, and becomes order and beauty when we rise to their comprehension.—JOHN TYNDALL.

CHAPTER VII

THE FREEZING OF ROCK MAGMA

WHEN we turn from artificial silicate melts to natural rock-magma, the difficulty of investigating the course of crystallization is enormously increased. In the first place, we can only learn the chemical composition of the magma by inference from that of the rock or rocks to which it gave rise in cooling. But a statement of the chemical composition of the rock ignores the fugitive constituents which were lost during crystallization, and it takes no account of other factors which may cause the rock to differ in composition from the parent magma. In lavas of Vesuvius, both the sinking of olivine and the flotation of leucite crystals have been demonstrated; thus a single body of rock may be enriched in one part with leucite crystals and in another part with crystals of olivine or augite. Specimens taken from different parts of this body of rock would differ considerably in chemical composition, and none of them would reproduce accurately the composition of the parent magma.

Unpredictable changes in the composition of the magma may result from its reaction with wall rocks. We saw in chapter V that a certain amount of assimilation of country rock has unquestionably taken place in some magma chambers, and that it must be expected wherever liquid magma has stood for a time in contact with solid rock of different composition. Reaction between the liquid and the solid will bring about an increase of some components and a decrease of others, in the marginal portion of the magma, thus the marginal rocks may differ widely in chemical and mineral composition from the uncontaminated magma and rock. Large gains of lime and magnesia by acid magma have been demonstrated in the vicinity of limestone contacts; also gains of alumina from argillaceous sediments and of silica from sandstones. The transference of large quantities of silica from magma to limestone is familiar to all students of contact metamorphism. Loss of alumina and alkalis from the magma is less obvious in the field, but it has been proved in many cases by chemical analysis of the contact rocks.

Since rock exposures are often inadequate, and nearly always so in the vertical direction, it must often happen that the rocks which we see in the field, and from which we deduce a magma of similar chemical composition, are in reality abnormal products, due either to local accumulation of certain crystals or else to local contamination of a magma by reaction with wall rock during its emplacement. To convince oneself of this, it is only necessary to think of the pyroxenites, the peridotites, the lamprophyres, and certain feldspathoidal rocks such as ijolite and italite. We must accordingly agree that it is seldom safe to draw conclusions, from the chemical composition of a given rock, as to the initial composition of the magma which gave rise to it.

Our problem is complicated, not only by irregular changes of concentration, but also by unknown fluctuations of temperature and pressure in the magma during crystallization. The cooling of a body of deep magma, enclosed in a chamber which has no outlet to the surface of the earth, is likely to proceed with great regularity. This is not true of magma which is brought suddenly into contact with cold rocks, or rushed quickly to the surface through a volcanic pipe or fissure. Under these conditions a period of slow cooling is followed by one of rapid cooling. This will generally give rise to a rock with porphyritic texture, or else crystallization may be so retarded by increasing viscosity that the magma congeals largely as a glass. But in the case of magma that is rushed quickly to the surface, the expected fall of temperature may be preceded by a sharp, local rise of temperature, for which several causes suggest themselves. Most important of these is the combustion of hydrogen and other volcanic gases in contact with air. T. A. Jaggar has shown that the temperature of the surface layer of the lava in the fire-pit at Kilauea may be raised two or three hundred degrees in this way. Another possibility of reheating may be seen in a sudden onset of crystallization, perhaps initiated by the loss of dissolved gases. This possibility is illuminated by an experiment which was described by W. Guertler, a student in G. Tammann's laboratory. A sodium silicate glass (an undercooled liquid) was heated slowly to about 550° C., whereupon crystallization set in with such vigour that the temperature of the mass was raised to 750° within five seconds. Still another possible source of heat deserves consideration. The viscosity of basaltic lava is of the order of a million times that of water (R. L. Nichols), so the ascent of lava, driven by tremendous gas pressure through tortuous, rubbled-filled channels, can hardly fail to give rise to some mechanically developed heat. L. H. Adams has expressed the opinion that "ordinary rock material, if extruded from a pressure corresponding to a depth of 20 kilometres, would heat itself up nearly 200° C. above its original temperature."

Since some of the processes that we have discussed would lead to local cooling of the magma, and others to local heating, there may be many departures from the expected course of crystallization, even in rocks of identical chemical composition. Local reheating may bring about corrosion and destruction of early-formed crystals, a phenomenon often observed in lavas. Small fluctuations of temperature account for the interrupted and reversed zoning that is so common in plagioclase crystals. F. Homma has supposed that these crystals were alternately floated up by attached gas bubbles and allowed to sink when the bubbles became detached; thus the crystals may have wandered repeatedly through layers of lava of slightly different temperature and concentration, where they acquired successive mantles of feldspar of appreciably different composition. Oscillatory crystallization of two different minerals, such as that illustrated in *Fig.* 14, may be explained on similar lines.

Too rapid cooling may lead to the preservation of unstable phases. For example, a magma rich in magnesia and silica may yield olivine (forsterite) during an early stage of cooling. If it is then chilled quickly, the normal reaction of olivine with the siliceous magma to form pyroxene may be prevented. The resulting rock will then hold crystals of olivine in a highly siliceous glass base, an unstable association which is only perpetuated by the extreme slowness of reaction between solids below their melting points.

Change of pressure probably has little effect upon the crystallization of rock magma, as long as a gas phase is absent. Most silicates are believed to contract in freezing, so the freezing point should be raised by compression; but the elevation of the freezing point of a silicate under compression was estimated by J. H. L. Vogt (1) to be only some 5° C. per 1000 atmospheres; a quantity scarcely large enough to have any important petrological consequences. Some petrographers maintain that under extremely high pressure, familiar minerals such as augite and plagioclase must be replaced by species of greater density such as garnet and jadeite; thus a basaltic magma crystallizing at great depths in the earth-crust might give rise to eclogite (garnet-omphacite rock) instead of gabbro or norite. There is no compelling field evidence in favour of this opinion. J. Johnston contrasted the influence of pressure and temperature upon physical equilibria in the following terms :—

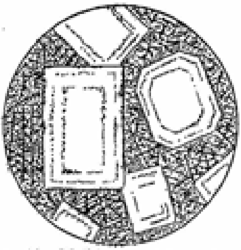

FIG. 14. Melilite crystals enclosing two successive generations of magnetite grains and embedded in a groundmass of magnetite, melilite and fayalite. (After J. H. L. Vogt.)

(1) In systems without a vapour phase, a change of pressure of 1000 atmospheres will in general effect no more than a change of temperature of 10° C.

(2) The effect of pressure upon viscosity and similar factors is in general insignificant as compared with the effect of temperature.

(3) Solubility is affected only slightly by change of pressure, so slightly that the effect is negligible as compared with that of change of temperature.

But if a body of magma rises to high levels in the crust, and especially if communication is established with the atmosphere, then the pressure may be lowered so far as to permit vesiculation (boiling) and rapid loss of gases, especially water. The effects that may follow the loss of gases from the magma have already been indicated.

THE ORDER OF CRYSTALLIZATION IN ERUPTIVE ROCKS

If now, aware of all the complications and difficulties that we have discussed in the last few pages, we try to establish the order of crystallization of a particular rock, we find that there are only three rules for our guidance, namely:—

(1) When one crystal encloses another, the latter is of earlier formation;

(2) The earlier crystals will generally be well shaped, the later ones less so owing to their mutual interference;

(3) When a rock contains both large and small crystals, it is to be presumed that the large crystals started to grow first.

But straightforward as these rules seem, they are not wholly reliable, as the following observations show. In some of their experiments on synthetic lavas, Fouqué and Lévy observed that leucite crystals were formed which held enclosures of glass, and that this glass subsequently crystallized and formed minute augite crystals within the leucite. J. Morozewicz (1) has described synthetic cordierite crystals with glass enclosures which, on continued heating, deposited magnetite crystals within the host. In either of these cases, the application of rule 1 would lead to the wrong conclusion as to the order of crystallization. F. Walker has found olivine crystals enclosing small crystals of augite, and has shown that the augite is the younger of the two minerals, the magma having made its way along cracks in the olivine crystals and crystallized there as augite and feldspar. G. P. Merrill, too, observed augite growing along cracks in olivine crystals. Quite recently, F. Homma has identified plagioclase, augite and magnetite that crystallized from occluded drops of magma within large crystals of plagioclase, in a Japanese andesite.

N. L. Bowen points out that it depends entirely on the direction in which a rock-section happens to be cut, whether an enclosure seems to lie at the centre or near the margin of the host. An enclosed crystal may appear in the centre of the section, yet it may have lain only just within the margin of the host. In such a case it is possible that the core of the host crystallized before the enclosed crystal, and the mantle afterwards; thus the apparent order of crystallization, as judged by enclosures, is really the order of completion of crystallization, not that of beginning of crystallization as it is usually held to be.

The second rule is also doubtfully reliable, for some minerals, such as garnet and staurolite, are able to form well-shaped crystals even when they grow in a solid matrix; this is shown by the appearance of perfect crystals of these minerals in metamorphic rocks which have been recrystallized without melting. On the other hand, a mineral which lacks crystal outlines was not necessarily formed at a late stage, for crystals of early formation, such as quartz and olivine crystals in certain lavas, are often partly resorbed at a later stage in the cooling process and so become rounded.

As regards the third rule, when one constituent such as feldspar occurs in enormous crystals which are embedded in a fine-grained crystalline matrix, one naturally concludes that the large crystals started to grow long before the little ones. Yet feldspar crystals of the largest size in granites and syenites are often full of minute enclosures of the other minerals right to the centre. The big leucite crystals in the lavas of the Eifel district of Germany often enclose grains of every other mineral that is associated with them. In these cases the third rule and the first lead to different conclusions.

It is such difficulties and contradictions as these that give force to J. H. L. Vogt's remark that "many of the petrographical contentions about the order of crystallisation in igneous rocks are partly questionable and partly wrong." The contentions to which Vogt takes exception are those that are embodied in the Rosenbusch Rule, according to which the normal order of crystallization in eruptive rocks is as follows :—

(1) Accessory minerals, such as zircon, xenotime, apatite, magnetite, chromite, sphene : or more generally those constituents which are present in smallest quantity;
(2) Ferromagnesian minerals, orthosilicates such as olivine preceding metasilicates, and ortho-pyroxene preceding clino-pyroxene;
(3) Feldspars and other alkaline silicates, the lime feldspars first :
(4) Quartz, and any glassy residue.

It must be said at once that the great petrographer whose name is associated with the rule was fully aware that his rule is one with many exceptions; the question before us is whether the exceptions are important enough to invalidate the rule. It will be seen that the very first section of the rule appears to contradict the laws of solution, for the components that are present in least amount ought to be among the last to crystallize. But when, in a granite, one sees minute crystals of zircon, xenotime or allanite enclosed in biotite, and little biotite scales enclosed in orthoclase, what conclusion can one draw except that the zircon crystallized first, then the biotite, and lastly the feldspar? It is true that if the minerals had all crystallized *at the same time,* then the large and abundant feldspar crystals would be almost certain to enclose some of the smaller biotites and zircons; but then it is almost as unlikely that the minerals should all have crystallized at the same time as it is that zircon should have crystallized first.

The strange contradiction now appears, that whenever the proportion of an accessory mineral such as zircon rises to a higher value than usual, then that mineral appears to have crystallized at a later stage than usual. So in a nepheline rock occurring at Mariupol, Russia, which contains about 2 per cent of zircon, that mineral has enclosed albite and ægirine. According to J. Morozewicz (2) the order of beginning of crystallization in this rock was albite, ægirine, zircon, nepheline; and the order of completion of crystallization was zircon, nepheline, ægirine, albite. Late crystallization of zircon has also been noticed in the zircon-syenites of Norway.

and the writer has observed it in a dyke rock in South West Africa; in each case the rock is unusually rich in zircon.

Sphene, like zircon, is often present in more than the average quantity in some syenites and foyaites. A dyke-foyaite in Assynt, Scotland, which was described by J. Horne and J. J. H. Teall, contains much sphene which has crystallized in skeleton plates, enclosing orthoclase, nepheline, and pyroxene. (*Fig.* 15.) Poikilitic sphene is abundantly present in the "melteigite" of the Fen district, Norway (W. C. Brögger), and in the foyaite of Beemerville, New Jersey.

FIG. 15. Skeleton crystal of sphene (dotted) enclosing pyroxene (P), orthoclase and nepheline; Assynt, Scotland. (After J. J. H. Teall.) Magnification 28.

Apatite also shows such anomalies, for plagioclase, hornblende, augite and olivine have all been observed to be enclosed in apatite at one time or another. W. S. Bayley figures skeleton crystals of apatite from a diabase in Minnesota, and P. A. Wagner has figured similar skeleton crystals from "blue ground" at Kimberley, in which apatite is one of the last minerals to crystallize. (*Fig.* 16.)

The lime-iron garnet, melanite or andradite, is an accessory constituent of many syenitic rocks, and it frequently crystallizes at a very late stage. C. W. Drysdale mentions an augite syenite at Franklin, British Columbia, carrying melanite which has crystallized after biotite, hornblende and diopside. Still later crystallization of this mineral has occurred in Finland, where it follows nepheline (V. Hackmann); and the writer (1, 2) has described instances of very late formation of melanite in Scotland and in the Transvaal. (*Fig.* 17.) Similar observations were made by W. C. Brögger (1) in the Fen district, Norway. The relatively rare titanosilicates and zirconosilicates which appear in the role of accessory constituents in many soda-rich rocks are among the latest minerals to crystallize. In the foyaites and tinguaites of Pilansberg, Transvaal, the writer (3) has found eudialyte, astrophyllite, molengraaffite (lamprophyllite) and pectolite in the form of poikilitic plates which enclose every other mineral in these rocks.

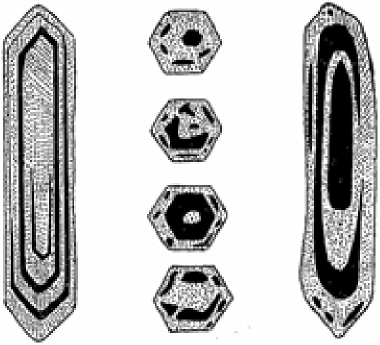

FIG. 16. Skeleton crystals of apatite from kimberlite; Kimberley. (After P. A. Wagner.)

Even the oxides crystallize late at times. W. C. Brögger (2) recorded the occurrence of nepheline in magnetite in Norway, and A. Streng showed that, in certain German diabases, magnetite had crystallized after plagioclase and olivine. Ilmenite sometimes encloses orthoclase; and even chromite, which, as a rule, precedes olivine, has been observed to be moulded on olivine crystals in the island of Rum (A. Harker).

When we come to the second section of the Rosenbusch rule, the exceptions are no less numerous. It has always been recognized that plagioclase begins to crystallize before augite in many dolerites and basalts. A. Streng's observations (2) on the basalts of Giessen, Germany, illustrate this point very clearly. These rocks contain plagioclase, augite and olivine, and the upper levels of the flows are partly glassy. In the more light coloured facies, there are crystals of plagioclase and olivine in a glassy base which retains the augite; while in the darker facies the crystals are of augite and olivine, and it is the feldspar which is retained in the glass.

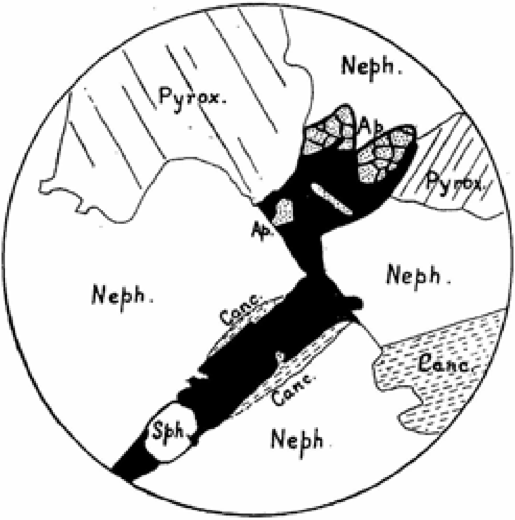

FIG. 17. Melanite (black) with pyroxene, nepheline, apatite, sphene and cancrinite. The melanite is interstitial to all constituents except cancrinite. Magnification 20. Sekukuniland, Transvaal.

A. Harker showed that among the anorthite-olivine rocks in the island of Rum, Scotland, those varieties which hold much olivine show that mineral crystallizing first; the anorthite-rich varieties show anorthite crystallizing first; and some rocks of intermediate composition show the two minerals to have crystallized simultaneously. (*Fig.* 18.)

It has lately been shown by N. L. Bowen and J. F. Schairer that whether olivine or pyroxene crystallizes first (from an anhydrous melt) depends upon the composition of the olivine. If the latter is a highly magnesian variety then it crystallizes before pyroxene; but if both minerals have a high content of iron then the pyroxene crystallizes first. Observations made by H. G. Backlund and D. Malmquist among the basalts of Greenland are in agreement with the conclusions of Bowen and Schairer.

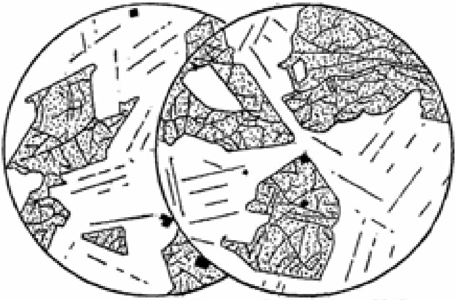

Fig. 18. Olivine crystals enclosing and moulded upon anorthite; island of Rum. (After A. Harker.) Magnification 20.

With the third clause of the Rosenbusch rule, as far as it can be considered apart from the preceding ones, there is less fault to be found. The crystallization of lime-rich plagioclase before soda-rich plagioclase is a necessary consequence of the course of the freezing curves (page 93 and *Fig.* 11). The earliest-formed plagioclase contains more anorthite than the liquid with which it is in contact, and subsequently reacts with the liquid as already described. If reaction takes place without interruption, the resulting crystals will either be homogeneous or else their composition will vary in a continuous manner from centre to margin. But many plagioclase crystals exhibit an oscillatory succession of zones which are alternately more sodic and more calcic. The passage from one zone to the next may be either gradual or abrupt, and the course of crystallization has often been interrupted by periods of corrosion. F. Homma has furnished an exact study of a plagioclase crystal from an andesitic lava of Japan, which shows fifty-one zones separated into five groups by corrosion surfaces. It is certain that such a crystal has had a very complex history.

Many attempts have been made to explain oscillatory and reversed zoning in feldspars. In the case of the Rapakiwi granites of Sweden and Finland, where the large orthoclase crystals have mantles of oligoclase around them, it has been assumed that the orthoclase crystals, after growing to a certain size, sank, or by some other means came into a layer of magma in which the crystallization of plagioclase had not been completed, and there received their mantles.

A more probable explanation of this phenomenon has been given by H. H. Thomas and W. Campbell Smith, who have shown that in the granite of Trégastel, Brittany, the formation of mantles of oligoclase about the alkali-feldspar is a consequence of cross-assimilation between the granitic magma and a basic igneous rock which it invaded. The matter was discussed in chapter V.

J. E. Hibsch described large plagioclase crystals in basalts of the central Caucasus, which show a core of the composition Ab_1An_1, followed by a sharply defined outer zone of which the innermost layer is Ab_2An_3, which in turn becomes more sodic towards the outside. Thus there is a reversal of the normal order in passing from core to mantle, but the usual order is followed in the successive zones of the mantle. Hibsch explains this by supposing that the cores were formed in an andesitic magma which at a later stage was mixed up with a basaltic magma rich in anorthite; but such an explanation is simply a confession of ignorance as to what happened. Other

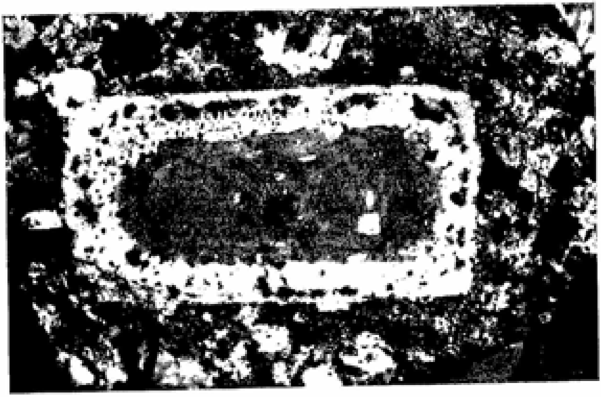

PLATE IV(A).—A large inset of microcline (grey) with mantle of sodic plagioclase (white), in granite. Cape Town, South Africa. (See *Bull. Geol. Soc. America*, vol. 60, 1949, p. 1213.)

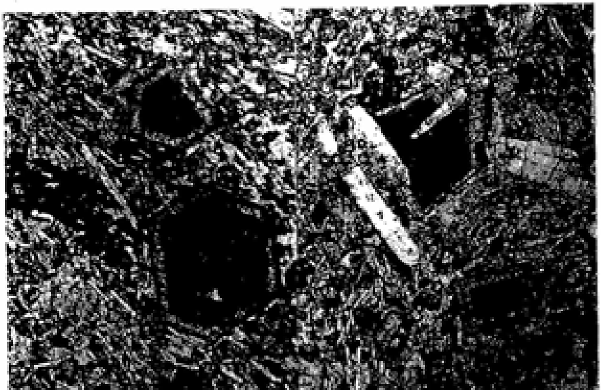

PLATE IV(B).—Nepheline crystals in phonolite, showing zonal structure due to variation in silica content. Bohemia. (See *Amer. Mineralogist*, vol. 24, 1939, p. 508.)

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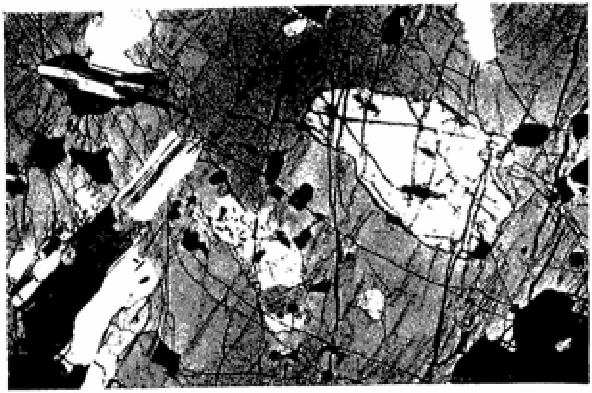

PLATE V(A).—Poikilitic texture. A large plate of hornblende (forming the entire background) is full of corroded relics of plagioclase (showing polysynthetic twinning), pyroxene, and iron ore. Cortlandt complex, New York State.

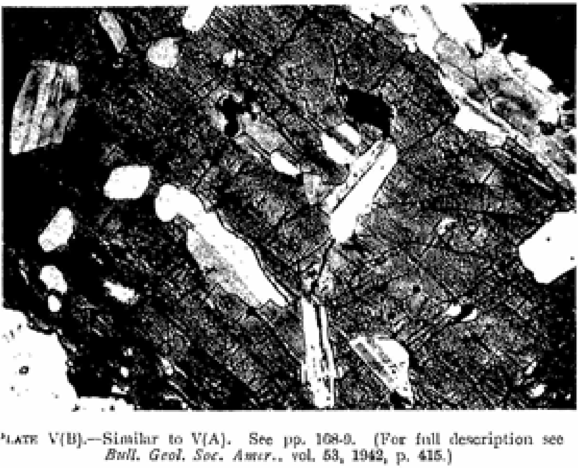

PLATE V(B).—Similar to V(A). See pp. 108-9. (For full description see *Bull. Geol. Soc. Amer.*, vol. 53, 1942, p. 415.)

[*To face p.* 113.

attempts to explain oscillatory zoning in feldspar depend upon fluctuations in the temperature of the magma or in the concentration of fugitive constituents; upon a recurrent condition of supersaturation in the magma; or upon a lack of balance between the rate of diffusion and the rate of crystallization. The problem of zoned crystals has been discussed in terms of physical chemistry by J. Phemister and by E. S. Hills.

Between feldspars and feldspathoids the relations are irregular. Sodalite is sometimes enclosed by orthoclase, sometimes interstitial to it, and again the two minerals may form micrographic intergrowths. This is also true of nepheline. In the green foyaite of Pilansberg, Transvaal, large insets of green microcline have crystallized before nepheline, while in the white foyaite of the same place the feldspar insets are full of little enclosures of nepheline and sodalite (S. J. S., 3). Leucite sometimes crystallizes at a very early stage, for some of the lavas of Vesuvius contain large crystals of leucite at the time when they are poured out of the volcano. Analcime, in the few rocks which contain it, is usually the very last mineral to form; yet the "blairmorite" of Canada, which holds over seventy per cent of analcime, has large crystals of that mineral set in a fine-grained groundmass consisting of smaller analcime crystals together with pyroxene, nepheline, orthoclase and melanite. Porphyritic lavas with insets of analcime have been described in the Highwood Mts. of Montana (Larsen and Buie).

As regards quartz, the Rosenbusch rule is very often at fault. Apart from the fact that many rhyolites and pitchstones show crystals of quartz in a glass-base, we have evidence even from granites to show crystallization of quartz before feldspar, or at least before the completion of the feldspar period. A. Harker and J. E. Marr described a granite-porphyry at Wasdale Crag, Westmorland, in which the larger feldspars enclose perfectly formed quartz crystals rising to one-fifth of an inch in diameter. In the Rapakiwi granites of Sweden and Finland the quartz typically appears as perfect little bipyramids enclosed in feldspar. A. Lagorio, by examining the composition and order of crystallization of many rhyolites and porphyries, brought out the approximate rule that in rocks of moderate acidity (about 55 to 65 per cent) the glass base is more acid than the entire rock, while in very acid rocks the base is less acid. In other words, quartz begins to crystallize early from highly siliceous rocks and late from those of moderate acidity. J. H. L. Vogt (2) confirmed this, and showed that the boundary lies at about 74 per cent of total silica.

CONCLUSION.

The only conclusion we can draw from the foregoing discussion is that there is no such thing as a constant order of crystallization, applicable to all magmas or even to one magma under different physical conditions. The Rosenbusch rule has no theoretical or experimental justification, and even on the basis of observation it cannot be upheld. Yet the strange fact remains that the rule seems to be applicable to a large number of rocks. How can this be?

The explanation lies partly in the matter that we discussed at the beginning of this chapter, namely, the inadequacy of the microscopic method of study. If little crystals of mineral A are seen to be enclosed in larger crystals of mineral B, does that really demonstrate that the period of crystallization of A was earlier than that of B? It depends, as Bowen showed, on the position of the enclosures within the host. If an enclosed grain lies at the centre of its host, then it is probably safe to say that the former crystallized first; but that conclusion applies only to the grains actually studied. It does not prove that all of A crystallized before all of B, for the same accidental enclosure of one mineral within the other might happen during simultaneous crystallization of both minerals at a cotectic surface. Again, if an enclosed grain of A lies just within the margin of B, then it is possible that crystallization of A may only have begun as that of B was finishing. Thus, in such a case as we have pictured, where large crystals of an abundant mineral enclose smaller crystals of a less abundant one, the enclosed mineral may have crystallized entirely before the host mineral, or at the same time, or partly after it. And if the enclosed mineral was formed either by exsolution, as albite in microperthite, hypersthene in clino-pyroxene, sphene in melanite; by inversion, as nepheline and orthoclase replacing leucite; or by the crystallization of occluded drops of liquid, then it may be wholly younger than its host.

There is indeed good reason for thinking that such diverse minerals as plagioclase, augite, hornblende and biotite have often crystallized *at the same time* in eruptive rocks. We saw that T. F. W. Barth has established (at least provisionally) the position of the cotectic surface for plagioclase and pyroxene in basalts. S. Tsuboi has made the curious discovery that in Japanese granites the refractive indices of biotite and hornblende show a progressive increase as the albite content of the associated plagioclase increases. One can hardly avoid the conclusion that all three minerals were growing in the magma at the same time.

But not all the discrepancies between theory and observation can be ascribed to the inadequacy of the microscopic method of study. Some of them are undoubtedly to be explained by the action of residual solutions of magmatic origin, which bring about recrystallization and replacement of some of the early-formed minerals. This matter will be discussed in chapter X. These residual solutions are often rich in soda, and they can be shown to have reacted with plagioclase, replacing lime-feldspar by albite. It is highly probable that other sodic minerals such as ægirine, soda-amphiboles, cancrinite, eudialyte and pectolite were introduced by residual solutions and that they crystallized in the interstices of a rock that was already almost completely solid, to some extent replacing the earlier minerals; that is, they did not crystallize in the magmatic stage at all but in the hydrothermal stage. This may also be true of some of the lime minerals such as calcite, wollastonite, melanite, vesuvianite and even of some sphene and apatite. It is conspicuously true of tourmaline, which sometimes replaces feldspar completely. The residual solutions seem to have a remarkable power of penetrating minute openings such as cleavage planes and twinning planes, in which they may deposit crystals which are actually younger, although they appear to be older, than their host.

When we make allowance for all these disturbing factors, we must conclude that microscopic study alone is not always a safe guide to the cooling history of a rock. On the other hand, the experimental study of anhydrous melts in the laboratory may be almost equally misleading since the influence of water is excluded. This leaves us in a quandary from which the only way of escape is the realization that observation must go hand in hand with experiment if we are ever to understand the whole story of the genesis of an eruptive rock. The first step is to realize that the emplacement and consolidation of a large body of magma in the earth-crust is a highly complex process, involving repeated changes of pressure and temperature, accompanied by the loss of some constituents to the wall-rock or the atmosphere and the gain of other constituents from the wall-rocks; that conditions of mixing are far from perfect, and that different parts of the same body of magma may in consequence follow different courses of development; that these partial magmas may be brought together again by later movements, and that at any stage in their cooling they may receive fresh increments of magma or magmatic solutions from below. Under such conditions as these there can be no permanent equilibrium in the system until it has cooled almost to atmospheric temperature. It is certain that no arbitrary rules of crystallization can ever cover such a complex history.

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The intellectual progress of mankind has been a continual disentanglement and simplification, leading to increased grasp and power.—H. G. Wells.

CHAPTER VIII

COMPATIBLE AND INCOMPATIBLE PHASES

It is a familiar observation to petrographers that leucite, nepheline, olivine and a few other minerals are almost never found in rocks which contain free quartz or tridymite. The implication is that the molecules of these minerals are unstable in the presence of free silicic acid, under such conditions as are realized in rock-magma, because they can combine with it to form higher silicates. The writer, in 1913, described such minerals as *unsaturated* with silica, while those minerals which are capable of crystallizing from rock-magma even in the presence of an excess of silica were said to be *saturated* with regard to silica. A saturated mineral is not necessarily a silicate at all; even certain oxides, titanates, and phosphates are saturated in the above sense, since they are stable in the presence of an excess of silica under magmatic conditions. Saturated minerals and quartz are compatible phases; unsaturated minerals are incompatible with quartz in a normal eruptive rock. The common saturated and unsaturated minerals are named in the following lists:—

Saturated	*Unsaturated*
All feldspars	Leucite
All pyroxenes	Nepheline
All amphiboles	Sodalite, haüyne, nosean
All micas	Cancrinite
Tourmaline	Analcime
Fayalite (iron olivine)	Olivine (magnesian)
Spessartite	Melanite (andradite)
Almandine	Pyrope
Sphene	Perovskite
Zircon	Melilite
Topaz	Corundum
Magnetite	Calcite
Ilmenite	
Apatite	Spinel?

The experimental studies of silicate melts which have been made in the course of the last twenty years have justified this grouping of the rock-forming minerals; and the actual temperatures at which certain of these minerals are stable or become unstable (in the crucible) have been determined. While it is of the greatest importance to have experimental confirmations of the petrographic evidence, one must always remember the differences between magma and dry melt which were pointed out in an earlier chapter. Some minerals form readily from dry melts; others require water for their generation. The presence of fugitive constituents in the magma affects every equilibrium temperature. Thus, from our point of view, it is the petrographic evidence that is of the foremost importance; the chemical evidence, in the present stage of investigation, is only confirmatory. In the following discussion we shall make use of both kinds of evidence.

That quartz and magnesian olivine are incompatible in eruptive rocks is shown by a multitude of observations, of which only a few can be mentioned here. The following table shows the relation between silica content and mineralogical development in a series of basalts in the Eureka district of Nevada (A. Hague).

Number	*Silica*	*Olivine*
1.	49·23	Rich in macroscopic secretions
2.	51·86	Rich in microscopic secretions
3.	57·42	Abundant, microscopic
4.	58·06	Easily recognized under microscope
5.	58·26	Only a trace
6.	58·60	None detected
7.	58·64	Detected under the microscope
8.	59·51	None detected
9.	59·64	None detected
10.	60·11	None detected

G. W. Tyrrell has described a series of dolerite intrusions at Kilsyth, Scotland, which contain olivine when quartz is absent, but hypersthene in all cases where quartz is present.

When a normally olivine-bearing rock makes contact with siliceous country rocks, it is often observed that reaction has taken place between them, leading to the disappearance of the olivine. Sills of olivine dolerite at Nordingra, Sweden, lose their olivine and become quartz-bearing when they come into contact with sandstone or granite (A. G. Högbom). Olivine grains in the Palisade diabase of New Jersey develop mantles of enstatite near the sandstone walls of the intrusion (J. V. Lewis). Many examples of this reaction are furnished by A. Lacroix in his volume on the inclusions of volcanic rocks; quartzose enclosures in olivine-bearing lavas have been partly assimilated by the lava, which in the neighbourhood of the enclosures loses its olivine and develops pyroxene instead.

A certain number of rocks undoubtedly carry both a little olivine and a little quartz. In some cases the explanation is that the quartz grains are of foreign origin, as in the instances just mentioned. Other well known examples are the "quartz-basalts" which J. S. Diller described from New Mexico. A recent re-examination of these rocks by R. H. Finch and C. A. Anderson has shown conclusively that the quartz is a foreign substance. A more general explanation of this abnormal association was supplied by N. L. Bowen and O. Andersen, who showed that in the crucible enstatite dissociates at 1557° C. into forsterite and silica; thus above the temperature of transformation olivine (forsterite) is compatible with free silica, but below that temperature it reacts with silica to form enstatite (clino-enstatite in the laboratory experiments). From a melt of the composition $MgSiO_3$ forsterite begins to crystallize at 1580° C. and continues to crystallize until the temperature falls to 1557° C.; thereafter forsterite is resorbed and clino-enstatite crystallizes until the whole mass is solid. Now, such a temperature as 1557° C. is well above the normal range of the magma, but this temperature must be lowered by the presence of other components in the magma and especially by the presence of some iron in the olivine; so it is in the "Bowen-Andersen effect" that we may seek the explanation of the occasional conjunction of olivine and quartz in rocks. If the magma is cooled rapidly after olivine has begun to crystallize, then the reaction between olivine and silica is prevented and we have a rock composed of a few crystals of olivine with or without other minerals embedded in a siliceous glass. Lavas of this character have been described by H. S. Washington (1) at Monte Ferru, Sardinia, and by H. H. Robinson at San Francisco

Mountain, Arizona. A similar effect may be produced by the olivine crystals becoming enclosed in pyroxene and so prevented from reacting with the siliceous residual magma; the latter, if it crystallizes completely, must then yield some quartz.

A later investigation by Bowen and J. F. Schairer (1) of the system MgO-FeO-SiO_2 has confirmed Bowen's earlier conclusions and has furnished no support for the idea that ordinary basaltic olivine can form in equilibrium with silica.

As regards fayalite, or iron-olivine, the circumstances are different. Fayalite appears in association with quartz or tridymite in a number of lavas and even in some granites and syenites. H. S. Washington (2) described acid lavas from the island of Pantelleria which contain a little fayalite, and the writer has found this mineral to be quite common in the quartz-trachytes of Kenya. Wager and Deer have shown that olivine containing more than 60 per cent of fayalite is stable in presence of quartz in the remarkable "ferrogabbro" of East Greenland. V. Sobolev has described pegmatitic quartz-fayalite diabase in Siberia, which holds 8 per cent of an olivine containing 80 per cent of the fayalite molecule, along with 15 per cent of quartz. According to the field evidence, therefore, fayalite is not incompatible with quartz. This need not surprise us, since the pure compound $FeSiO_3$ is unknown in nature. Bowen and Schairer (1) were able to prepare pyroxenes containing as much as 87 per cent of this molecule, but they failed in the attempt to synthetize the pure compound. On the other hand, they demonstrated the existence of a fayalite-tridymite eutectic at 1178° C.

We see, then, that the petrological evidence of the incompatibility of magnesian olivine with free silica, and of the compatibility of fayalite, has been completely confirmed by experiment. All common olivine is magnesian and therefore unsaturated.

Coming now to leucite, we find very few records of the association of leucite with quartz. In some of them the quartz is demonstrably of foreign origin. A. Lacroix described several occurrences of this nature. Thus the "leucitite néphélinique" of Korretzberg, Rhenish Prussia, normally contains no feldspar, but where it holds quartzose enclosures it develops feldspars at the expense of its leucite and nepheline. Similar phenomena are described at Bertrich in the Eifel and at the Kaiserstuhl in Baden. In other cases, the material is the mysterious substance called "pseudoleucite," and it proves nothing at all about the stability of fresh leucite. The remarkable "leucite-granite-porphyry" which E. Hussak found at Serra de Caldas, Brazil, contains pseudo-crystals of sharp trapezohedral form, composed of orthoclase and nepheline with a little spinel and corundum, set in a fine-grained base of granitic composition. Quartz is never found in the pseudo-crystals, or nepheline in the base, and a narrow mantle of orthoclase laths separates the core of each pseudo-crystal from the siliceous base. Hussak supposed that a granophyric magma had invaded and broken up a leucite-bearing rock; but the question requires further investigation in the field.

Experiments made in the last decade of the 19th century by C. and G. Friedel, and afterwards repeated by E. Baur and by C. N. Fenner, showed that either leucite or orthoclase could be produced, from melts of suitable composition, by varying the proportion of silica. G. W. Morey and N. L. Bowen have since proved that leucite bears the same relation to orthoclase that olivine (forsterite) does to enstatite. Orthoclase heated to 1170° C. dissociates into leucite and a siliceous liquid; consequently leucite is compatible with free silica above 1170° C., but not below that temperature. It is therefore possible that leucite may crystallize at an early stage from some highly potassic magmas and afterwards be transformed into orthoclase; or, if the rate of cooling was rapid, some crystals of leucite might be preserved in a base of siliceous glass. Such a rock was discovered by W. Cross in Wyoming and named wyomingite. It consists of phlogopite, leucite, diopside and nosean with some interstitial glass. Chemical analysis of the rock shows that there is enough silica present to have converted all the leucite into orthoclase if the conditions of cooling had been favourable.

The systems nepheline-leucite-silica, anorthite-leucite-silica, and diopside-leucite-silica have now been worked out experimentally by J. F. Schairer and N. L. Bowen. The equilibrium diagram for each of these systems shows that the leucite field is completely separated from the tridymite field by an orthoclase field; from which it follows conclusively that leucite and quartz or tridymite are incompatible under conditions of equilibrium.

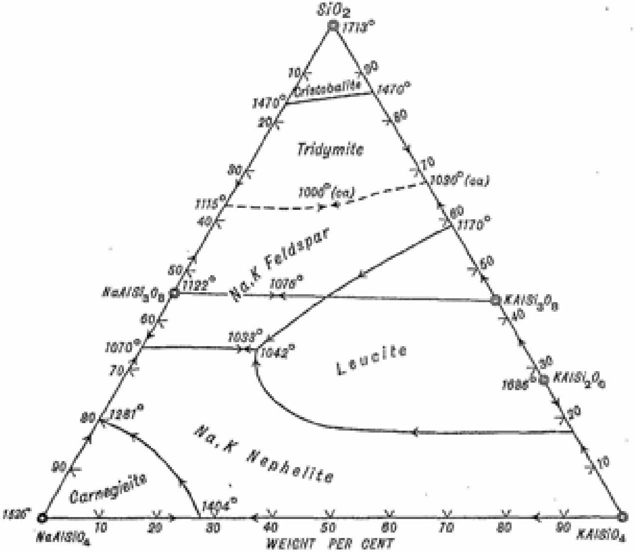

FIG. 19. Equilibrium-diagram for the ternary system, $NaAlSiO_4$—$KAlSiO_4$—SiO_2 (arrows indicate falling temperature; paired concentric circles indicate the composition of a compound). (Bowen and Schairer.)

In the case of nepheline, we have again very few records of quartz and nepheline appearing in the same rock, and all of them are explained as the result of contamination of the magma. R. Heinrich describes some basaltic lavas containing both nepheline and olivine, which cut through granite near Rothau in Bohemia. Fragments of the granite which are enclosed in the basalt have been much corroded, and in their neighbourhood nepheline does not appear at all in the basalt, while the olivine grains have been attacked and converted partly or wholly into pyroxene. In other cases it is the quartz grains which are changed. Lacroix records a case in the Vosges Mts. where quartz grains enclosed in nephelinite have been converted into soda-zeolites. F. Cornu observed that fragments of granite enclosed in nepheline basalt at Salesl, Bohemia, have their quartz changed into apophyllite; this observation has been repeated by A. Scheit at Teplitz.

We have already seen (page 85) that the nepheline rocks of Alnö, Sweden; those of Iivaara, Finland; and some others have been acidified at their junctions with granite or gneiss, resulting in the non-appearance of nepheline in the contact rocks. The behaviour of the nepheline-rich pegmatites of the island of Seiland, Norway, is also significant. According to T. F. W. Barth these nepheline-pegmatites cut the peridotites of the island, but are never found cutting the acid gneisses. All the pegmatites in the gneiss are feldspathic.

A single record of idiomorphic nepheline in a highly acid granophyre at Horns Nek, Transvaal, is mistaken. The mineral is cordierite, not nepheline (S. J. S., 2).

From the experimental side we have the long-standing evidence of Friedel, Baur, and Fenner that either albite or nepheline may be produced at will by adding a larger or smaller quantity of silica to melts containing soda and alumina in the correct proportions. Referring again to the work of Schairer and Bowen (also Bowen and Schairer (2)), we find that in the system nepheline-albite-silica the albite field completely separates the fields of the other two phases, which are consequently wholly incompatible under equilibrium conditions.

Nepheline and leucite are not only incompatible with free silica, but they are incompatible with free magnesium metasilicate. The magnesian pyroxenes enstatite, hypersthene, clino-enstatite and clino-hypersthene, as well as the pigeonitic pyroxenes, are never found in rocks that contain feldspathoids, although magnesium orthosilicate (olivine) is very common in such rocks. This clearly indicates that the nepheline field and the enstatite field, in the system olivine-nepheline-silica, are separated from each other by the feldspar field; and that the same condition obtains in the system olivine-leucite-silica. Thus we find plagioclase-enstatite rocks, plagioclase-olivine rocks, feldspathoid-olivine rocks, but never feldspathoid-enstatite rocks. But if molecules of calcium metasilicate are present the case is altered, for calcium metasilicate combines with magnesium metasilicate to form the very stable compound diopside, $CaMgSi_2O_6$; under these circumstances any deficiency of silica in the magma falls upon the feldspar, because the orthosilicate $CaMgSiO_4$ (which bears the same relation to diopside as olivine does to enstatite) is rarely stable under magmatic conditions. Thus, although the feldspathoids are incompatible with enstatite and the other magnesia-rich pyroxenes, they are quite compatible with diopside and augite.

The remaining feldspathoids, sodalite, haüyne, nosean and cancrinite are closely related in composition to nepheline and are nearly always associated with it in rocks. There is therefore an initial probability that they will behave as nepheline does towards an excess of silica in the magma. There is little direct evidence of acidification, but no lack of indirect evidence of the incompatibility of these minerals with quartz. So far as is known, none of them has ever been found in a quartz-bearing rock; their home is among the nepheline syenites and phonolites. Analcime is in a different position, for some of it seems to be primary whereas other examples are definitely of secondary origin, occurring along with zeolites in fissures and vapour cavities. In the latter case the mineral may be associated with rock-crystal, amethyst or chalcedony, or it may occupy cavities in an acid rock; but those occurrences of analcime which are considered to be magmatic are all in quartz-free rocks, most of which hold either olivine or nepheline.

Melilite is confined to a very uncommon group of lavas in which olivine is nearly always present in abundance, and in which the silica content scarcely rises above 35 per cent. It is also well known as a constituent of very basic furnace slags. Melilite is a very complex solid solution of rather uncertain composition, but one of the simple molecules composing it is akermanite, $Ca_2MgSi_2O_7$. This is one of the phases formed in the system CaO-MgO-SiO_2 which was studied experimentally by J. B. Ferguson and H. E. Merwin. In this system the stability field of akermanite does not touch the silica field at all, but is separated from it by the fields of diopside and pseudo-wollastonite. Another of the simple molecules that enter into melilite is gehlenite, $Ca_2Al_2SiO_7$. This appears as one of the phases in the system lime-alumina-silica which was studied by G. A. Rankin and F. E. Wright. The stability field of gehlenite is separated from that of silica by the fields of wollastonite and anorthite.

Perovskite is a very rare mineral in eruptive rocks; it is commonly associated with melilite and nepheline, but avoids the more silicic rocks, in which its place is taken by sphene. It was shown by A. Bourgeois, more than fifty years ago, that sphene is formed in highly siliceous melts and perovskite in melts of low silica content.

The lime-iron garnet, melanite or andradite, is only known in quartz-free syenites and phonolites, and is probably due in every case to the assimilation of limestone by the magma; that is, it is a product of desilication. The melanite syenite of Assynt, Scotland, occupies an intermediate place between a quartz-syenite and a limestone, and melanite only appears when quartz has disappeared. (S. J. S., 3.) This matter is discussed at greater length in chapter XVI.

The magnesian garnet, pyrope, is restricted to very basic rocks such as peridotites, pyroxenites and serpentines, which are generally free from feldspar. There is nothing in the known facts of distribution of either melanite or pyrope to suggest that these minerals are capable of forming in presence of an excess of silica.

In the case of corundum, the common association is with plagioclase (Transvaal; Ural Mts.), or nepheline (Canada; India), in quartz-free rocks. The important corundum deposits of the Transvaal are in plagioclase-biotite pegmatites, and A. L. Hall remarks that "not a single example was observed where a corundiferous reef also carried quartz." It is possible, however, that at the very low temperature under which crystallization takes place in some pegmatites, corundum and quartz may form a stable association, just as calcite and quartz do below 500° C. Thus Adams and Barlow remark that in Ontario a small amount of quartz has sometimes been found in the corundiferous reefs; and Barlow states that in a pegmatite in Colorado "the corundum seems to favour association with quartz." But E. S. Larsen contends that these coarse corundum pegmatites are low-temperature hydrothermal products, and H. W. Tomlinson has described the production of corundum from plagioclase by the action of magmatic solutions.

As regards spinel there is little evidence, but C. E. Tilley notes that in the garnet-gneisses of South Australia spinel is never found in contact with quartz, but is always separated from it by a shell of garnet or sillimanite.

Calcite has been recognized as a constituent of nepheline-syenites in Ontario, Sweden, and elsewhere, and of quartz-bearing aplites at Gowganda, Canada. The reaction between calcite and silica has already been discussed (p. 57), and in the light of the experimental evidence it would seem to be correct to regard calcite as an unsaturated mineral above 500° C., and a saturated one below that temperature, under atmospheric pressure.

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Man longs for causes, and the weaker minds, unable to restrain their hunger, often barter for the most sorry theoretic pottage the truth which patient enquiry would make their own.—JOHN TYNDALL.

CHAPTER IX

ERUPTIVE ROCK COMPLEXES

THE crystallization of a body of eruptive magma does not always yield a homogeneous body of rock. Indeed, there are so many factors tending to oppose homogeneity that it is rather a matter for surprise that many large bodies of rock have nearly the same composition throughout. But other bodies are visibly complex, and even the most uniform-looking granite or gabbro will show a certain degree of variability when studied in detail. Many examples of lack of homogeneity were discussed in Chapter V, where they were ascribed to reaction between the magma and its walls. There is no doubt that assimilation of country rocks is one of the most frequent causes of heterogeneity in eruptive rocks, but it is not the only cause.

The variations of composition observed within bodies of eruptive rock are all too frequently referred to a mysterious process called "magmatic differentiation." This doubly-unhappy expression implies that rock-magma has a capacity, not possessed by common liquids, for spontaneously changing its composition; and it carries the further implication that all the variations observed in rock masses are the result of a single process. Neither of these implications is correct. Rock magma is subject to the same laws as other solutions, and if a homogeneous magma gives a heterogeneous product it must be for reasons with which other solutions have made us familiar. In addition to the assimilation of country rocks, which we have already discussed, the following factors are capable of producing heterogeneity in bodies of eruptive rock:

(1) Density stratification in the liquid magma;
(2) The mingling of two or more magmas;
(3) Differential movement of crystals and residual liquid during crystallization;
(4) Varying concentration of the fugitive constituents;
(5) Post-magmatic alteration and replacement. (Discussion in Chapter X.)

(1) DENSITY STRATIFICATION IN THE LIQUID MAGMA

There are two ways in which a density-gradient may be set up in a body of liquid that has not begun to crystallize. One of them depends on the separation of immiscible liquid phases; the other is independent of that complication. The theory of immiscible liquid phases has always proved very attractive to petrologists who have observed the frequent association of gabbro with granite or granophyre, and basalt with trachyte. Indeed, the spontaneous splitting of the magma into distinct liquid phases seems to have been the original idea underlying the use of the term magmatic differentiation.

It is known that certain organic liquids, such as phenol and nicotine, are completely miscible with water within a certain range of temperature, but only partly miscible at higher or lower temperatures. Applying this idea to rock-magma, it has been supposed that with falling temperature the magma will tend to split into two liquids, the lighter of which will rise and float upon the heavier. After crystallization has taken place, it is supposed that we should then have a lighter rock, such as granite, resting upon a denser rock such as gabbro, with a sharp junction between them. Since two fluids of different composition would be unlikely to finish their crystallization at the same temperature, one of these rocks might even show an intrusive relation towards the other.

This hypothesis is attractive in its simplicity, but it must be said that it is quite unsupported by experimental evidence. Many silicate melts of composition very similar to that of eruptive rocks have been studied in the laboratory and they have shown no tendency whatever towards immiscibility. A. S. Ginsberg and K. S. Nikogosyan melted diabase and granite together in various proportions and found them completely miscible. Certain mixtures of lime, magnesia and other oxides with an excess of silica, which were studied by J. W. Greig, did indeed give immiscible liquid fractions at temperatures in the neighbourhood of 1700° C.; but the composition of these melts is very different from that of any known eruptive rock and the temperature much higher than any recorded magmatic temperature. We seem justified in concluding, as Greig himself did, that his observations "do not offer support to the idea of differentiation by the splitting of a magma into two immiscible liquids." D. P. Grigoriev has demonstrated immiscibility in melts of silicates with fluorides, but these melts differ very greatly from natural magmas.

N. L. Bowen (1) has strongly opposed the hypothesis of liquid immiscibility in natural rock-magma. He points out that the unmixing of two liquids takes place by the slow growth and coalescence of many small globules, and that one might reasonably expect these globules to have been preserved in some instances. It is most unlikely that all the globules would collect into a separate and distinct layer, leaving no evidence of their existence. If immiscibility were a common phenomenon in nature, such globules would be common in glassy lavas, but this is not so. In a single instance of a "globule-bearing" quartz-porphyry, described by J. L. Tanton, it was demonstrated by J. W. Greig that the supposed globules are actually spherulites formed by partial crystallization of an already solid lava. F. Levinson-Lessing described, as long ago as 1884, a glassy diabase at Jalguba in Karelia which contains globules (varioles) of acid glass. In 1935 he again drew attention to this phenomenon, claiming that it illustrates unmixing in the liquid state. The variolitic rocks are associated with, and form a textural variety of, a body of normal diabase. Little grains and idiomorphic crystals of pyroxene are scattered through both kinds of glass, but are more abundant in the basic kind. Apart from these the acid glass is shown to be equivalent in composition to 23 parts of quartz, 75 parts of oligoclase, and 1 part of magnetite; and the basic glass to 70 parts of bytownite, 24 parts of olivine, and 5 parts of magnetite. In short, the varioles are of dacitic glass, the groundmass of olivine-basaltic glass. Since pyroxene crystals appear both in the varioles and in the groundmass, it is suggested that the splitting of the original magma into two immiscible liquids took place after the period of crystallization of pyroxene.

This very remarkable occurrence is unique, and the evidence produced certainly seems to justify Lessing's interpretation; yet doubts and questions present themselves. If unmixing was able to take place even in the rapidly cooled glassy facies, why is there no indication of the same process or its results in the normal diabase? Lessing shows by chemical analysis that the diabase without varioles has exactly the same composition as the variolitic rock, from which it must follow that the former has not been affected at all by unmixing. This paradox must be resolved before Lessing's conclusion can be adopted.

In spite of a complete lack of encouragement from experimental chemistry, not a few petrologists still invoke liquid immiscibility to explain their field observations. F. F. Grout (1) has used this hypothesis to elucidate the differentiation observed in the gabbro lopolith of Duluth, Minnesota. The lower part of this great intrusive body is strongly banded; its average composition is olivine-gabbro but individual bands range in composition from anorthosite to peridotite. The upper division is comparatively homogeneous, consisting of various facies of feldspathic gabbro. Between the lower and upper divisions, which differ slightly in age, there is a zone of "red rock," and at the top of the upper division there is another such zone. The red rock is an acid granite or granophyre, and the passage from gabbro to red rock is "somewhat abrupt," the gray gabbro rapidly giving place to a bright red rock of totally different character. Grout ascribes the banded structure of the lower gabbro to the action of convection currents in conjunction with crystal settling; but to explain the abrupt separation of granophyre from gabbro he suggests the splitting of the magma into immiscible fractions, of which one may have been "wet" or highly charged with water and the other relatively "dry" or anhydrous.

On the island of Rödön, Sweden, H. von Eckermann has described a dyke six metres wide, cutting Rapakivi granite. This dyke shows a lateral gradation from diabase, in the middle of the dyke, through monzonite to a granite almost identical with the enclosing Rapakivi. This is clearly not a case of successive intrusions of basic, intermediate and acid magmas. Von Eckermann conceives that the Rapakivi granite represents the solidified top of a body of magma which separated into an upper layer of acid magma and a lower layer of basic magma; that a fissure opened through the granite and reached down to the underlying basic magma which was still fluid; and that basic magma rose in the fissure, but as it rose it was succeeded, without any interruption of flow, by a further supply of acid magma generated by the continued fractionation of the basic magma.

It will be observed that in the first of these two illustrations there is an abrupt passage from gabbro to granite, but in the second there is a perfect gradation from diabase to granite. It is an accommodating theory that can be used to explain two contradictory observations!

But without the separation of immiscible liquid fractions, it is still possible that in a long vertical column of liquid the composition may vary progressively from top to bottom in consequence of the influence of pressure upon solubility; as the pressure increases steadily from top to bottom of the column, the concentration should vary correspondingly. This phenomenon has been studied in the case of salt solutions, but the effect is found to be small and there is no reason to suppose that it would be any more significant in a silicate magma than in any other solution. If it were, then it would be possible for partial magmas of slightly different composition to be withdrawn from different levels in the same reservoir.

An unequal distribution of temperature might have a similar effect. Since the osmotic pressure of a dissolved substance is proportional to the absolute temperature, a pressure gradient would be established between the hotter and the cooler parts of a body of magma, which would drive the less soluble molecules towards the cooler parts and increase the concentration there. It has been shown that a difference of temperature of 100° C. may cause a difference of concentration of as much as seven per cent. But all processes involving molecular flow are excessively slow, and long before an appreciable difference in concentration could be produced it is likely that convection currents would have equalized the distribution of temperature throughout the magma. (But see W. Wahl, 1946.)

It is important to observe that under the conditions discussed in either of the last two paragraphs the composition and the density of the magma would show a *continuous* variation. Now a rude stratification more or less in accordance with density is exhibited by many large eruptive masses, but the variation is seldom a continuous or regular one. This point is convincingly illustrated by *Fig.* 20. which shows the variations of density and composition within the Lugar sill, in Ayrshire, according to G. W. Tyrrell.

W. H. Collins has recently revived the theory that a magma of intermediate chemical composition may give rise, by a spontaneous change taking place in the magma itself while it is still wholly liquid, to a more acid magma of the composition of granite and a more basic magma of gabbroic composition. Collins finds support for this view in the geology of the famous nickel-bearing intrusive at Sudbury, Ontario. As this intrusion has been used to illustrate other theories of differentiation, a brief description is necessary.

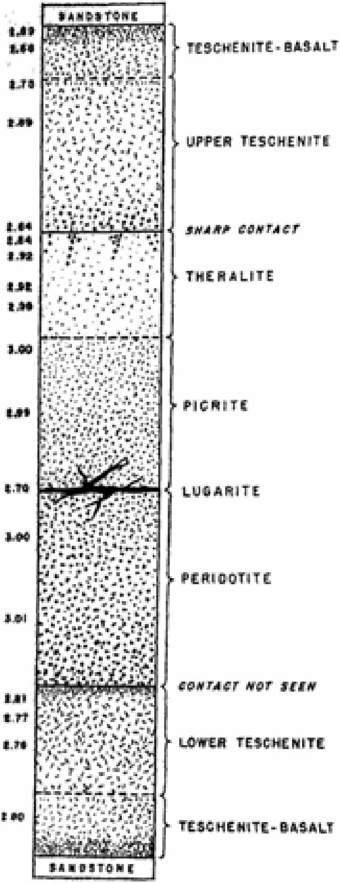

Fig. 20. A vertical section through the Lugar sill, Ayrshire. (G. W. Tyrrell.) Numbers represent specific gravity.

The Sudbury intrusion is a great curved sheet or lopolith, with a boat-like outcrop some 37 miles long and 17 miles wide. It is formed essentially of two kinds of rock, granophyre (or micropegmatite) above and gabbro (or norite) below, the combined thickness amounting to about 1·5 miles. The former rock is mainly composed of feldspar and quartz, with only a small proportion of biotite, hornblende, and other dark minerals; the specific gravity of this rock may be as low as 2·67. The gabbro contains plagioclase, pyroxene (often altered to hornblende), biotite, and other minerals in minor proportions; its specific gravity rises from 2·7 to more than 2·9. The junction between the two kinds of rock is not always well exposed, which may be the reason why there has been much argument about its character. The earliest students of the area (T. L. Walker and A. P. Coleman) believed that there was a gradual transition from one rock to the other. T. C. Phemister (1) in 1925 made a careful study of the passage zone and found it to be very narrow, the maximum width 286 yards, the minimum almost nothing. From this Phemister concluded that the gabbro and the micropegmatite were formed from separate magmas, the second following the first before it had completely solidified. After a fresh study of the evidence, Coleman, Moore and Walker reaffirmed in 1929 their opinion that the norite and the micropegmatite were formed from one body of magma which has undergone differentiation in place, while still completely fluid. Collins (1934) confirmed Phemister's observation of the narrowness of the passage zone, but nevertheless continued to hold, with Walker and Coleman, that "The norite and micropegmatite were formed in place by gravitative separation from a single body of molten rock matter (magma) that was homogeneous when intruded," and further that "separation of the two layers had been completed, or nearly completed, before crystallization began and must therefore have taken place in the liquid state."

Except for the words "gravitative separation" we are given no clear picture of the process involved. Nothing is said about immiscible liquid phases, and since Collins insists upon the presence of a transition zone between the two layers the hypothesis of immiscibility would seem to be ruled out.

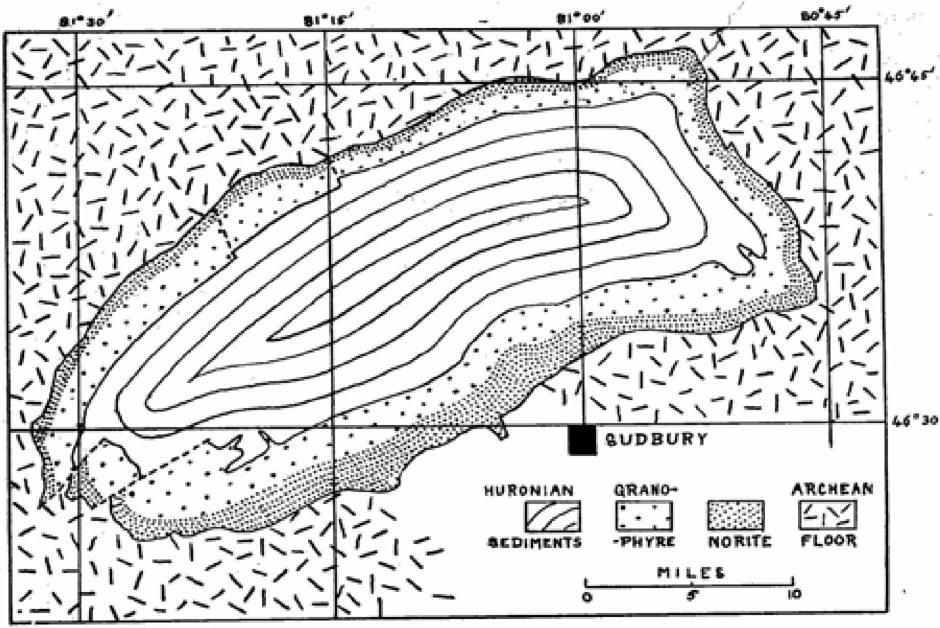

FIG. 21. Plan of the Sudbury lopolith (after W. H. Collins).

Yet it is shown that the variation in density and composition, from the top to the bottom of the sheet, is not of a continuous nature. Collins says "there is not a regular change in acidity and specific gravity across the irruptive. The SiO_2, CaO, MgO and K_2O curves all show pronounced and characteristic changes across the transition zone, but only minor departures from uniformity within the norite and micropegmatite." Thus the type of differentiation favoured by Collins is a kind unknown to chemistry, for it is neither an abrupt separation of immiscible phases nor a gradual change of composition. The evidence advanced by Collins in support of his theory is not strong enough to destroy the conviction that rock magma is governed by the same laws as other liquids and cannot change its composition arbitrarily.

In a later contribution to the subject (1935), T. L. Walker abandons the idea that the differentiation at Sudbury took place in a wholly liquid medium and tries to interpret it in terms of the sinking of heavy crystals from the magma during cooling. But Phemister in 1937 gave additional evidence in support of his claim that the acid member came into place slightly later than the basic one and that the intermediate zone has resulted from the "imperfect mixing of liquid acid member and partly solid basic member."

A strong argument against differentiation having taken place in an entirely liquid magma was furnished by N. L. Bowen; it is the restricted range of chemical composition in the glassy rocks (which represent former liquids) compared with that in the crystalline rocks. There are no glassy rocks corresponding to anorthosite, pyroxenite, or dunite.

2. The Mingling of Two or More Magmas

We have seen that T. C. Phemister holds the transition zone at Sudbury to be a hybrid product formed by an acid magma following a basic one before the latter had cooled completely, and soaking into the basic one to some extent. Hybridism in this sense is just a special case of assimilation, the conditions assumed being those most favourable to reaction. An earlier theory of hybridism, which is associated with the great name of Bunsen, attributed the variable composition of the lavas of Iceland to the intermingling of two primitive magmas, one of "feldspathic" and one of "pyroxenic" composition. The products of this commingling of magmas were called hybrid rocks by J. Durocher in 1857. The terms hybrid and hybridism have since been used in a double sense to cover both the intermingling of two liquids and also the contamination of liquid magma by reaction with its walls. Since we already have an adequate number of synonyms for contamination, it would be a gain if hybridism could be restricted to its original meaning; but perhaps it is too late to do this now.

A. Harker (1, 2) described several instances, in the island of Skye, of the injection of acid magma into basic rock while the latter was still hot. He says "the nature of the mutual reactions which have taken place among the several rocks indicates that they were intruded in somewhat rapid succession, and even in certain places that one was not completely solidified before it was invaded by another." This is nearly, though not quite, the hybridism of Bunsen and Durocher. H. H. Read has described a remarkable development of hybrid rocks in the counties of Sutherland and Ross, in the north of Scotland. They form small, elongated bosses in a country of granites and acid schists. At least two hundred of these bosses have been mapped and the average size is 200 by 50 yards. The rocks are very variable in appearance and composition, ranging from light-coloured granitic types to dark rocks made up of olivine, pyroxene, hornblende and biotite in various proportions. Whatever the nature of the dark minerals, the feldspars are always orthoclase and oligoclase. The commonest type of rock is intermediate between the light and dark extremes, and Read claims that these intermediate rocks have been formed by "a fairly complete mixing of granitic and ultra-basic magmas," followed in many cases by brecciation of the intermediate and basic hybrids and their envelopment in later acid and intermediate magma.

That the rocks described by Read are hybrids can hardly be doubted, but it is less certain that they were formed by the mixing of two magmas. It seems likely that the contamination of an acid magma by assimilation of a *solid* basic rock, coupled with the infiltration of granitic matter into the latter, might account for all the phenomena observed in the region. If so, these phenomena belong to assimilation rather than to hybridism in the older sense.

D. L. Reynolds has described the production of augite-biotite-diorite by reaction between a biotite-pyroxenite magma and a plagioclase magma, near Newry, Ireland. Every gradation is observed between biotite-pyroxenite, the same rock containing small pockets of andesine, and augite-biotite-diorite. It is claimed that the latter rock has resulted "from the permeation of the ultra-basic rock, at least after partial crystallization," by a plagioclase magma composed almost entirely of andesine. The writer, having studied the evidence in the field, finds himself unconvinced of the reality of the "plagioclase magma" and of the mixing of two magmas. He agrees that the dioritic rock is a hybrid in the newer sense of the term, but he thinks the facts are capable of explanation in terms of reaction between the magma of the Newry granodiorite (which envelops the contaminated rocks) and solid magnesian rocks.

F. Homma has described an abnormal lava from one of the Unzen volcanoes in Kyushu, Japan, which has insets of olivine and a plagioclase with reversed zoning (see p. 112), also some altered hornblende, biotite with reaction-rims, and grains of corroded quartz. Homma deduces from his microscopic study of the rock that an olivine-basalt magma was injected into a basin containing a hornblende-dacite magma after the latter had begun to crystallize, and that the abnormal lava is a hybrid. This interpretation may be correct, but it does not help us to understand how a magma with an excess of silica came into existence side by side with another magma deficient in silica.

E. S. Larsen and colleagues have been led, by a detailed study of the lavas of the San Juan region, Colorado, to the conviction that there must have been "a thorough mixing of very large masses of magma" in that region. Rhyolitic and basaltic lavas or andesitic lavas are associated in an intimate and erratic manner, basaltic horizons being interlayered with nearly all the rhyolites and rhyolitic horizons with some of the andesites. Thus acid and basic lavas must have been erupted in quick succession from the same vents or from vents very close together. Study of the larger feldspar crystals in these lavas has led to the conclusion that many of them did not grow in the environment in which they are now found; thus zoned plagioclase crystals with highly calcic cores are found in the rhyolitic lavas and others with dominantly sodic cores in the basaltic lavas. It is thought that some of these crystals may have settled down or floated up from one layer of magma to another, but most of them are believed to have reached their present environment in consequence of the mixing of two partly crystallized magmas. Nevertheless, reaction between liquid lava and solid lava or other solid rocks remains a possible alternative.

To summarize: it seems to the writer that evidence of the mixing of two completely or even partly fluid magmas is almost as unconvincing as evidence of unmixing was found to be. If silicate liquids are not immiscible then, of course, they must be capable of mixing, but before they can mix they must occupy the same space at the same time. It is known that certain volcanoes have emitted basalt at one time and rhyolite or trachyte at another, but not both at the same time. It is perhaps conceivable that under deep-seated conditions a basic intrusive mass might be succeeded by an acid one while the former was still hot, as Harker supposed in Skye and Phemister at Sudbury; under these conditions a limited amount of intermingling might indeed take place; but the writer is of opinion that most cases of the supposed intermingling of liquids are really cases of interaction between liquid and solid.

3. Differential Movement of Crystals and Mother-Liquor

A liquid rock-magma has no sharp freezing-point, but it has a long freezing interval, crystallization beginning perhaps in the neighbourhood of 900° C. and continuing down to 500° C. or thereabouts. Between these temperatures the magma is a "mush" of liquid and crystals. In the upper part of the temperature range there are few crystals in a large volume of liquid; in the lower part of the range there is a network of interlocking crystals with a little residual liquid in their interstices. It has been said with reason that "a mass of this composite nature will have some peculiar physical properties, unlike those either of a crystalline body or of a liquid" (A. Harker).

It is obvious that any agency by which the early-formed crystals can be segregated in one part of the magma-chamber will bring about a high degree of differentiation in the resulting rocks. Such an agency may be gravitation. The crystals of earliest formation in most rocks belong to comparatively dense minerals like olivine, pyroxene, or a lime-rich plagioclase; and if these crystals sink to the bottom of the magma-chamber they will give rise there to a body of rock which will differ greatly in composition from that formed at higher levels by the crystallization of the residual magma. A similar effect will be produced if the early crystals are lighter than the magma and tend to rise to the top; and the most pronounced difference will result if some crystals tend to rise and others to sink. N. L. Bowen (2) has actually demonstrated the sinking of olivine crystals, and in another instance the rising of tridymite crystals, in a few cubic centimetres of liquid in a crucible. W. E. Trommsdorff has pointed out that if there is differential movement of this nature in a body of crystallizing lava, it should be possible in favourable cases to detect it under the microscope by observing stream-lines in the ground mass surrounding the moving crystals. In *Fig*. 22 (A) the stream-lines indicated by the microliths in the groundmass show clearly that both the large plagioclase and the large augite crystals were sinking through the magma just before it congealed. In *Fig*. 22 (B) the upward movement of a large leucite crystal is indicated.

Other agents besides gravitation are capable of setting up differential movement of crystals and liquid. A. Harker (3) indicated the effect of crustal stress upon the spongy mixture of crystals and liquid; he claims that the crystalline network may yield to deformation and the interstitial liquid may be squeezed out of the meshes "as water is squeezed from a sponge." Since the early-formed crystals and the residual liquid must differ widely in composition, this process may indeed be one of the most effective agents in rock differentiation. N. L. Bowen (3) has given some consideration to the possibility of the "squeezing out" of residual magma, in connection with the genesis of alkaline rocks. Both Bowen and A. C. Waters have recorded a straining or filtering effect where magma containing crystals has been injected into narrow fissures. R. Balk has pictured a similar process, on a vastly larger scale, in the Adirondack Mts.

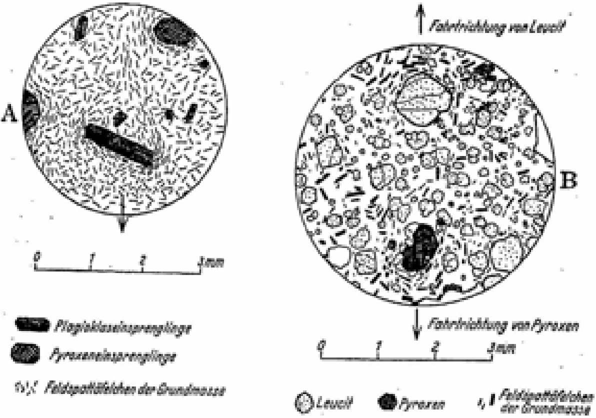

FIG. 22. Flow-lines in the ground mass of (A) a trachyte, (B) a leucite-basanite, demonstrating the settling of augite and plagioclase crystals (in A) and the simultaneous settling of augite and rising of leucite crystals (in B). (W. F. Trommsdorff.)

The escape of dissolved gases may be another cause of differential movement within the magma. R. A. Daly has pointed out that the formation of bubbles must lend buoyancy to the magma, and that local vesiculation would almost certainly set up convection currents. He describes this process as "two-phase convection." So long as the magma is wholly fluid, it is not clear that any marked chemical or mineralogical differentiation could arise in consequence of two-phase convection; but in a mush of crystals and liquid some effect is certainly to be expected. The crystals may be swept aside by the currents, just as logs are swept into the backwaters of a flooding river, with the result that some parts of the rock will be highly porphyritic and other parts not. In the case of magma that reaches the surface, one lava flow may be full of large feldspar crystals and another almost free from them.

The remarkable layered structure of the gabbro in the Skaergaard intrusion, East Greenland, is supposed to have been brought about by convection currents due to cooling, in a magma which contained a proportion of early-formed crystals. It is thought that differences of density among the crystals, and variation in the velocity of the currents, may have had a "winnowing" effect upon the crystals. (Wager and Deer.)

Another possibility that deserves consideration is that bubbles of gas may attach themselves to the growing crystals and float them upwards. This explanation presented itself to the writer (1) in connection with a lava-flow at Kijabe, Kenya, which is both vesicular and exceptionally rich in large insets of plagioclase. It is admitted that it might be difficult or impossible to distinguish between the effect of gas-flotation and that due to two-phase convection or simple gravity flotation.

There is still a third possibility to be taken into account. The vapour-pressure of the dissolved gases of the magma must be greatest when there is only a small residue of liquid left in a large mass of crystals. At this stage boiling may set in, and there may be an active expulsion of the residual liquid from the interstices of the crystal network by means of streams of gas bubbles rushing towards a vent in the roof of the magma-chamber. This is something different from two-phase convection, for it depends on the passage of expanding bubbles of gas through a multitude of narrow openings in a network of crystals, driving or blowing the residual liquid along with them. This process has been described as "gas streaming," and the writer (2) has used it to explain the occasional expulsion of trachytic or phonolitic lavas from volcanoes that normally yield only basalt, as in Mauritius and St. Helena.

We see that gravity, earth-pressure, convection currents, and escaping gases are all competent to bring about some degree of separation of the residual liquor from the crystalline portion of a body of partly solidified magma. Now this residual liquor, in the case of the more feldspathic magmas at least, is always rich in alkalis, and N. L. Bowen (4) has proved that many phonolites, trachytes and rhyolites actually have a composition which places them in the lowest-melting region of the system $NaAlSiO_4$—$KAlSiO_4$—SiO_2 (nepheline, kaliophilite, silica). Consequently Bowen makes the claim that "the process exerting the dominant control over the composition of liquids in these lavas has been fractional crystallization," that is, separation of the crystals from the residual liquor.

From the discussion of possibilities we pass to actual examples of crystal fractionation that have been demonstrated in the field.

Charles Darwin noticed the accumulation of plagioclase crystals towards the bottom of a scoriaceous lava-flow on James Island in the Galapagos group. He attributed the phenomenon to the action of gravity, recalling a similar observation made by L. Von Buch on Teneriffe; and he stressed the importance of the phenomenon as "throwing light on the separation of the trachytic and basaltic series of lavas." Another instance of the same kind is described by L. L. Fermor among the basalts of the Deccan, India. A thickness of over 1200 feet of these lavas was intersected in a borehole, and Fermor shows that most of the flows carry insets of labradorite from a quarter to half an inch in diameter. The insets are generally absent from the upper parts of the successive flows, present in moderate quantity in the middle and abundant in the lower parts. H. T. Stearns (1) has described a thick lava flow on the island of Oahu, Hawaii, which contains crystals of transparent feldspar half an inch across. These are so densely concentrated near the base of the flow as to form a nearly pure feldspar rock.

In all these cases we observe that crystals of plagioclase seem to have been denser than the parent basalt. But in the basalt of Kijabe, Kenya, to which reference has just been made, large crystals of plagioclase (labradorite) have accumulated at the top of the lava-flow. If the normal tendency of the early crystals of plagioclase in feldspathic basalts is to sink, then it follows that some other agency such as gas-flotation or two-phase convection must have operated at Kijabe. Among deep-seated rocks there is a good deal of field evidence which indicates that the normal tendency of plagioclase crystals in basic magmas is to rise. N. H. Winchell considered that the bodies of anorthosite which are developed in connection with the Duluth gabbro of Minnesota were formed by the upward accumulation of labradorite crystals from the gabbro magma. F. F. Grout (2) has explained the genesis of small bodies of anorthosite at Pigeon Point, Minnesota, in the same way.

Although the minerals of early crystallization are most commonly the denser ones, such as olivine and augite or hypersthene, there is one conspicuously light mineral that often crystallizes at an early stage, namely, leucite. W. E. Trommsdorff has been able to demonstrate under the microscope that the larger leucite crystals in a leucite-basanite of Vesuvius were actually rising through the lava at the same time that the augite crystals were sinking. (*Fig.* 22, B). It is likely that the remarkable rock that H. S. Washington named "italite," and which contains ninety per cent of leucite, is simply a portion of a lava-flow which has been specially enriched with leucite crystals by gravity flotation.

The sinking of augite crystals has often been recorded, for instance by Darwin in the Galapagos Islands and by King in Hawaii. J. P. Iddings described a sheet of diabase at Electric Peak, Yellowstone National Park, in which "a layer of the sheet four or five feet thick near the bottom contact is full of large porphyritical augites." The upper contact, it is interesting to note, carries small insets of feldspar. R. E. Fuller has described the gravitational accumulation of olivine during the advance of basalt flows in south-eastern Oregon, and claims that "the grains accumulated in a manner analogous to the deposition of sand at the delta of a river." H. T. Stearns (2) has given some remarkable illustrations of the settling of olivine and pyroxene in lavas of Samoa.

Perhaps the most convincing example that has yet been investigated of the sinking of heavy crystals through a magma, was described by F. Walker in the Shiant isles in the west

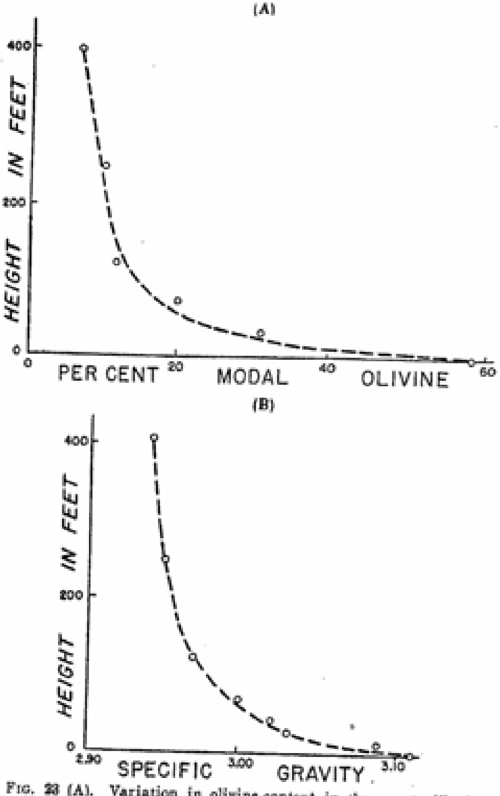

Fig. 23 (A). Variation in olivine-content in the great sill of the Shiant isles. (B). Variation in specific gravity in the same sill. (F. Walker.)

of Scotland. Walker remarks that the ideal condition for this type of differentiation would be one where the magma was intruded in a completely liquid form and no disturbance took place during cooling. Such a condition will seldom be realized, but the basic sill of Garbh Eilean in the Shiant isles shows a close approach to the ideal. Walker has demonstrated that there is a continuous increase in specific gravity, with a corresponding increase in the concentration of olivine, from top to bottom of this sill. The data are presented in a convincing manner by the two graphs in *Fig.* 23.

The Palisade sill in New Jersey presents evidence of the settling of both olivine and pyroxene. In this case almost the whole of the olivine is concentrated in a single level — the Olivine Ledge—near the base of the sill. (See *Frontispiece.*) The data contained in the following table were furnished by F. Walker.

THE COMPOSITION OF THE PALISADE SILL

Height above base.	*Olivine.*	*Pyroxene.*	*All Dark Minerals Together.*	*Plagioclase.*	*Anorthite in Plagioclase.*	
997	2	50	58	42	57	A little quartz is generally present above the olivine ledge.
980	1	44	49	50		
950		37	41	52		
900		32	34	60		
700		31	34	61	47	
650		29	30	66		
550		33	35	61		
320		36	41	55	50	
230		43	44	54		
170		44	47	52		
120		45	47	52		
90		53	54	45	57	
80		53	54	45		
73	1	52	56	44		
71	17	43	63	37		
58	22	37	65	35	63	The olivine ledge.
50	21	44	68	32	65	
49	15	45	63	37		Quartz is hardly ever present below the olivine ledge.
40	4	51	58	42		
30	9	49	60	40		
15	7	51	61	39	60	
6	2	43	47	52		
3	1	43	45	55		
1	2	46	53	48		

However the details of the process may vary, one seems compelled to admit that differential movement of crystals and magma has been one of the most potent causes of differentiation in many eruptive masses.

4. VARYING CONCENTRATION OF THE FUGITIVE CONSTITUENTS

If a body of magma contained in a closed reservoir cools and eventually crystallizes from the margin inwards, then as crystallization progresses the concentration of the fugitive constituents must show a continuous increase in the still fluid central portion. In consequence of this change in concentration it is likely that the central rock, after crystallization is complete, will show an appreciable difference, not only in texture but also in composition, from the marginal facies. A magma of basaltic composition, for instance, might give rise to a marginal facies containing pyrogenetic minerals such as augite and olivine, whereas in the central facies hydatogenetic minerals such as hornblende and biotite would be favoured. This type of differentiation would affect phases more than components, at least in the earlier stages; but we might expect to find some concentration of the more soluble components of the magma, especially the alkalies, in the central facies. The process pictured here is in fact an earlier stage of the process which ultimately gives rise to pegmatite veins.

Mount Girnar, in the state of Kathiawar, India, is one of those remarkable intrusive bodies built up of concentric rings of different kinds of rock, of which many examples have been described in the last thirty years. Beneath a cover of basalt which has been partly removed by erosion, there is a central mass of dioritic or monzonitic character, composed of andesine, orthoclase, green pyroxene, hornblende and biotite in various proportions. Round about the central mass is a body of olivine-gabbro containing labradorite, olivine and augite, with only minor amounts of biotite, hypersthene and orthoclase. Beyond the gabbro there is a nearly complete ring of granophyre which is younger than the other rocks and cuts the basalt cover.

According to the interpretation put forward by K. K. Mathur, V. S. Dubey and N. L. Sharma, the olivine-gabbro crystallized first and the residual magma, enriched with fugitive constituents, was concentrated towards the bottom and centre of the reservoir. Here the reaction principle came into operation and olivine was entirely replaced, and augite largely so, by hornblende and biotite; a more sodic plagioclase was formed, and the proportion of orthoclase was greater, giving diorite and monzonite instead of gabbro. The facts described by Mathur, Dubey and Sharma are consistent with the view that variation in the concentration of the fugitive constituents (water and alkalis) was the main cause of differentiation among the rocks of Mount Girnar.

Another rock complex to which this explanation may be applicable is Shonkin Sag, in Montana. This is a laccolith composed of several kinds of rock with a rudely concentric arrangement. The marginal facies was described by W. H. Weed and L. V. Pirsson as leucite-basalt porphyry; it is a compact porphyritic rock which must have cooled quite rapidly. Within this chilled skin the bulk of the magma crystallized as a medium-grained shonkinite, a type of syenite relatively rich in dark minerals; and within that again there is a comparatively small core of more normal syenite. The shonkinite contains some 50 per cent of augite, olivine, biotite and apatite in a matrix of potash-feldspar. The syenite holds about 20 per cent of augite and biotite, with 40 of potash-feldspar and 30 of zeolites.

The original description of this remarkable complex by Weed and Pirsson has been supplemented quite recently by J. D. Barksdale; and even more thoroughly by C. S. Hurlbut, who explains the observed differentiation in terms of sinking of the heavy minerals combined with flotation of the early formed leucite crystals. There can be no doubt that crystal settling is responsible for the variation in the concentration of the heavy minerals, but it remains to find an explanation of the unusual degree of zeolitization in the rocks at the centre of the complex. It is clear that the magma, which formed a lens not more than 250 feet thick in the middle, was chilled very rapidly at first. The chilled phase at the roof and floor of the laccolith is 12 to 15 feet thick. As cooling and crystallization proceeded from the margin inwards, there would be a tendency for the residual magma, enriched with fugitive constituents which were prevented by the chilled phase from escaping outwards, to be concentrated in the middle of the laccolith. That there was such a concentration of fugitive constituents is indicated, not only by the abundance of zeolites, but also by the increasingly coarse grain of the later syenitic rocks, and the change in the character of the dark minerals from augite and olivine in the chilled margin to augite and biotite in the shonkinite, to a more or less sodic pyroxene in the syenite, and to ægirine and soda-amphiboles in the pegmatitic phase. These reactions may have begun in the magma, but in part they are post-magmatic.

It must now be abundantly clear that the variations of mineralogical and chemical composition within bodies of eruptive rock are due to many causes, some of which we know and others we can only guess at. Certainly there is no single process of "magmatic differentiation," and the sooner that pretentious and misleading term is dropped the better for petrology. In the foregoing pages we have described a few of the simpler cases of rock differentiation, in which a single process seems to have predominated; but even in these cases it is likely that more than one process was involved. Thus the Lugar sill (*Fig.* 20) clearly shows evidence of the settling of olivine crystals, especially in its middle member which consists of picrite passing down into peridotite; but the complexity of this intrusion is such that G. W. Tyrrell was driven to assume three successive injections of magma varying in composition from teschenitic to picritic and peridotitic. In other words, the greater part of the problem was left unsolved.

J. S. Flett has made a series of studies of basic sills in the Carboniferous of Scotland. In some he finds evidence of the settling of olivine crystals, but he finds it impossible to explain all the observed differentiation within the sills by crystal settling alone. "Other evidence indicates that the liquid part of the magma was not homogeneous throughout, but had undergone differentiation that cannot be ascribed to crystallization and precipitation."

In a study of the Braefoot sill, in Fife, which is another member of the same group of intrusions, R. Campbell is compelled to assume not only crystal settling but also assimilation of quartzite, filter-press action, and pneumatolytic and hydrothermal processes in order to construct a satisfactory explanation of the observed differentiation.

G. M. Schwartz and A. E. Sandberg have lately described three sills of diabase with associated granite or micropegmatite at Duluth, Minnesota. The authors discuss the alternative hypotheses of successive intrusions, of metamorphism of the roof, of hydrothermal alteration, of syntexis, and of gravitational differentiation, and find it impossible to explain the phenomena by any one of these factors alone.

Extrusive Rocks.

If it is so difficult to explain the variation observed within an intrusive body, where all the facies and their mutual contacts are exposed to view, how can one hope to understand the relation of one extrusive rock to another? We know that some volcanoes have yielded only one kind of lava throughout their history, while others have emitted different kinds of lava at different periods. The lavas of Etna seem to have varied very little during centuries; sometimes they are more andesitic, sometimes more basaltic, but the difference is slight and the average Etnean lava is andesitic in composition (Washington, Aurousseau and Keyes). Vesuvius on the other hand has changed its character considerably in the last two thousand years. The oldest known product of Monte Somma (the ancestor of the modern Vesuvius) is a trachytic pumice; leucite tephrite appeared later and still continues to be emitted, but A. Rittmann has shown that the composition of the lavas has changed steadily, silica, alumina and alkalis decreasing, while lime, magnesia and iron increase with the passage of time. H. J. Johnston-Lavis pointed out that on the island of Vulcano, for two thousand years, eruptions of basalt and obsidian have been going on at vents only a mile apart. H. Williams showed that rhyolite and basalt have been emitted alternately from the Newberry volcano in Oregon. In the small area of the Siebengebirge on the Rhine, half a dozen volcanoes so close together that they almost touch have yielded lavas varying from typical trachyte (Drachenfels) to andesite (Wolkenberg) and olivine-basalt (Löwenburg), as well as dyke-rocks containing feldspathoids. In the Rift Valley region of Kenya the lavas vary from obsidian holding 32 per cent of excess silica to phonolite with 30 per cent of nepheline, basalt with 20 per cent of olivine, and "augitite" with 50 per cent of augite. The most abundant lava of the Hawaiian volcanoes is a basalt with little olivine; but W. Cross and H. S. Washington have recorded the following varieties in addition: trachyte, trachyandesite, oligoclase-andesite, andesine-andesite, olivine-rich basalt, picritic basalt, limburgite and melilite-nepheline basalt.

A great deal of ingenuity has been expended in guessing at the meaning of these remarkable variations in the composition of the lavas emitted from a common vent, or within a limited region. It is a wise conclusion that until we know more about the causes of differentiation in deep-seated rocks we cannot begin to understand the genesis of lava-flows.

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A plurality of suffrages is no guarantee of truth where it is at all difficult of discovery.—DESCARTES.

CHAPTER X

LATE-MAGMATIC AND POST-MAGMATIC REACTIONS

IN the last three chapters we have discussed the crystallization of silicate liquids with little regard to the presence of dissolved water. But we know that most rock-magma—probably all rock-magma—holds some water in solution. It is important to inquire what becomes of this water as the magma crystallizes. Some of it is used up in forming hydroxyl groups in hornblende or mica, and consequently remains in the rock; but a large proportion of the dissolved water escapes either to the atmosphere, in the case of surface lava, or into the wall rocks, in the case of deep magma. This we know from the study of volcanic gases and contact metamorphism. From the point of view of the student of ore-deposits, the escape of aqueous solutions is the most important incident in the cooling history of the rock.

It is surprising to find that H. C. Sorby, as long ago as 1858, had formed a clear picture of the crystallization of a hydrous magma and of the consequences that must follow if the escape of water is prevented.

> "In my opinion, the water associated with thoroughly melted igneous rocks at great depths does not dissolve the rock, but the rock dissolves the water. . . . If the fused rock passed by cooling into anhydrous crystalline compounds, the water would necessarily be set free; and if the pressure was so great that it could not escape as vapour, an intimate mixture of partially melted rock and liquid water would be the result. It is difficult to form any very definite opinion as to the actual amount of this water, and to decide whether or not it exercised an important influence over the crystalline processes that took place during the consolidation of such rocks as granite. . . . I think the amount, though limited, must nevertheless have been *considerable*, and that its presence will serve to account for the connexion between granite and quartz veins, and the intimate relation of both to the metamorphic rocks, and explain many peculiarities in the arrangement of the minerals in the cavities in granite or in the solid rock, even if it was not the effective cause of their elimination and crystallization."

In this chapter we can do little more than try to fill in the details of Sorby's picture.

It was long assumed that the solubility of water in rock-magma is unlimited, and that during crystallization the proportion of dissolved water increases steadily until the liquid acquires the composition of a hot aqueous solution of silicates. But R. W. Goranson's investigation of the solubility of water in molten granite has demonstrated that granite magma, under such conditions of pressure and temperature as concern a petrologist, cannot retain in solution more than some nine per cent of water. Whether such a concentration is ever reached in nature will depend on the resistance offered by the wall rocks to the escape of gases. If we try to picture the course of crystallization of a magma having a small initial content of water, we may begin by assuming that the wall rocks are impermeable to gases.

The earliest crystals formed from rock-magma are frequently anhydrous species such as olivine, pyroxene, and plagioclase; in this case the concentration of water in the liquid must increase as crystallization advances. Sooner or later a stage is reached at which olivine and pyroxene are no longer stable; hydroxyl-bearing silicates such as hornblende and mica are now generated, partly by direct crystallization from the magma and partly by transformation of the earlier, anhydrous species. If a sufficient quantity of magmatic water is used up in this way, then the limiting concentration determined by the prevailing temperature and pressure may never be reached, and the hydrous magma will simply decrease in quantity until it is all used up, exactly as in the crystallization of an anhydrous system. But if the limiting concentration of water is reached, then further crystallization of the magma will liberate the excess of water as a second fluid phase which, if it cannot escape, must of necessity form an emulsion with the magma. Vesicular lava is just such an emulsion, in which the second fluid phase is a gas. In the case of deep magma the second fluid will not be a gas, but in all probability a liquid below its critical temperature. It will not be pure water, but a saturated solution of the strong bases, soda, potash, and lime, combined with silica, alumina, and other weakly acidic oxides such as titania, zirconia; and with a very small proportion of the strongly acid chloride, fluoride, sulphate, or phosphate radicles. In the case of a gabbroic magma, potash will be reduced and magnesia may be present.

It is often assumed that any fluid liberated from rock magma must be in the supercritical state; but this does not appear to be necessary, and many authorities oppose the assumption. Although weak solutions of salts such as KCl exhibit critical phenomena, it has been established by experiment that a saturated solution of salt does not do so, the vapour pressure rising to a maximum and then declining again (M. Benedict).

Whether an emulsion of magma and aqueous solution was actually formed, in any particular instance, we do not know and do not need to know, because it would cause no interruption in the crystallization of the system and no change in the character of the minerals crystallizing out. Nevertheless the abundance of aqueous inclusions in minerals of late crystallization, especially quartz, seems to indicate the presence of liquid water during their growth. From such an emulsion, feldspars and hydroxyl-bearing silicates will continue to crystallize until the magma is used up and only the aqueous-alkaline solution remains in the pores and shrinkage-cracks of the rock. The amount of this aqueous-alkaline residue will depend on the opportunities which were presented for the escape of water from the system by diffusion or boiling. The assumption made at the beginning of this discussion, that the rock walls are impermeable to gases, is probably extreme. If some water is able to escape by diffusion, during the cooling of the system, then the aqueous residue may be insignificant in amount, and if much water escapes, there may be no aqueous residue at all; yet there is an abundance of field evidence which suggests that the aqueous residue, especially from granitic magma, is by no means insignificant. It is to this residue that we credit most of the metasomatism and at least some of the ore deposition that we observe in the contact zones of large intrusions. To it we also concede a large share in the generation of the aplites and pegmatites, as well as the vein-quartz, which fill contraction fissures in the eruptive rock and its walls. These vein-filling aggregates could not form until there were fissures for them to form in, and the opening of fissures cannot begin until the system is too rigid to flow as a whole; that is, until the crystallization of the magma is far advanced. Day, Sosman, and Hostetter have shown that quartz undergoes a rapid contraction, amounting to 2·4 per cent of its volume, between 575° and 500°; this is likely to be the period of most rapid contraction in granitic rocks.

At the stage when contraction fissures begin to form, crystallization must clearly be so far advanced that the network of crystals is strong enough to support the gravitational and orogenic pressure on the system. The liquid contained in the pores and interstices is accordingly relieved of this pressure, and for the first time it may become possible for a gas phase to develop in the closed system. By its expansion this gas will tend to drive the residual liquid into the gradually opening fissures in the rock. N. L. Bowen has discussed the further history of the boiling liquid in the fissures, especially in relation to contact metasomatism and ore-deposition. For the moment we shall confine our attention to that portion of the residual alkaline solution which remains in or travels through the pores of the rock and continues to react with the surrounding crystals. It may be supposed that this small residue of aqueous solution is too scanty and too dilute to make by itself any significant contribution to the bulk of the rock; yet it may perform an important function by promoting and facilitating reaction between the minerals already present. It is likely, too, that as water vapour escapes through the walls and roof of the intrusion, as it must eventually do in consequence of the greatly increased vapour pressure, it will be replaced continuously by the boiling and upwelling of more of the pore-solution from deeper levels; thus the total amount of reaction and deposition that is accomplished in the upper levels of a large intrusive body may in the end be very great.

The reactions thus induced are of many kinds, but the general tendency is to replace anhydrous species by hydrous ones and moderately hydrous species by still more hydrous ones. All such reactions are in a literal sense *hydrothermal*, but a useful distinction has been made between reactions which tion of the magma itself." This definition seems to have the same meaning as Becke's words, quoted above, and the term deuteric has been used by a number of petrologists in this sense. It is suggested, therefore, that deuteric be retained and applied to those high-temperature hydrothermal reactions which take place at the close of the magmatic stage, under the influence of residual aqueous solutions derived from the magma itself. If the reactions appear to have been brought about by fluids from an external source, the term deuteric should not be used. For example, the production of poikilitic hornblende in the norite at the heart of the Cortlandt complex (see on, p. 167) seems to have been effected by solutions derived from the norite itself, so this has been described as a deuteric reaction. But the alteration of the Salem gabbro by the Quincy granite and solutions derived from it, in Essex Co., Massachusetts, is a high-temperature hydrothermal reaction which is not deuteric.

Field evidence only can show us whether the transforming solutions, in such a case, came from within or without the body of rock that was transformed. If they came from within, is it certain that they were derived from the crystallizing magma? Might not the effect be due to ground water, heated and set in circulation by the intrusion? This is no new question. We have met it already in connection with volcanic gases, and we shall meet it again in connection with pegmatites and ore-deposits. Every time it is asked, whether in relation to volcanic gases, deuteric reactions, pegmatites, or ore-deposits, we get two irreconcilable answers. According to one view, the main source of mineralizing solutions and volcanic gases is rock magma in course of crystallization; according to the other, all mineral-forming solutions and volcanic gases are derived from ground water, sea water, or the atmosphere.

This difference of opinion makes it only too clear that there is no rule by which the effects produced by magmatic solutions can be distinguished from those due to heated ground water. The same minerals will grow under the same physical conditions, if the same bases and acids are present, no matter where or how the solutions had their origin. It would be easy to show by simple enumeration that the magmatic hypothesis commands most support at the present day; but it is well to remind ourselves that "a plurality of suffrages is no guarantee of truth where it is at all difficult of discovery."

May not both hypotheses be true? Perhaps some volcanic steam was originally ground water or sea water; perhaps some ore-deposits were formed by lateral secretion, and some pegmatites too. The same admission may be made for some post-magmatic reactions. But if we agree that rock-magma contains dissolved water, some of which is set free during crystallization, then we must also agree to the proposition that as crystallization proceeds beneath an impermeable cover, the conditions will become increasingly favourable to post-magmatic reactions of the nature of hydrolysis, base exchange, and the concentration of the scarcer chemical elements in the later crystallizations. Deuteric action is *expectable* under deep-seated conditions.

In the remainder of this chapter we shall describe some phenomena which are widely believed to be deuteric

THE TOURMALINE-TOPAZ-CASSITERITE TYPE OF MINERALIZATION

The hypothesis that an already solidified eruptive rock may be locally transformed and recrystallized by the action of its own residual solutions was clearly expressed by H. C. Sorby in 1858, and was also developed in the writings of A. Daubrée. In a study of the genesis of tin deposits, begun in 1841, Daubrée pointed out that these deposits are distinguished from other ore-bodies by a characteristic suite of associated minerals, namely, quartz, topaz, tourmaline, lithia-mica, apatite, and wolframite. It is to be observed that these minerals contain the elements fluorine, boron, phosphorus, lithium and tungsten, between which there is no chemical relationship whatever. They differ as widely as possible in valence, in atomic weight, and in ionic radius, and they do not form isomorphous compounds. Is it conceivable that ground water, seeping through rocks of all kinds, could make such a strange selection of unrelated elements?

Daubrée remarked that the familiar replacement of feldspar by tourmaline seems to demand the presence of an agent which was able to decompose feldspar and introduce boron at the same time; and that pseudomorphs of cassiterite and quartz after orthoclase, such as are found near St. Agnes, Cornwall, require an agent which can decompose feldspar and introduce tin. Daubrée found such an agent in fluorine. He pointed out that tin fluoride is volatile at a moderate temperature (actually 705°), and that the fluorides of boron (−101°), silicon (−160°), phosphorus (−75°) and tungsten (+19°) all have remarkably low boiling points. When these fluorides come into contact with water they are decomposed, liberating hydrofluoric acid which readily attacks feldspar and other silicates.

Daubrée discussed the almost invariable association of tin deposits with granite and granite-porphyry, and he observed that the fissures in which cassiterite, topaz and tourmaline are deposited are typically simple contraction fissures, not fractures of tectonic origin. He accordingly advanced the theory that tin and the elements that accompany it were transported in the form of fluorides from "the depths where one finds the general reservoir of metals." Geology knows no such reservoir of metals except the deep-seated rocks and their magmas.

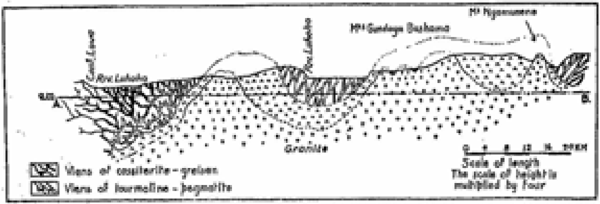

FIG. 24. Section through the Bushema batholith, Belgian Congo. The central biotite-granite has a marginal facies of muscovite-granite from which issue veins of tourmaline-pegmatite, passing out into cassiterite-greisen. (F. Blaise and N. Boutakoff, *Ann. soc. géol. de Belgique*, 57, 1937, p. c97.)

Some ten years after Daubrée had expressed these ideas. R. Bunsen introduced the adjective "pneumatolytic" to describe a group of reactions caused by acid gases escaping from active lava flows. These reactions are of an entirely different character from those visualized by Daubrée, but in 1890 W. C. Brögger took over Bunsen's adjective and applied it to the group of minerals, typically developed in pegmatites, which contain fluorine, chlorine, boron, phosphorus, and other so-called "mineralizing agents." In consequence of this perversion of meaning, the view became firmly established that minerals containing these unusual elements must have been formed by the agency of magmatic gases; and to this day geologists who encounter rocks or mineral deposits containing tourmaline or topaz too often describe them as "pneumatolytic" products. The reactions that Bunsen described were truly pneumatolytic, because they took place under atmospheric pressure, but there is no reason to suppose that the fluids which generated tourmaline and topaz in plutonic rocks were gaseous. It is important to realize that the *assumption of gas as the medium of transport was made to meet the special case of cassiterite*. But it is now known that tin dioxide, like silica, can be transported as a sodium salt, or even in colloidal form in water, so all necessity for a high temperature and gaseous transport has disappeared, both for cassiterite and for the associated minerals. In view of this, it would be an advantage if the term "pneumatolytic" were to disappear too. The condition under which the topaz-tourmaline-cassiterite group of minerals was formed is preferably described as high-temperature hydrothermal; either deuteric or not.

F. G. Smith finds that tourmaline is unstable in strongly acid solutions, but stable in those of low alkalinity; hence he concludes that "if cassiterite and tourmaline are found in nature to have been deposited from the same solutions at the same time, then the solutions must have been alkaline, not acid."

An important study of tourmalinization in the Dartmoor granite and its surroundings was made by A. Brammall and H. F. Harwood, who describe the following modes of occurrence of tourmaline:—

(1) Little euhedral crystals enclosed in biotite. These are thought to be primary magmatic products;
(2) Large irregular crystals replacing plagioclase or biotite;
(3) Tongues of tourmaline selectively replacing the threads and patches of albite in microperthite;
(4) Euhedral crystals in geodes, along with feldspar, quartz, fluorite and apatite;
(5) Tourmaline and quartz together replacing euhedral orthoclase crystals;
(6) Radiating groups of needles ("tourmaline suns") in nodules in the granite;
(7) Quartz-tourmaline veins;
(8) Tourmalinized country rocks,

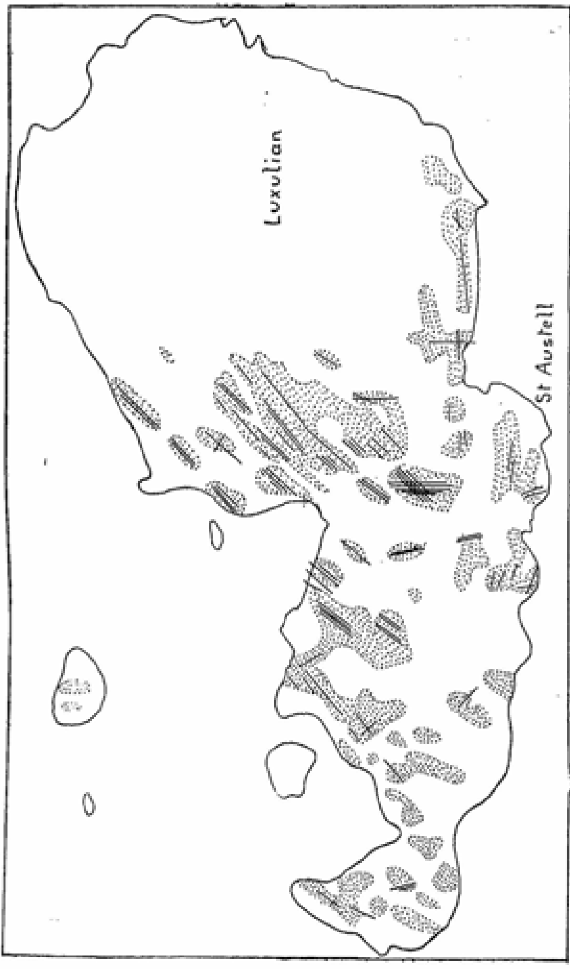

FIG. 25. Geological map of the St. Austell granite mass, Cornwall, showing the distribution of china-clay areas (dotted) and tin lodes (black lines). (After D. A. MacAlister.)

Brammall and Harwood summarize the cooling history of the granite as follows: "The magma fraction represented by the earliest and most basic type of Dartmoor granite was essentially free from those volatile constituents which promote the genesis of tourmaline. The magma fractions which yielded the later and progressively more acid types contained these volatile constituents in increasing concentration. In the products of the latest stage of intrusion, the concentration was such that the intrusive bodies were more or less severely pneumatolyzed (*sic!*) before crystallization was complete. During the closing phase, the granite of east Dartmoor was affected by severe tensions. Numerous fissures opened and afforded passageway for fluids and gases from depth. These cracks and fissures were infilled with schorlaceous veinstuff, quartz-schorl rock, vein quartz, and ore-bodies carrying specular iron or cassiterite. The closing phase was marked by local tourmalinization of granite in place."

In a later paper Brammall and Harwood discussed the temperature range in the formation of tourmaline. In the granite, tourmaline is associated with high-quartz and some of it is " demonstrably contemporary in origin with quartz in the normal groundmass of the biotite granite." In the geodes it is associated with low-quartz, but the lower limit of temperature of formation is not known. The authors record that little idiomorphic crystals of tourmaline have been observed in the Skiddaw slates. It may be added that P. D. Krynine has described overgrowths of new tourmaline upon detrital grains of the same mineral, in Paleozoic sandstone in Pennsylvania; so it appears that tourmaline may continue its growth right down to atmospheric temperature.

E. O. Teuscher has discussed post-magmatic reactions in the granites of the Erzebirge, in Saxony. These acid granites have a well developed pegmatitic end-phase, with albite-, topaz-, and lithionite-pegmatites. The Eibenstock granite contains many horizontal sheets of aplite, some containing topaz, others tourmaline. Even andalusite, cordierite, and dumortierite are occasionally present in these rocks. There is a continuous passage from albitized granite through tourmalinized granite into topaz-bearing aplite-granite. Teuscher writes:

"It can be said that the aplites are no offshoots of a distant magma, but that they stand in the closest relation to the main granite of the neighbourhood; they can best be regarded as a residual liquid from the crystallizing magma."

J. G. W. Dunn described topaz metasomatism and tourmaline metasomatism of granite-porphyry dikes at Mount Bischoff, Tasmania. In the earliest stage feldspar was replaced by topaz; occasionally quartz was replaced too, leaving a topaz rock in which the only other constituents were some radial groups of blue tourmaline and a little cassiterite. Tourmalinization was on the whole a later process; where it was most intense, an entire body of rock might be replaced by radiating groups of tourmaline needles with a little cassiterite. In a further stage of metasomatism, both topaz and tourmaline, as well as feldspar, might be replaced by cassiterite. Dunn considered the reactions to be pneumatolytic and wrote that "The direct association of the deposits with the porphyries indicates that the mineralization period represents the end phase of magmatic intrusion and differentiation." In other words, the granite-porphyry dikes were altered by their own residual fluids; there is no reason to believe that these fluids were gaseous rather than liquid.

The evidence that we have discussed gives strong support to the opinion that the generation of tourmaline and topaz, as well as minerals containing lithium, tin, and tungsten, must be ascribed to fluids which escaped from deep-seated magma in the final stage of crystallization. It does not seem possible to explain this remarkable association of unusual elements by the agency of ground water, more especially as the special minerals we have considered are almost entirely restricted to the most siliceous types of granite and granite-porphyry, and have even been found in rhyolite. It would perhaps be possible to imagine a process by which these elements, and these alone, were leached by ground water out of a completely solidified body of granite, and deposited about its margins; but there is no reason to think that ground water possesses such remarkable powers of selection. Extensive tourmaline and topaz metasomatism seem to demonstrate convincingly the reality of the deuteric stage of crystallization in acid eruptive rocks.

THE PYROXENE-AMPHIBOLE-EPIDOTE TRANSFORMATION

Another well-established process, which may be either late-magmatic or post-magmatic, is the replacement of pyroxene by hornblende. In an early stage of reaction this process yields only narrow mantles of hornblende about the pyroxene crystals, but in a more advanced stage large crystals of hornblende are generated which may be filled with little, more or less corroded grains of pyroxene, olivine, and other minerals. For this texture, which is beautifully developed in some of the hornblendic rocks of the Cortlandt complex in New York State, G. H. Williams introduced the adjective "poicilitic" (usually spelt poikilitic), meaning "mottled." In the rock to which Williams gave the name "cortlandtite," the hornblende forms allotriomorphic crystals several centimetres in diameter, which enclose many little, rounded grains of olivine, augite and hypersthene.

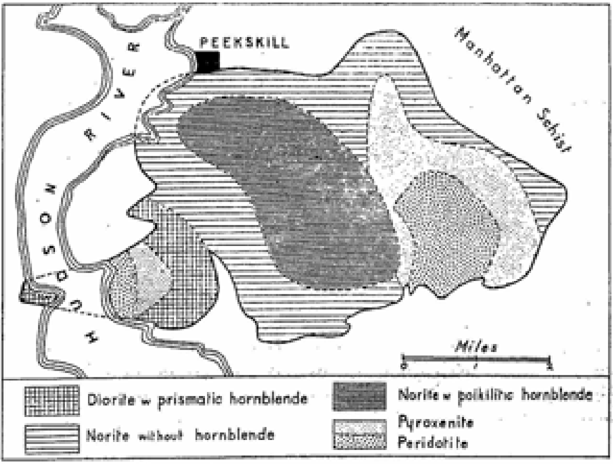

FIG. 26.—Geological map of the Cortlandt complex, New York State.

The Cortlandt complex, to which these rocks belong, is a funnel-shaped pluton with an area of some 25 square miles. At the east and west ends of the roughly elliptical outcrop there are bodies of pyroxenite and peridotite, but the central mass is norite with both ortho- and clino-pyroxene. The rocks are banded, coarse-grained bands alternating with finer-grained, and the bands dip inwards, towards the centre of the funnel. About the margin of the funnel the norite has developed a hornblendic facies (diorite in the sense of Zirkel) in which the hornblende is prismatic and apparently primary. Further inwards the norite is free from hornblende for a distance of half to one mile; then hornblende appears again and becomes increasingly abundant towards the centre of the complex. But whereas the hornblende in the marginal region is prismatic and generally free from inclusions, that in the central region forms large allotriomorphic plates and skeleton crystals which are full of corroded grains of plagioclase, augite and hypersthene (S. J. S., 2).

Since poikilitic texture was first described in rocks of the Cortlandt complex, we may digress for a moment in order to discuss the significance of this texture. When two substances crystallize side by side from a solution, a large crystal of one may accidentally enclose some smaller crystals of the other, especially if crystallization takes place rapidly. But in a slowly crystallizing magma, the chance that a mineral of later crystallization may enclose a large number of grains of the earlier minerals seems to be small. Crystal settling is a reality; and in the present instance the early-formed crystals of hypersthene, augite (and sometimes olivine) must have been considerably denser than the partly feldspathic magma in which they grew. It is significant, too, that to a large extent the enclosed grains are not idiomorphic crystals but corroded relics. Can we believe that crystals of density about 3·4 remained suspended in a moderately feldspathic magma long enough for crystals of hornblende, many times their own size, to grow from and around them? Evidence to the contrary is given by certain norites which have been subjected to crushing. In these rocks the grains of plagioclase and pyroxene show wavy extinction and partial granulation, but the hornblende which surrounds them is not affected in this way, hence it must have been formed after the rock was completely solid. The writer concludes that the poikilitic hornblende in the norites, pyroxenites, and peridotites of the Cortlandt complex is mainly deuteric, and was formed by transformation of the original minerals of these rocks under the influence of ascending solutions of magmatic origin. In rising through the centre of the funnel, the solutions followed especially the more coarse-grained bands in the norite, but some hornblendization also took place along irregular cracks.

It is a legitimate conclusion that a well-developed poikilitic texture, in any rock that has not undergone metamorphism by external forces, is good evidence of post-magmatic crystallization. The very same texture is characteristic of minerals such as andalusite, cordierite, and garnet which undoubtedly grew in a solid matrix in metamorphic rocks. It is also characteristically developed in crystals of calcite and gypsum which have grown in loose sand.

The Caribou stock in the Front Range of Colorado resembles the Cortlandt complex in some respects. It is mainly composed of monzonite, but it includes bodies of pyroxenite and gabbro which preceded the monzonite in order of intrusion. All the main varieties of rock contain augite, generally mantled by hornblende. Some of this hornblende may be late magmatic but some is certainly post-magmatic because veins of hornblende follow planes of fracture in the rock. In a facies described as porphyritic hornblende-pyroxenite, large crystals of hornblende lie in a fine-grained matrix of pyroxene and biotite. The large crystals are poikilitic and enclose grains of augite, biotite, apatite, and magnetite. Ward Smith reported that "One can see all stages of the development of large hornblende crystals from an initial stage in which hornblende is merely a spongiform, optically continuous network that extends through a matrix of pyroxene and biotite, through stages in which the hornblende is more and more predominant, to completely grown hornblende crystals that are solid except for inclusions of magnetite and apatite. This gradational series is interpreted as evidence that the large crystals developed metasomatically in solid pyroxenite."

The Sooke gabbro, in Vancouver island, shows similar phenomena of hornblendization. After the consolidation of the gabbro, there was a period of faulting accompanied by the formation of shear zones and joints. Through these fissures there came hot solutions that altered the rock with which they came in contact into hornblendite. The veins of hornblendite are mostly just a few inches wide, but some are 50 to 250 feet wide, grading into unaltered gabbro. The replacement of the wall-rock by hornblende can be observed in all its stages (H. C. Cooke).

B. K. N. Wyllie and A. Scott described a rock at Garabal Hill, Scotland, which is composed almost entirely of quarter-inch crystals of glossy black hornblende. Many of them have cores of diallage, and even those crystals which have no such cores still show relics of the "schiller structure" of diallage, indicating that the hornblende grew at the expense of pyroxene crystals. The authors express doubt whether a mere equilibrium change during crystallization of the magma could account for the formation of a strongly coloured hornblende (holding some 25% of ferrous and ferric oxides) from a practically colourless pyroxene. They suggest that the transformation was effected by "heated vapours" from a younger diorite magma. It is difficult to believe in the transport of large amounts of iron oxide by vapours (unless halogens were abundantly present, and of this there is no evidence), so it seems more likely that the transformation was effected by hot residual solutions from the magma that generated both the original pyroxenite and the dioritic and other rocks of the complex. This is the opinion expressed by S. R. Nockolds, who has re-examined the region.

TRANSFORMATIONS AFFECTING FELDSPAR

The commonest example of this nature is *albitisation*, which may effect either plagioclase or orthoclase. An enrichment of plagioclase with albite may be brought about (1) by simple addition, in the post-magmatic period, of layers of albite or anorthoclase to an existing plagioclase; (2) by substitution of Na-Si for Ca-Al atoms; (3) by transformation of anorthite into zoisite, leaving the feldspar relatively enriched with albite.

Bailey and Grabham described the albitization of plagioclase in a number of Scottish diabases and basalts. Plagioclase containing as much as 80% of anorthite has been replaced by pure albite, and since the crystals do not contain visible cavities (indicating removal of anorthite), there must have been a large introduction of soda into the crystal lattice. Silica must also be introduced, as the equation shows:—

$$CaAl_2Si_2O_8 + Na_2CO_3 + 4SiO_2 = 2NaAlSi_3O_8 + CaCO_3.$$

Bailey and Grabham picture the process as one of self-digestion or autolysis, claiming that "the soda introduced into the altered basic feldspars probably belongs to the lava itself"; they suggest that "the lava at this stage was stewing in a concentrated solution of sodium carbonate." P. Eskola and colleagues were able to reproduce in the laboratory the conditions described by Bailey and Grabham. Powdered bytownite or anorthite was heated in a steel bomb along with sodium bicarbonate, silica and water, at various temperatures between 230° and 550°. Under these conditions the lime-rich feldspar was converted into albite or oligoclase, the most complete albitization being realized between 260° and 330°.

But the demonstration that plagioclase can be albitized by hot soda solutions in the laboratory tells us nothing about the source of these solutions in nature. G. Beskow maintains that the soda necessary for the extensive albitization observed in pillow-lavas (spilites) has come from heated sea water. On the other hand many rocks have been albitized which are not known to have been exposed to sea water, and for these rocks Bailey and Grabham's explanation seems more acceptable. If the reaction was indeed due to sea water, one wonders why there was no production of scapolite (see on, page 175).

A careful study of the spilite problem, with a review of the abundant literature, has been made by J. Gilluly (1). The same authority (2) has described the production of albite-granite by the silicification and albitization of quartz-diorite, in Oregon. The diorite is composed of andesine and hornblende, with a little orthoclase and quartz; it passes gradually into a rock in which quartz is graphically intergrown with albite, the latter containing cores of andesine which are riddled with tongues and threads of albite. The hornblende is simultaneously transformed into deep blue hastingsite. The passage from diorite to albite-granite involves an addition of more than 20% of silica, an increase of soda from 2·7 to 4·5%, and large reductions of lime, magnesia, and ferrous oxide.

A relative enrichment of plagioclase with soda may be effected by the elimination of anorthite. In the reaction known as "saussuritization," the anorthite fraction of a calcic plagioclase is converted by addition of water into an aggregate of zoisite and sericite, with which calcite and chlorite are often associated. The following equation will serve to illustrate the transformation :—

$$\underset{\text{(anorthite)}}{2CaAl_2Si_2O_8} + H_2O = \underset{\text{(zoisite)}}{Ca_2(AlOH)Al_2Si_3O_{12}} + \underset{\text{(base of mica)}}{HAlSiO_4}$$

Plagioclase crystals which have suffered this alteration become semi-opaque by the development of minute scales of sericite and granules of epidote or zoisite; only the outermost zones show comparatively clear albite. In some cases augite and clear albite are reported to form ophitic intergrowths, and the albite then has the appearance of a primary mineral; but Bowen's study of the system diopside-plagioclase (see p. 95) makes it most unlikely that albite and augite can crystallize simultaneously from a magma.

E. S. Bastin traced the development of reddish veins of aplite or pegmatite, having all the appearance of igneous rocks, in the Nipissing diabase of Ontario. The veins may be 1-2 feet wide and a few hundred feet long. Some of them have sharp contacts with the diabase but others pass through reddish transition facies. The Nipissing diabase consists of labradorite, augite, and iron ore, with some interstitial micrographic intergrowths of quartz and albite. Bastin showed that with commencing alteration the labradorite becomes filled with sericite scales, while the augite is chloritized. Passing from the grey diabase toward the red aplite, the turbid feldspar clears up and eventually appears as simple albite, more or less reddened by scales of hematite. Bastin commented that "in general the red rock is so fresh appearing that it would naturally be interpreted as a direct crystallization from a magma, were it not for the evidence of progressive mineral transformation that we have just described." In his opinion the solutions that effected the transformation came from a deep magmatic source, perhaps the same source that supplied the diabase.

E. V. Shannon described a thick diabase sill at Goose Creek, Virginia, which contains veins of diabase pegmatite in its lower part. Some of the pegmatites contain essentially the same minerals as the normal diabase, namely labradorite, augite, magnetite, and microperthite, but others have albite as their only feldspar. There are also veins of simple albite-quartz rock with aplitic texture.

One more illustration of this effect may be given. R. C. Emmons has described a thick sill in Bridgland township, Ontario, which shows a progressive change, by minute gradation, from pyroxene-gabbro at the base to pegmatitic quartz-hornblende diorite at the top. Emmons considers that the residual liquor of the crystallized magma, containing alkalis and silica, tended to rise through the network of crystals. In doing so it reacted with the crystals, making the feldspars progressively more and more sodic, causing the transformation of augite into hornblende, and depositing quartz in the final stage of crystallization. Two completely similar instances have lately been described by R. Merriam in Oregon.

J. L. Gillson has shown that feldspar which has been albitized may undergo a further replacement by orthoclase. He has described, in the Pioche district of Nevada, the transformation of a noritic rock composed of labradorite, augite, and hypersthene, into a quartz-monzonite in which the feldspar is andesine containing corroded cores and ghosts of labradorite. Some potash feldspar was formed in the magmatic period, but after the rock has solidified there was a further replacement of quartz and plagioclase by microcline and sanidine. Gillson has also described the replacement of quartz by microcline, and of plagioclase by both minerals, in the granodiorites of the Pend Oreille district, Idaho. The "adularization" of basalt in the Sunda Islands was described by W. P. de Roever.

Remarkable evidence of replacement of soda by potash was found by C. N. Fenner in geyser basins in the Yellowstone Park. Rhyolitic and dacitic lavas which were penetrated by the drill, in deep boreholes, were found to be largely glassy in texture, with insets of quartz, plagioclase, and orthoclase. The ratio of $Na_2O : K_2O$ in a perfectly fresh sample was 4·24 : 2·94. In certain other lavas the groundmass had been recrystallized to an aggregate of quartz and orthoclase, and in these the ratio was 1·02 : 6·16 and upwards. The pyroxene is the same in both types of rock, but Fenner points out that the soda-rich insets in the second type are not compatible with the potash-rich groundmass. He holds that the geyser waters attacked the glassy base of the lavas and replaced soda by potash.

A fine study of potash enrichment in lavas at Esterel, France, has been made by R. Terzaghi.

REPLACEMENT OF FELDSPAR BY ANALCIME OR SCAPOLITE

Instead of being albitized, plagioclase is sometimes replaced by analcime. In this case no addition of silica is required and one may write the illustrative equation

$$\underset{\text{(plagioclase)}}{CaAl_2Si_2O_8 + 2NaAlSi_3O_8} + Na_2CO_3 + 4H_2O = \underset{\text{(analcime)}}{4NaAlSi_2O_6.H_2O} + CaCO_3.$$

J. T. Lonsdale found analcime-bearing rocks to be very common in the Terlingua-Solitario region of Texas. The rocks range from syenites to quite melanocratic types. The analcime is of late formation, the microscope indicating that "its period of formation extended from a very late magmatic stage into a period definitely post-magmatic." In the larger bodies of rock, which cooled relatively slowly, extensive reaction took place between the residual liquid and the earlier formed crystals. The initial stages of reaction were thought by Lonsdale to be late-magmatic. Among the products were aegirine rims on the augite crystals, replacement of pyroxene by biotite, and the development of analcime in ocellar form. Later reactions which are certainly post-magmatic gave such products as vesicles filled with analcime, calcite and zeolites. Euhedral crystals of analcime were even produced in the adjacent sedimentary rocks.

The replacement of plagioclase by analcime has taken place in many Scottish and Moravian teschenites, which have been described by G. W. Tyrrell and F. Walker. The process of replacement is beautifully demonstrated by a dike rock (lamprophyre) that cuts limestone at Rudeville, New Jersey, and was described by J. F. Kemp. The rock consists of plagioclase, pyroxene, and biotite, with analcime forming irregular patches in the groundmass and also little spheroids which Kemp supposed to be pseudomorphs after leucite. It is perhaps more likely that they are former gas vesicles. However that may be, the spheroids consist of clear analcime containing "ghosts" of plagioclase in every stage of corrosion and replacement.

C. S. Ross described a dike of olivine-basaltic composition, in Colorado, which contains patches of analcime-syenite. The analcime fills angular interstices and also replaces feldspar. The oligoclase cores of the feldspar crystals are more or less completely replaced by analcime, but the orthoclase mantles are scarcely affected. Ross writes that "the analcite that is secondary to feldspar is without doubt a deuteric mineral formed near the close of crystallization by reaction between residual solutions . . . and sodic plagioclase." R. H. Jahns has described other analcime-bearing dikes in Colorado and has arrived at a similar conclusion. J. Gilluly recorded sills of analcime-diabase in Utah, which have cores of analcime-syenite just like the rock described by Ross. The same conclusion is reached, that the analcime is partly primary and partly a replacement of feldspar.

H. W. Fairbanks observed four modes of occurrence of analcime in diabase dikes of San Luis Obispo county, California. Hexagonal or rounded grains of analcime are partly enclosed in feldspar; these are assumed to be primary. Analcime also fills angular interspaces between the feldspars, and it partly replaces feldspar. Finally, crystals of analcime line the walls of cavities.

The replacement of leucite by analcime, and the production of "pseudoleucite," are described in chapter xviii.

Replacement of plagioclase by scapolite is widespread in the Kiruna district of Norway. W. C. Brögger (2) gave the special name "ödegardite" to a rock formed by the transformation of hyperite into an aggregate of hornblende and scapolite. All stages of transition from hyperite into ödegardite may be found, and Brögger maintains that the change is due to "the influence of hot solutions and pneumatolytic gases in connection with the magmatic genesis" of the hyperite. J. H. L. Vogt also thought the necessary solutions containing NaCl were derived from the hyperite magma. A. Bugge was of a different opinion; he held that the origin of the scapolite and the apatite rocks of this region might be connected with the intrusion of dikes of granite-pegmatite; that is, that the scapolite rock is a product of thermal metamorphism of hyperite. This view derives a certain probability from the observation that some of the altered hyperites contain relics of coronas, and there is reason to think that coronas are products of thermal metamorphism (S. J. S., 3). N. Sundius has described the extensive scapolitization in the Kiruna district as a "regional pneumatolytic metamorphism."

On the other hand, A. Lacroix has described the case of a diabase dike which, in his opinion, has been scapolitized by salt water from adjacent saline clays.

This striking conflict of opinion emphasizes the truth that we do not always know the source of the solutions that promote post-magmatic reactions. We must be guided in each case by the field evidence.

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We circle around truths and, if patient, get ever nearer.—
J. E. Spurr.

CHAPTER XI

THE GENESIS OF PEGMATITE

The origin of pegmatite has been hotly debated for more than a century. The oldest recorded opinion on the subject declared that "pegmatites are injections of granitic matter which, arising from the still liquid granite in the interior, was injected into the upper, already solidified granite" (Charpentier, 1823). Opposed to this was the "lateral secretion" theory, first applied to ore-deposits but later extended to pegmatite. H. Credner wrote in 1875 that "the mineral matter of granitic dikes . . . comes from the partial decomposition and leaching of the wall rocks by seepage water."

As far as the simple quartz-microcline pegmatites are concerned, either of these theories might find support to-day; in fact, G. H. Williams claimed that in Maryland, intrusive pegmatites and segregation pegmatites which are quite similar in appearance may occur side by side. But a powerful argument in favour of injection rather than segregation is the frequent lack of any chemical or even mineralogical resemblance between a pegmatite and its country rock. W. C. Brögger illustrated this argument by reference to the granite-pegmatites of Hitterö and Arendal, Norway, which are mineralogically similar to one another and also to an adjacent body of granite; yet the Hitterö pegmatites occur in anorthosite and norite, the Arendal pegmatites in gneiss and schist. Brögger pointed also to the ægirine-granite pegmatite of Rundemyr, which cuts limestone at a distance of half a kilometre from a body of ægirine-granite.

G. H. Williams observed in Maryland that "there seems to be no relation between the character of the adjoining rock and the particular kind of pegmatite which occurs within it." Crosby and Fuller reported that in New Hampshire "we have found pegmatites of strikingly uniform composition traversing in succession, and without sensible change, a whole series of granites, gneisses, and schists of the most diverse origin." W. T. Schaller showed that the famous lithia-pegmatites of Pala, California, occur mostly in gabbro but sometimes in granodiorite; and O. Andersen found that in the pre-Cambrian region of southern Norway, typical microcline-granite pegmatites are most abundant in gabbro, diorite, and amphibolite. A parallel observation was made by T. G. McLaughlin, at Wind River Canyon, Wyoming, where veins of microcline-granite pegmatite invade a schist composed of hornblende and labradorite. A body of microcline-granite occurs in the vicinity.

Further evidence to the same effect is given by the corundum-bearing pegmatites of the United States, Transvaal, and India, which are found in peridotite or serpentine, usually at no great distance from an intrusive body of granite. In Seiland, Norway, according to T. F. W. Barth, both corundum pegmatites and nepheline-pegmatites occur in peridotite but not in the adjoining gneiss. It seems impossible to explain these facts in terms of lateral secretion.

W. C. Brögger made the curious observation that the rare minerals in the granite-pegmatites of the Christiania (Oslo) district of Norway are to a large extent different from those in the syenite-pegmatites of the same region. The granite-pegmatites hold especially columbates, tantalates, and phosphates; the syenite-pegmatites and nepheline-pegmatites hold zirconium minerals and species containing fluorine and boron. F. L. Hess (1) recognized four distinct types of pegmatite in North Carolina, namely zircon-sphene pegmatites, uranium-rare earth pegmatites, lithium-tin pegmatites, and simple granite pegmatites without any rare minerals. G. Switzer found four types of pegmatite within a single small granite stock in Colorado. These observations are not easily reconciled with the lateral secretion theory.

F. L. Hess (2) observed that the country rock about some pegmatites is greatly altered; large quantities of silica as well as certain uncommon minerals have been introduced into the country rock, but there is little evidence of any transference of matter from country rock into pegmatite. Thus at Camp Harding, New Mexico, the wall rock (quartz-mica schist) was so largely replaced by lepidolite that it was at one time mined, along with the pegmatite, for its lithium content. In India, the joint work of Roy, Sharma, and Chattopadhyay showed that the pegmatites of Kodarma "have not taken any material from the wall rocks, but on the other hand they have added plenty of material to the latter." The mica-schist near the pegmatite veins has been feldspathized, and eyes of oligoclase have developed in it, while abundant crystals of tourmaline and apatite appear both in the pegmatite and in the adjoining schist.

In the case of large plutonic intrusions, we have seen that reaction with the wall rocks has generally been of a reciprocal nature, the magma giving up some constituents to the wall rock and receiving others in exchange. It need not surprise us, then, to find some evidence of reciprocal action between pegmatite and its walls. Large crystals of andalusite are sometimes found in granite-pegmatite which cuts slate or schist. J. J. Runner has recorded both andalusite and sillimanite in pegmatites of the Black Hills, South Dakota, and holds that they are due to contamination of the pegmatite magma with aluminous sediment. R. W. Webb has described pegmatite containing giant crystals of andalusite, in Riverside County, California. O. Andersen described a plagioclase-cordierite pegmatite within cordierite-gneiss at Risör, Norway; also garnetiferous pegmatite in garnet-gneiss and epidote-bearing pegmatite in epidote-mica schist. It was his opinion, however, that in these pegmatites "the reaction has been more a recrystallization than an assimilation" of the introduced material.

J. Lehmann conceded that "a certain dependence upon the country rock cannot be denied" in the pegmatites of the Saxon granulite region. He found orthoclase-rich pegmatites in association with orthoclase-granulite; oligoclase-biotite pegmatites with plagioclase-granulite; and rare hornblende-epidote pegmatites with pyroxene-granulites and amphibole-schists. We may recall Crosby and Fuller's observation that the mica-pegmatites of New Hampshire are more micaceous where they cut mica-schist than where they cut granite. In brief, it need not surprise us to find that certain pegmatites have taken on some of the characters of their wall rocks while others show no such relation. The same thing is true of granites in general. But the recognition of a certain mineralogical resemblance between a crystalline rock and its wall falls short of proof that the former was generated from the latter. Not one of the examples cited above can be held to support the lateral secretion theory *in toto;* whereas the numerous instances in which there is no mineralogical or chemical relationship between pegmatite and wall rock afford strong evidence against that theory, at least in the instances referred to.

About the year 1870, G. Vom Rath was impressed by the abundance of tourmaline, beryl, and lithia-mica in the pegmatites of Elba, and their absence from the surrounding granite. He assumed that the uncommon elements in these minerals had been supplied by ascending solutions from some deep-seated source. W. C. Brögger commented that this assumption could only be accepted if it was admitted that both the pegmatite and the massive granite were formed from a "hydatopyrogenic" magma, or a mutual solution of silicates and water. In this matter Brögger was of the same opinion as J. Lehmann, who wrote in 1884 that the granitic dykes of Saxony "have indeed been formed by the co-operation of water, yet it was not atmospheric water seeping into fissures but eruptive water, charged with mineral matter in solution, which was yielded by the underlying granite." Lehmann compared the pegmatite-forming solution to a hot silica jelly.

Between the "seepage water" theory of Credner and the "eruptive water" theory of Lehmann and Brögger it can hardly be denied that the advantage, as regards the explanation of the rare-element pegmatites, is with the latter. The seepage theory requires us to believe that water at little more than atmospheric temperature, seeping through the interstices of solid rocks and making contact only with the surfaces of the mineral grains, can leach the rare elements out of crystalline particles which only hold minute traces of these elements in solid solution. According to the eruptive theory, the rare elements were dissolved in a body of rock-magma which also held water in solution; thus water and these elements were mingled on an atomic scale from the beginning. When the water was liberated as a second liquid phase, in consequence of the crystallization of the bulk of the magma, it held in solution not only a residue of alkalis, alumina, and silica but also all the rare components which had not yet reached their saturation point. The solution collected in fissures and cavities, where it deposited the rare minerals together with feldspar and quartz.

W. C. Brögger was the first to show that chemical replacement has played a part in the generation of the complex pegmatites. He distinguished three periods of growth, a magmatic period, a pneumatolytic period (see back, p. 162) and a hydrothermal period; and he described the replacement of orthoclase by albite, of nepheline by sodalite, and other post-magmatic changes. A Lacroix, in his great work on the minerals of Madagascar, contributed further observations of this nature, especially in relation to the lithia-pegmatites. He wrote: "There is no doubt that the lithia minerals are of secondary, pneumatolytic origin. . . . The lithia phase is always the last act in the history of granitic magmas."

Since 1925 much attention has been directed to replacement in complex pegmatites. In that year F. L. Hess (1) discussed the natural history of the pegmatites and emphasized the importance of replacement of country rocks by the pegmatite-forming solutions. W. T. Schaller demonstrated that the lithia-pegmatites of Pala, California, have been formed by a succession of replacements from an original graphic granite composed simply of quartz and microcline. In the first stage of replacement both microcline and quartz were partly replaced by albite, but the graphic texture of the rock was not wholly destroyed. The displaced potash took the form of muscovite. At a later stage, garnet and black tourmaline were formed at the expense of both kinds of feldspar; and still later, lithium silicates replaced to some extent all the minerals already named. In the pegmatites of Maine a similar series of replacements was traced by K. K. Landes, who judged that black tourmaline, beryl, and garnet were formed in the magmatic stage; lithia-tourmaline, spodumene, and lepidolite in the high-temperature hydrothermal stage; and lithium-manganese phosphates and other minerals at lower temperatures. A. L. Anderson claimed with regard to the pegmatites of Idaho that even tourmaline, beryl, and garnet are not primary crystallizations but replacements of earlier minerals. W. T. Schaller holds that "any one of these minerals may have more than one generation, and reversals of the order of crystallization also occur."

The succession potash-pegmatite, soda-pegmatite, lithia-pegmatite, has now been recognized in many lands and many localities. This order of succession is that of diminishing ionic radii (K. 1.33; Na ·98; Li ·78), which suggests that the small sodium ions are able to diffuse into a lattice containing potassium, and the still smaller lithium ions into a lattice containing sodium, while the reverse procedure is difficult or impossible. This is an attractive idea, but it is not always supported by the observed order of formation of the minerals. Thus J. G. McLaughlin has recorded pseudomorphs of albite after petalite (sodium replacing lithium) in a pegmatite in Wyoming; and Brush and Dana described pseudomorphs of microcline after spodumene in pegmatite at Branchville, Connecticut. B. M. Shaub says that at Newry, Maine, albite both preceded and followed the lithia minerals.

It is not necessary that all pegmatites should have had the same history, but most students of the complex pegmatites have found evidence of mineral replacement, usually in the order described by Lacroix, Schaller, and others. T. W. Gevers claims that in the great pegmatite region of Namaqualand, S. Africa, " there is hardly one species outside feldspar, β-quartz, zircon, and small crystals of apatite and spessartite, that was not shown by field evidence and laboratory examination to have originated by replacement." The solutions that effected these replacements in pegmatite are believed to be derived from the cooling magma. At least we are sure of this: that the solutions were hot and charged with mineral matter; that they followed channels within normal pegmatite veins and sometimes invaded the wall rocks; and that they followed the same channels for long periods, all the while cooling down and changing their composition by reacting with the solid silicates. Since residual solutions of the necessary character are expectable as end-products of the crystallization of granitic magma, most geologists have been satisfied to accept the magmatic origin of the mineralizing solutions and to see the complex pegmatites as products of the deuteric alteration of simple granite-pegmatite.

THE ASSOCIATION OF PEGMATITE WITH APLITE

One of the most puzzling of the many problems connected with pegmatite is the frequent association of aplite and pegmatite, representing opposite extremes of granularity, within the same body of granite. Sometimes there is a difference of age between them, and then either of them may cut the other; but in many instances pegmatite and aplite occupy the same fissure, sometimes one and sometimes the other assuming the central position. Alternatively, pockets of pegmatite may develop within an aplite dike, and streaks and patches of aplite may be preserved within a body of pegmatite. Exceptionally there may be a rhythmic alternation of aplite and pegmatite in successive curved sheets, resembling Liesegang rings on a large scale.

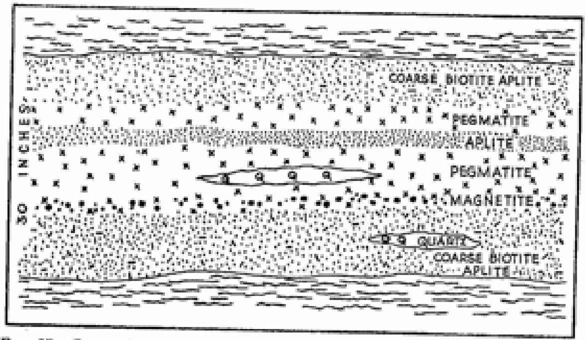

FIG. 27. Pegmatite with marginal and central aplite. Vermilion, Minnesota. (F. F. Grout.)

Aplite, like pegmatite, is a sack name which covers a wide range of mineral associations. It is one of the anomalies of our old-fashioned classification of rocks that the mineralogically interesting and commercially important pegmatites have received just one generic name, while the relatively uninteresting aplites have been burdened with more than a score of "specific" names of the usual geographical type. We shall ignore all such trivial names and confine our attention, in this discussion, to what Rosenbusch called "aplite in the narrower sense," and particularly to the only really common variety, granite-aplite. This is a fine grained dike-rock containing the same minerals as the parent granite but differing from it in two important respects. There is first the textural difference. In typical granite the minerals of early crystallization are idiomorphic; those of late crystallization (generally quartz and microcline) are typically allotriomorphic, having adapted themselves to such spaces as remain between the crystals of the earlier minerals. Many granites are porphyritic, or at least uneven-grained. But in aplite all the mineral grains have approximately the same size and none is better shaped than another, as if all had crystallized at the same time and at the same rate; thus every grain of the abundant minerals, whatever its composition, interferes to the same extent with every other grain. This texture is well described by the adjective "sugar-grained."

The second point of distinction between granite and aplite is that the latter is markedly deficient in dark minerals. Muscovite is common, but biotite is scarce and tourmaline and garnet are patchy and inconstant. The feldspar is the same as in granite, namely orthoclase, microcline, and sodic plagioclase. Mineralogically, therefore, aplite differs from normal granite in the same sense that pegmatite does. No satisfactory comparison of chemical composition can be made on account of the difficulty of sampling a really coarse pegmatite. Washington's Tables include some thirty analyses of pegmatites and graphic granites, presumably of the less coarse type, and nearly a hundred analyses of aplites, microgranites, microsyenites, and paisanites. Many of these analyses are incomplete, minor constituents having been neglected, but most of them include determinations of TiO_2 and P_2O_5 and a few supply data on ZrO_2, Cl, F, S, and Li_2O. A rough comparison shows that these minor elements and oxides have been recorded just about as often in the aplites as in the pegmatites, and that their proportions are of the same order of magnitude. Even the proportion of combined water is about the same. The rare minerals that give a special interest to pegmatites have not generally been observed in aplites, but the following minerals have been recorded in certain examples: zircon, xenotime, tourmaline, topaz, allanite, lepidolite, spodumene, ægirine, riebeckite. It is fair to conclude that there is no significant chemical difference between aplites and pegmatites which occur within the same body of granite.

To what, then, may we ascribe the astonishing difference in texture between aplite and pegmatite? Nobody doubts that the coarse crystallization of pegmatite, compared to that of normal granite, is connected with the greater concentration of water in the residual solution, and many writers refer confidently to the presence of "mineralizers" in the solution. But aplite is also a product of residual solutions, and the chemical data which we have cited fail to establish any chemical difference between aplite and pegmatite. The explanation of the textural difference must therefore be a physical one. In chemical laboratory practice one prepares large crystals by permitting a saturated solution to cool very slowly, without disturbance; whereas rapid cooling and constant agitation of the solution

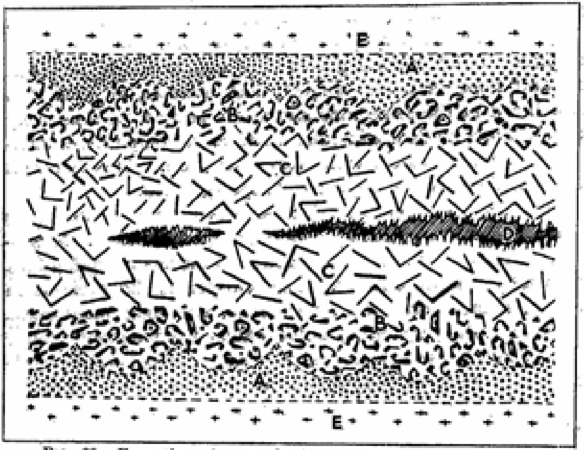

FIG. 28. Formation of pegmatite by recrystallization of aplite. (N. M. Uspensky.)
A, aplite; B, graphic granite; C, coarse graphic granite; D, druses.

will yield a crop of small crystals. It is difficult to believe that the crystallization of aplite was rapid, even in a relative sense, yet the number of centres of crystallization in aplite is many times that in pegmatite.

There is some evidence that the change from aplite to pegmatite may be due to recrystallization. N. M. Uspensky has discussed this possibility in connection with the pegmatites of the Borshov range, in Transbaikalia. These veins show bilateral symmetry, with coarse, drusy graphic granite in the middle. Next to this, on both sides, is a zone with perfect graphic texture; then a marginal zone of aplite, towards which the pegmatite presents a series of curved surfaces and a radial structure. The description reminds one of the "scalloped growths" in the pegmatite of Kinkel's quarry, at Bedford, New York, which are formed of alternating bands of pegmatite

FIG. 20. A vein of aplo-pegmatite in the Snoqualmie granodiorite, Washington State. Magnification 1¼. Description on p 188.

and aplite; and of the feldspar "brushes" described by E. S. Bastin at Topsham, Maine, and the hemispherical growths of cleavelandite found by V. C. Shainin at Branchville, Conn.

Fig. 29 shows an interesting relation between aplite and (micro-)pegmatite in a narrow vein traversing the Snoqualmie granodiorite, in Washington State. The middle member, only 15 millimetres wide, is a quartz-orthoclase aplite having a grain-size of less than 0·2 m.m. Some larger quartz grains lie irregularly along the middle line. The aplite is bordered on each side by a 10 mm. selvedge of micropegmatite consisting of blades of orthoclase, 5 to 10 mm. long, which enclose fan-like groups of quartz fibres. The wall rock is hornblende-biotite granodiorite in which the larger feldspars, measuring about 2 x 1 mm., are zoned plagioclase with mantles of alkali-feldspar. The passage from granodiorite to micropegmatite is strongly marked on account of the disappearance of the dark minerals, yet the junction is not an intrusive one, for the radiating texture of the micropegmatite shows clearly that the quartz-orthoclase intergrowth grew out from the granodiorite walls. The passage from pegmatite to aplite marks the limit of this growth. There is no indication that the pegmatitic crystals were truncated by the intrusion of an aplitic liquid, but only that the growth of these crystals was stopped by some change of physical conditions. These observations are favourable to the idea that the micropegmatite was formed by a recrystallization of the margins of the aplite vein, under the influence of residual solutions escaping from the granodiorite.

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Whosoever shall entertain high and vaporous imaginations, instead of a laborious and sober inquiry of truth, shall beget hopes and beliefs of strange and impossible shapes.—FRANCIS BACON.

CHAPTER XII

ERUPTIVE ROCKS AND ORE DEPOSITS.

THIS is not a study of ore-deposits. It does not start out from known occurrences of ore and attempt to establish their mode of origin. It is strictly a continuation of the subject matter of the last two chapters, and it attempts to investigate the *capability* of eruptive magma and solutions of magmatic origin to generate bodies of ore. If certain ores have their source in rock-magma, then it should be possible to trace the ore-forming elements all the way from place of origin to place of deposition, and to study the chemistry of the transporting solutions and of the reactions that caused the ore to be precipitated where we find it. We shall therefore reverse the usual method of approach. Instead of setting out from an ore deposit and taking sights on the nearest eruptive rock, we shall start from the petrologist's end of the road and see where it leads us. If we meet a body of ore-geologists on the way, well and good; if not, it will show that the road must be re-surveyed.

In this discussion we shall use the word "ore" in the broadest sense, to include such uncommon minerals as beryl, lepidolite, pitchblende, monazite, since all of these are "ores" in modern metallurgical practice. We shall be concerned only with those primary ores which are closely connected with eruptive rocks. With secondary or oxidized ore we have nothing to do.

The belief that residual solutions from crystallizing rock magma, which may already have generated pegmatites and quartz-veins, are capable of depositing ore in and about the parent rock, grew out of the studies of A. Daubrée on the genesis of tin deposits. The evidence, which we have already summarized (p. 161), proved so convincing that it had the effect of blinding many geologists to other possible sources of ore. How widely a belief in the magmatic origin of ore has been accepted, we may deduce from the following statements by recent writers on ore deposits :—

> " It is held that hypothermal and mesothermal deposits are formed by emanations from intrusive rocks, generally batholiths." (W. Lindgren, 1933.)
>
> " It is now generally conceded that a very large group of ore-deposits was deposited by solutions differentiating from magmas." (F. F. Osborne, 1937.)
>
> " Probably most geologists would agree that ore-deposits are generally derived from large bodies of igneous material." (C. N. Fenner, 1940.)
>
> " (Magmatic) waters and vapors have become almost universally accepted as the causative agents of most epigenetic ore deposits." (A. M. Bateman, 1942.)

It is interesting to note the choice of verbs in these statements : it is *held;* it is *conceded;* would *agree;* universally *accepted*. Rightly or wrongly, one gets the impression that each of these writers was expressing the fashionable view rather than a deep personal conviction. It is our purpose in this chapter to examine the foundations of this widely held belief in the power of magmatic solutions to generate ore deposits.

We have already discussed the composition of the residual solutions from crystallizing rock-magma. We need only recall that the average eruptive rock contains nearly sixty per cent of feldspar; that the later the feldspar crystallized, the richer it is in alkalis; and that the great majority of rocks contain no significant proportion of any acid stronger than silicic acid. The conclusion is unavoidable that the residual aqueous solution from the crystallization of such a rock must contain abundant sodium and potassium ions, together with weak aluminium-silicon-oxygen anions; that is, it must be definitely alkaline in character.

This is contrary to the opinion of many ore-geologists, who find it necessary to postulate acid solutions as carriers for the common heavy metals, and who point to the presence of chlorine, fluorine, and sulphur in most eruptive rocks that have been completely analyzed. But according to the data of Clarke and Washington, which we quoted on p. 31, the average eruptive rock contains only 0·09 per cent of chlorine, 0·08 per cent. of sulphur, and 0·03 per cent of fluorine. Even if these elements are entirely primary (a questionable assumption as regards both sulphur and chlorine), and even if the whole amount of each should be concentrated in the residual liquid (another questionable assumption), it is surely beyond dispute that these small amounts of strong anions must be outweighed by the weak aluminium-silicon-oxygen anions and the strong sodium and potassium cations, and consequently that the residual liquid from any feldspathic or feldspathoidal magma must be alkaline.

This conclusion does not shut out the possibility that a certain amount of acid *vapour* may be formed if the presence of open channels permits boiling of the alkaline liquid to take place. This possibility has been explored by N. L. Bowen, and it is thought to be illustrated by the Katmai fumaroles, the chemistry of which was studied by Allen and Zies. The steam was found to contain 0·117 per cent of HCl, 0·032 per cent of HF, and 0·029 per cent of H_2S. The mineral incrustations about the fumaroles were studied by E. G. Zies, who found the most abundant metalliferous deposit to be magnetite. Chemical and spectroscopic analysis of the incrustration showed a small content of lead, zinc, copper, and manganese, as well as traces of molybdenum, tin, nickel and cobalt.

If acid solutions are necessary to explain the genesis of certain ore deposits which are assumed to be of eruptive parentage, then boiling of the residual liquid from crystallizing rock-magma seems to be the only way in which these solutions can be supplied. With the Katmai illustration before us, it is not difficult to believe that certain near-surface (epithermal) ore deposits may have been formed or modified by acid vapours, but it would not be permissible to extend this conclusion to deep-seated (hypothermal) deposits. It is most unlikely that boiling can take place under plutonic conditions. G. W. Morey has discussed this question; he considers that under deep-seated conditions the pressure will always be so high as to make boiling impossible. P. Niggli seems to share this opinion, but Bowen thinks it possible that "pegmatitic liquid sometimes boils even at rather considerable depths." In view of this difference of opinion, and in the total absence of direct evidence, it will be wise to concede the possibility that acid vapours may participate in the formation of *some* ore deposits. Yet when we compare the high proportion of strong bases, in the average eruptive rock, with the trivial proportion of strongly acid radicles, it is difficult to evade the conclusion that the problem of eruptive ore deposits must be discussed primarily in terms of alkaline solutions.

In this discussion we shall find it convenient to follow J. H. L. Vogt (1) by making a distinction between *oxidic* ores and *sulphidic* ores. The former are oxides and oxygen salts (silicates, aluminosilicates, phosphates, titanates, etc.); the latter are sulphides and the related arsenides, antimonides, selenides, tellurides. A parallel distinction was made by V. M. Goldschmidt between *lithophile* and *chalcophile* elements. The latter term, meaning "copper-loving," does not seem well chosen; it will be more appropriate to use *thiophile* (sulphur-loving) for the elements in question. Lithophile elements yield oxidic ores; thiophile elements yield sulphidic primary ores. The basis of the distinction rests on the energy of combination of the elements. Those that have a high energy of oxidation have combined with oxygen in nature and are lithophile; those with a low energy of oxidation have generally failed to secure oxygen and have instead combined with sulphur or arsenic; they are thiophile. A few metals, particularly iron but also manganese and tin, form both oxidic and sulphidic primary ores.

THE GENESIS OF OXIDIC ORES.

We have already reviewed the evidence relating to the tin-tungsten ores (p. 161) and the pegmatites (p. 178). This evidence has convinced most if not all geologists that the tin-tungsten group of minerals and the lithium-beryllium-tantalum group were formed by magmatic residual solutions escaping from bodies of granite. Proof is beyond our power, but the evidence is so strong and so consistent that the conclusion is probably as soundly established as most geological beliefs.

This conclusion is justified by the chemistry of the metals concerned. In a solution which contains chiefly sodium and potassium ions, together with weak aluminum-silicon-oxygen anions, one might expect to find small quantities of certain related elements, namely:

(1) elements related to sodium and potassium;
(2) elements having some relationship with aluminium;
(3) elements whose oxides form soluble salts with the alkalis, in this respect resembling silica.

Group 1. The elements lithium, rubidium, and cæsium are so closely related to sodium and potassium that it is a matter of considerable difficulty to separate them. The ionic radii are as follows:

Li 0·78; Na 0·96; K 1·33; Rb 1·49; Cs 1·65,

hence lithium is to be expected in association with sodium, cæsium and rubidium with potassium. Lithium differs from the other alkali-metals in not forming a feldspar, yet the albite-rich pegmatites are the chief source of lithium ores, and sodium and lithium are mutually replaceable in such minerals as spodumene, tourmaline, amblygonite, lithiophilite and natrophilite. Rubidium oxide has been found to the amount of more than three per cent in microcline and over one per cent in mica (F. W. Clarke). A little cæsium accompanies the other alkali metals in some varieties of mica and beryl, and it forms the rare silicate pollucite which is related in composition to leucite.

Group 2. Nearly related to aluminium are the rare elements of the cerium group, together with yttrium, scandium, and gallium. These elements are both trivalent and quadrivalent. When trivalent, they seem to be interchangeable to a certain extent with aluminium, as in the mineral allanite. Boron too is believed to substitute for aluminium in some silicates, notably tourmaline and danburite. Beryllium, which of all the metals has the smallest atoms, is always associated with aluminium and silicon, the elements closest to it in ionic radius (Be 0·34; Si 0·39; Al 0·57). Goldschmidt suggests that beryllium may to a small extent replace either aluminium or silicon.

Group 3. The oxides of a number of metals, mostly those of high valence, are either amphoteric or definitely acidic in character and consequently form soluble salts or at least hydrosols with the alkalis. Aluminium and beryllium fall into this category, and the others are as follows:—

Valence 2: lead, zinc, manganese;
,, 3: vanadium, chromium; antimony, arsenic;
,, 4: zirconium, hafnium, thorium, titanium, tin, gallium, germanium, molybdenum;
,, 5: columbium, tantalum, bismuth;
,, 6: tungsten, molybdenum, uranium.

Most of these metals have more than one valence, but the primary ores, as a rule, exhibit only one of the possible valences. Five of the metals named are thiophile and may be omitted from this discussion; these are lead, zinc, arsenic, antimony, and germanium. Bismuth trioxide has no acidic character, but the pentoxide Bi_2O_5 is feebly acidic and forms salts such as sodium bismuthate; this compound, however, is immediately reduced by ferrous and manganous oxides, so it could not possibly be present in rock-magma. Nevertheless bismuthinite (Bi_2S_3) and native bismuth are rare constituents of some pegmatites. Molybdenum, in its commonest ore, molybdenite, is quadrivalent and then thiophile, but a little molybdenum is associated with tungsten in scheelite and is then hexavalent and lithophile. (Wulfenite and other rare molybdates are oxidized, not oxidic, ores.)

In the case of vanadium, chromium, and manganese, the stable oxides V_2O_3, Cr_2O_3, and MnO are basic and do not form salts with the alkalis. The higher oxides V_2O_5, CrO_3, and MnO_2 are acidic, but they are only generated under strongly oxidizing conditions which are not realized in deep-seated rock-magma; this is an unescapable deduction from the almost invariable presence of ferrous oxide in eruptive rocks. (The chromate, crocoite, and the vanadate, vanadinite, are oxidized, not oxidic ores.) Consequently one would not expect to find vanadium and chromium in association with the other

Chemical analysis, supplemented by the spectroscope, has revealed traces of all these elements, as well as many rarer ones, in eruptive rocks. Sedimentary and metamorphic rocks have not been analyzed so frequently or so completely as eruptives, but there is no reason to doubt that they too carry small quantities of the thiophile elements. Nickel, cobalt, zinc, copper, lead, silver, and gold have all been detected in sea-water and must therefore find their way into sedimentary rocks. A manganese nodule dredged from the ocean floor by the *Challenger* expedition was found to contain weighable proportions of nickel, cobalt, zinc, copper, lead, thallium, vanadium, and molybdenum. It would not be wise, therefore, to assume that all ore is derived from eruptive rocks. The data of Clarke and Washington show that the total amount of the five thiophile elements, nickel, copper, zinc, cobalt, and lead, in the average eruptive rock, is less than the amount of zirconium alone; and even the rare-earth metals (cerium, yttrium, etc.) are more abundant than copper, lead, or zinc.

Nevertheless a few sulphidic ores are common minor constituents of eruptive rocks. Chief among them are pyrite, pyrrhotite, pentlandite, and chalcopyrite. It used to be thought that these ores had crystallized along with the associated silicates; but the introduction of the reflecting microscope has shown in countless instances that the sulphidic ores are not contemporaneous with the rock-forming silicates but were introduced after the completion of the magmatic stage of crystallization. As an illustration we may cite the Engels copper deposit in California. It was claimed in 1914 that "the ores are of direct magmatic origin"; yet Graton and McLaughlin showed, a few years later, that "the ores, instead of being magmatic in the sense that they were initial constituents of the dioritic rock in which they occur, were introduced after the rock had solidified and had suffered notable dynamic and chemical changes."

It appears in an increasing number of instances that the distribution of ore in an eruptive rock is controlled by fissures. This is the case at Bingham and Butte, where the ore occurs in a much shattered quartz-monzonite. At Sudbury, according to Tolman and Rogers, "Very often the ore is concentrated in those portions of the intrusive which have suffered brecciation during intrusion." In a recent study of the Mount Prospect area, Connecticut, E. N. Cameron found that the distribution of the iron-copper-nickel sulphides in norite was controlled by "fractures, microfaults, microshear zones, and mineral cleavages." Evidence of this kind does not disprove the magmatic origin of the ore but makes it less certain than it seemed to an earlier generation of geologists.

The same pegmatites and quartz veins that carry oxidic ores may carry sulphides too, particularly pyrite, arsenopyrite, molybdenite, occasionally bornite or chalcopyrite, and sometimes bismuthinite or galena. Nevertheless the vast majority of pegmatites and quartz veins in and near eruptive bodies are barren of ore, and there is no evidence (within the writer's knowledge) that quartz veins of eruptive parentage ever lead or grade along the strike into large bodies of sulphidic ore. According to L. C. Graton, the experience of mining geologists is that when ore veins and pegmatites occur in the same region, the ore veins are not continuous with the pegmatites but nearly always intersect them. Graton therefore concludes that "most ore-deposits did not come through the pegmatite lineage."

From the chemical angle, there is no apparent reason why the thiophile metals should be present in magmas and solutions of which the main components are silica, alumina, and alkalis. According to J. H. L. Vogt (2) the sulphides of cobalt, nickel, lead, silver, and copper are only soluble in silicate melts in minute amounts of the order of 0·01 to 0·025 per cent. It is true that even these minute quantities might give rise to sizeable ore deposits if all the sulphides in a large body of rock-magma could be collected and concentrated in one spot; but then the question arises whether a hydrous magma or solution would deposit the metals as sulphides or as oxides? In the presence of water the following reaction may be expected to take place:

$$RS + H_2O \Longleftrightarrow RO + H_2S.$$

If H_2O were in considerable excess over H_2S, this reaction would proceed towards the right and the heavy metals would be precipitated as oxides (or other oxygen compounds) instead of sulphides. Only a very high concentration of H_2S in the magma or solution could bring about the precipitation of sulphides, and there is no evidence that the concentration is high. The gas collections made at Kilauea (see p. 35) show water in great excess over sulphur, and the Katmai fumaroles (p. 192) yielded only trifling amounts of H_2S. The chief deposit of these fumaroles was magnetite.

Some writers have invoked alkali sulphides as solvents for sulphidic ores. Concentrated solutions of alkali sulphides will dissolve the sulphides of certain metals, forming thio-salts such as sodium thiostannate (Na_2SnS_3). Magmatic solutions may perhaps contain a small proportion of alkali sulphides, but there is no suggestion that they are ever concentrated solutions of these, so it seems that their capacity for transporting sulphides of the heavy metals is negligible. R. M. Garrels found that iron is soluble only to a very small extent in sodium sulphide melts, so he claims that "alkali sulphide solutions are incapable of carrying more than minute amounts of iron in solution."

May we then suppose that the heavy metals were present in the magma in the form of silicates, and that the sulphide ions were contributed by ground water? This would remove one of our difficulties. It would be possible to explain on these lines the frequent occurrence of the sulphides of iron, nickel, and cobalt in magnesium-rich rocks such as gabbro and norite. The ionic radii of these four metals are:

Mg 0·78; Ni 0·78; Co 0·82; Fe″ 0·83.

It is well known that these metals are capable of a high degree of ionic substitution in their salts. Magnesian minerals such as ortho-pyroxene and olivine commonly hold a considerable percentage of ferrous iron and often a little nickel (sufficient when concentrated by weathering to form commercial nickel deposits). It is possible that nickel and cobalt, as well as iron, were originally contained in the magnesian silicates of gabbroic rocks, and that they were later converted into sulphides by percolating solutions containing hydrogen sulphide, derived from the country rocks. But then the three heavy metals could hardly be set free to combine with sulphur unless the magnesium silicates were deeply decomposed, and in many instances the rock is notably fresh.

The actual proportions of cobalt, nickel, vanadium, and chromium in certain magnesian rocks have been determined spectroscopically by R. L. Mitchell and by P. H. Lundegardh. Some of their data are reproduced below.

	Olivine-gabbro Greenland	*Quartz-gabbro* Sweden	*Peridotite* Sweden
V	·021	·05	·006 per cent
Cr	·023	·008	·0005 ,,
Co	·007	·004	·02 ,,
Ni	·020	·009	·045 ,,

To form a workable ore deposit from these trivial quantities of metal, dispersed through a large body of rock-magma, demands a remarkably efficient process of concentration. We have already formed the opinion that magmatic residual solutions do not have the necessary chemical character to transport and deposit sulphides of the heavy metals. Acid vapours escaping from the magma at places of reduced pressure may transport these metals in the form of chlorides or fluorides, but the example of Katmai shows how very small the amount so transported is likely to be. To explain the concentration of large bodies of sulphidic ore, another mechanism must be sought.

The very complete separation of matte from slag in metallurgical operations has suggested that *exsolution of liquid sulphides* from silicate magma may take place in nature. This hypothesis was first employed by A. P. Coleman to explain the copper-nickel ore bodies at the base of the Sudbury lopolith. Coleman showed that there is a direct relation between the thickness of the great gabbro sheet and the bulk of the ore bodies beneath it; and further, that the massive sulphide ore often occupies hollows in the floor of the intrusion, as if a heavy sulphide liquid had accumulated there. P. A. Wagner arrived at a similar explanation of the genesis of the copper-nickel ores in the Bushveld complex. Wagner expressed the opinion that "the disseminated sulphide specks (in the norite) were in their original form droplets of non-consolute iron-nickel-copper matte"; and that "there is the strongest presumptive evidence in favour of regarding the ore-bodies as local aggregations of such matte droplets."

Further support for this hypothesis comes from D. L. Scholtz, who has studied the copper-nickel ores of East Griqualand, South Africa. He maintains that "the ore which separated from the magma was concentrated in the basal zone by gravitational differentiation, to form pools within hollows and irregularities in the floor of the intrusion." Scholtz computed that droplets of ore, two millimetres in diameter, would sink through the magma almost four times as rapidly as olivine crystals of the same dimensions. Still another supporter of the exsolution hypothesis is G. C. Kennedy, who has applied it to the copper-nickel deposits of Yakobi Island, S.E. Alaska.

These observations made in different regions by Coleman and others almost compel us to recognize exsolution of sulphides as a real and active process. But this view has met considerable opposition at Sudbury, where there is evidence of hydrothermal action. A more serious objection is that the proportion of thiophile metals in ordinary gabbro (excluding iron in the oxidized state) is so small that it is difficult to understand how the ore could be so effectively separated from the abundant silicate matrix by gravity alone, and why it should have separated in the form of sulphide rather than oxide or silicate. Until we have experimental evidence of the solubility of sulphides of the heavy metals in liquids of the composition of gabbroic magma, it will be unwise to accept exsolution of sulphides as a complete explanation of the iron-copper-nickel ore bodies. Even if exsolution of these ores were proved, it would only supply part of the answer to the problem of sulphidic ore in general. It would leave the intrusive ore veins unexplained, especially those which contain gangue as well as ore. For these an explanation is still to be found.

In 1923, J. E. Spurr abandoned all the traditional concepts of ore-genesis and claimed that many of the massive sulphide deposits have been formed by direct injection of an "ore magma." He affirmed, with regard to certain Mexican deposits, that "the ore was intrusive in a fluid but highly concentrated state, containing no more water than basic igneous rocks"; and further, that "the Potosi veindike is the result of crystallization of a relatively dry and viscous ore magma." In the case of the Mendota vein, in Georgetown, Colorado, Spurr wrote that it opens up from a mere crack to "a strong vein one to four feet wide, filled with blende (with pyrite) inclosing numerous angular fragments of the wall rock. . . . These fragments are neither rounded, corroded, nor replaced." He therefore concluded that the vein was filled all at once, "as if a solution froze into metallic sulphides."

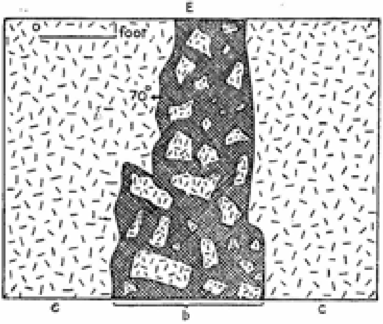

FIG. 30. Horizontal section of part of the Mendota vein, Colorado. Isolated lumps of granite are enclosed in solid ore. b, blende, c, granite. (J. E. Spurr.)

Spurr says that the ore-magma, as he conceives it, "is not the residual fluid expelled from igneous rocks during forced crystallization, but is a special magma developed at the deep zone where rock differentiation takes place. . . It is a highly specialized form of igneous magma."

At first sight, Spurr's hypothesis does not seem to differ widely from that of Coleman, Wagner, and Scholtz. All four authorities derive an ore fluid by exsolution (or more vaguely by "differentiation") from a silicate magma. But for Coleman, Wagner, and Scholtz, exsolution took place after the gabbroic magma had reached its final position and some degree of crystallization had occurred; whereas for Spurr the differentiation of the magma was deep-seated and the ore-magma *followed* the silicate magma into place. Spurr wrote as follows: "In general the igneous rock and the specialized ore magma are close together in depth, so that the first adjustment fractures of the consolidating intrusive tap the latter" (*i.e.*, the ore magma). It may be possible to harmonize these conflicting views, but only Spurr is in a position to do so.

A further elaboration of the ore-magma concept has given us the "abyssal theory," according to which the separation of ore from silicates took place when the earth was cooling down from the liquid stage of its development. C. H. White contends that "the ores of the heavy metals for the most part were depressed to abyssal depths before a solid crust was formed, and were later carried up and deposited in the crust when it was fractured by earth movements." White's idea seems to combine V. M. Goldschmidt's concept of a zone of sulphides and oxides, lying between the lithosphere and the iron core of the earth, with Spurr's theory of intrusive ore-magmas.

It is unprofitable to discuss what may happen in abyssal depths, but one is left wondering how a dense ore-magma could be raised from such great depths to the upper levels of the lithosphere. Goldschmidt's sulphide-oxide zone was supposed to lie seven to eight hundred miles beneath the earth's surface, and most geologists have assumed a thickness of this order for the silicate shell of the earth.

CONCLUSION.

We have seen that there is ample support, from the chemistry of rock magmas, for the claim that ores of tin, tungsten, tantalum, beryllium, and other lithophile elements are derived from bodies of eruptive magma. But in the case of the thiophile elements, evidence of this nature is scanty and unconvincing. The small amount of direct evidence that we possess indicates that the solubility of the thiophile metals, in silicate magmas containing sulphide ions, is very slight; and that the residual aqueous solutions from these magmas are incapable of transporting the sulphides of the heavy metals in appreciable amount. Graton's conclusion that sulphidic ore deposits do not come through the pegmatite lineage seems to be fully justified. Transport by acid vapours may account for some relatively small bodies of epithermal ore, but this explanation is not applicable to the much more extensive hypothermal deposits. Exsolution of sulphidic ore in liquid form, from magmas of gabbroic composition, is supported by strong field evidence; but this process does not seem capable of affording an explanation of all sulphidic deposits. Experimental research on the behaviour of metallic sulphides in hydrous silicate melts is greatly needed; without it, many of the claims that have been made about the magmatic origin of sulphidic ores must be held to lack adequate backing.

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> "*Die Petrographie ist die Urkundenlehre der Erdgeschichte, und somit eine historische, nicht eine lediglich beschreibende Wissenschaft. Hierin liegt es bedingt, dass eine natürliche Systematik der Gesteine historisch, d. h. genetisch sein muss.*"—H. ROSENBUSCH.

CHAPTER XIII

THE CLASSIFICATION OF ERUPTIVE ROCKS

CLASSIFICATION begins when people give names to things. When certain rocks, such as basalt, porphyry, and granite were given these distinctive names to mark certain differences in their appearance, the classification of rocks was set afoot. This kind of classification is what John Stuart Mill calls "a mere incidental effort consequent on the use of names," and because popular names are generally given to things in a very loose way, without careful discrimination, so the resulting classification is a very imperfect thing. We shall find it convenient to refer to it as a *classification of the first kind.*

At a later day, when the objects attract the critical attention of science, another kind of classification becomes necessary, in which the most suitable grouping of the objects is the first consideration, and the naming of them is a secondary matter which is governed by the grouping. This we may call a *classification of the second kind*. Of this kind of classification, Mill says

> "Classification, thus regarded, is a contrivance for the best possible ordering of the ideas of objects in our mind; for causing the ideas to accompany or succeed one another in such a way as shall give us the greatest command over our knowledge already acquired, and lead most directly to the acquisition of more."
>
> "The general problem of classification, in reference to these purposes, may be stated as follows: To provide that things shall be thought of in such groups, and those groups in such an order, as will best conduce to the remembrance and to the ascertainment of their laws."
>
> "The ends of scientific classification are best answered when the objects are formed into groups respecting which a greater number of general propositions can be made, and those propositions more important, than could be made respecting any other groups into which the same things could be distributed."

Now petrography has a highly developed classification of the first kind, for more than six hundred distinct names (not counting compound names) have been given to eruptive rocks to mark minute points of difference between them, and new names are still being coined at the rate of ten or twelve a year. But of the way in which this classification has grown up, it has been said with entire justice that "the method of petrographers in proposing names, so far as any method is discoverable, is to give an independent name to each rock which is slightly different from any previous rock found, without reference to any definite plan of nomenclature." (C. R. van Hise.)

To emphasize petty differences between one rock and another, while ignoring fundamental points of resemblance, is the very antithesis of classification. It is proper to establish varieties *within an accepted framework*, but to set up a multitude of so-called species without regard to any framework at all is simple anarchy.

To make matters worse, the names that have been longest in use, which were in the beginning mere popular names of the loosest kind, based on colour and other trivial characters, have gradually acquired a kind of sanctity and have dominated all attempts at classification, instead of the classification dominating the names. Some of our most used names, like granite, syenite, porphyry, obsidian and basalt are of great antiquity; others including gabbro, diabase, diorite, trachyte and phonolite date from the beginning of the nineteenth century, when geology was still undecided whether such rocks had been formed by fire or by water. The names have sometimes changed their meaning, but through all changes they have persisted and have dominated every attempt at classification.

If we wish to set up a classification of the second kind, to take the place of the disorderly rabble of names that passes for a classification now, we must first resolve that we will not allow ourselves to be dominated by names. If an existing name, indicating a supposed "species" of rock, cannot be made to conform to the boundaries of a new and well reasoned classification, it is not to be assumed that the method of classification is wrong. It is much more likely that the name is wrong, in the sense of having been given without due discrimination between like and unlike or essential and inessential.

But it is not only against ill-defined rock names that we have to be on guard; it is no less necessary to protect ourselves against premature and ill-conceived hypotheses about primordial magmas, magmatic differentiation, consanguinity, petrological affinity, and similar rigmarole. We shall make no progress in understanding and classifying rocks until we cast all such speculative lumber overboard, along with the belief that every rock name defines a "natural species."

In the quotation which we have placed at the head of this chapter, H. Rosenbusch affirmed that a natural system of rocks must be historical, that is, genetic. With this we agree, but add that it must be based on the *facts* of rock genesis and not on any hypothesis. The genesis of an eruptive rock is the cooling-history of a rock-magma. It follows that the character of a rock is determined by the chemistry of the magma, the physical conditions under which crystallization took place, and the stability-relations of the rock-forming minerals; in short, the problems of petrology are problems of physical chemistry. The classification of rocks must therefore be based on physical chemistry, as far as our slender knowledge permits.

Now the chemistry of a rock-magma can never be the subject of direct study; we must be content to deduce what we can from the chemistry of the rocks formed by it. The physical conditions can only be studied directly in the case of surface magma; for deep magma we have to rely upon deductions from the texture of the rock and its geological environment. But the minerals formed in the course of crystallization can be observed directly, and their stability-relations can be ascertained partly by microscopic study and partly by experiment in the laboratory.

We must now inquire what help we can derive from the chemical analysis of rocks, from their texture, their geological environment, and their mineralogical composition, in the attempt to set up a genetic or physico-chemical classification of eruptive rocks.

THE USE OF CHEMICAL ANALYSIS IN CLASSIFICATION

It is often assumed that the bulk analysis of an eruptive rock indicates the composition of the magma from which that rock crystallized, save for the loss of an insignificant proportion of water and other fugitive constituents. Those who make this assumption close their eyes to such possibilities as the sinking or flotation of early-formed crystals in the magma; the separation of residual liquid by earth-pressure or gas-streaming; the marginal assimilation of country-rock; and hydrothermal reactions such as albitization. We have already discussed these processes and have found them to be so important as to justify the claim that the composition of an eruptive rock must often be very different from that of the magma that gave birth to it. If any further illustration of this truth is required let us turn to F. F. Grout's account of the differentiated sill at Pigeon Point, Minnesota. This sill is believed to have been injected as a homogeneous basaltic magma, yet from this magma there were formed both olivine-gabbro and a red rock of granitic composition. Chemical analysis of these rocks shows how greatly they differ from each other, and how both differ from the intermediate rock which perhaps corresponds more closely to the parent magma.

	Olivine gabbro	*Intermediate Rock*	*Red Rock (Granite)*
SiO_2	49·88	57·98	72·42
Al_2O_3	18·55	13·58	13·04
Fe_2O_3	2·06	3·11	·68
FeO	8·37	8·68	2·49
MgO	5·77	2·87	·58
CaO	9·70	2·01	·66
Na_2O	2·59	3·56	3·44
K_2O	·68	3·44	4·97
TiO_2	1·19	1·75	·40
Rest	1·31	2·93	1·65
	100·10	99·91	100·33

Even if future investigation should show that the red rock at Pigeon Point is slightly younger than the gabbro, like the red rock at Sudbury, that would not invalidate the conclusion drawn here, for the two rocks are so intimately associated that one can hardly doubt their common origin.

There are certain deep-seated rocks that are composed almost entirely of a single mineral such as olivine, hypersthene, or lime-rich plagioclase. No lavas of similar composition have ever been discovered, and the reason is simply that the melting points of the minerals mentioned are well above the normal range of magmatic temperature. It seems to be beyond dispute that these rocks were not formed from magmas of their own composition. There are even rocks composed of almost pure chromite, which has its melting point in the neighbourhood of 1850° C.; it is inconceivable that these rocks crystallized from a magma of their own composition.

Whenever one finds a porphyritic rock with insets so large and so numerous that they would cause the chemical composition of the rock to differ widely from the average composition of rocks containing the same association of minerals, then there is reason to suspect enrichment by settling or flotation of crystals. This applies to plagioclase-rich basalts such as the "Porphyritic Central" type in the Hebrides; to olivine-rich basalts such as the "chrysophyres" of Hawaii; to augite-rich rocks such as the "augitites" of Kenya; to leucite-rich rocks such as the "italite" of the Alban Hills; and to many orthoclase-rich porphyries. The bulk composition of any one of these rocks must be very different from that of the magma that gave birth to it.

But if the chemical composition of a rock is no reliable guide to the fixed constituents of a magma (and it is certainly none to the fugitive constituents), then it must follow that *chemical analysis does not provide a suitable basis for the genetic classification of rocks*. It is simply misleading to refer a rock to a particular "magma-type" on the basis of chemical analysis alone. Only the most direct and complete field-evidence can put us in a position to correlate a particular rock with a particular magma.

Not only does the bulk analysis of a rock tell us little of the magma from which that rock crystallized, but it fails even to indicate the mineralogical composition of the rock itself. This is convincingly shown by the two analyses given below. Except for a small difference in silica content the two analyses are remarkably similar, yet one of the analyzed rocks is a hornblendite containing more than ninety per cent of hornblende, and the other a dolerite composed of plagioclase, augite and olivine. The figures are taken from Washington's Tables.

	A	B
SiO_2	44·78	42·58
Al_2O_3	9·38	9·58
Fe_2O_3	4·51	4·97
(Fe,Mn)O	9·60	10·47
MgO	16·85	16·97
CaO	10·85	11·54
Na_2O	2·24	2·01
K_2O	·20	·54
TiO_2	·74	·94

On the other hand, if the mineralogical composition of a rock is acurately known, it is possible to arrive at its chemical composition by simple computation. As an illustration of the degree of success which is attainable in this matter, we shall quote an investigation by J. P. Iddings of the granite of Butte, Montana. The minerals present in the rock are quartz, orthoclase, plagioclase, biotite, hornblende, pyroxene, magnetite, pyrite; and their proportions were determined by the method of linear measurement of thin sections. The average composition of each mineral is known; multiplying this in each case by the quantity concerned, Iddings estimated the chemical composition of the granite to be as in Column I. An actual chemical analysis of the rock, by H. N. Stokes, is quoted in Column II.

	I *Calculated.*	II *Determined by Analysis.*
SiO_2	64·83	64·34
Al_2O_3	16·42	15·72
Fe_2O_3	1·48	1·62
FeO	2·39	2·94
MgO	2·01	2·17
CaO	4·35	4·24
Na_2O	2·97	2·76
K_2O	4·11	4·04
H_2O	·67	1·01
Rest	·70	·97
	99·88	99·81

This is an extremely good agreement; much better, it must be said, than could be expected in every case. It shows that it is possible to deduce the chemical composition of a rock from its mineralogical composition much more accurately than one can perform the reverse operation of deducing the mineralogical composition from a chemical analysis of the rock.

It will be instructive to see how the results of chemical analysis are employed in the two best-known chemical classifications of rocks. In the Norm system of Cross, Iddings, Pirsson and Washington, the oxides are recalculated into simple mineral molecules, which are either actual rock-forming minerals or else end-members of the great ionic substitution series. Thus plagioclase feldspar is represented in the norm by albite, anorthite, and minor quantities of orthoclase and nepheline; augite appears as hypersthene, diopside, acmite and corundum; hornblende is represented by the same molecules as augite; and biotite by orthoclase, olivine and corundum or hematite. What the norm system does, in effect, is to take the complex solid solutions apart and exhibit their hypothetical components separately. The norms of the two analyses A and B given on page 211 are as follows :—

	A	B
orthoclase	1·11	3·34
albite	11·00	3·14
anorthite	15·00	15·57
nepheline	4·26	7·38
diopside	31·04	31·10
olivine	28·76	29·42
magnetite	6·50	7·19
ilmenite	1·37	1·82
apatite	—	1·01

In the case of rock B the molecules of orthoclase, albite, and most of the anorthite that appear in the norm are actually present in the rock in the form of plagioclase. The normative nepheline corresponds perhaps to soda-anorthite in the plagioclase, since no nepheline actually appears in the rock. Normative diopside corresponds closely to augite in the rock (although the alumina which is always present in small quantity in augite has been allotted to anorthite in the calculation); and olivine and iron-ore are actual minerals of the rock. In this case the norm gives a fairly close representation of the actual mineralogical composition (called the *mode*) of the rock.

The position with regard to rock A is entirely different. The actual mineralogical composition of this rock is as follows: hornblende 91·3, olivine 3·6, iron-ore 5·1 per cent. It is clear that the norm entirely fails to express the true mineralogical character of this rock, and does not help us either to form a picture of the rock or to understand its genesis. Rock A (hornblendite) must have had a totally different cooling-history from rock B (olivine-dolerite), yet this supremely important aspect of the subject is ignored in the norm classification. To discuss rocks in terms of ultimate chemical composition without reference to their actual mineralogical development is to discuss a chemical system in terms of its components and ignore the phases; it is to reject physical chemistry in favour of mere arithmetic.

As a method of recording chemical data, of checking mineralogical data, and as a means of comparing cryptocrystalline and glassy rocks with holocrystalline ones, the norm system has a great and obvious value; but there its usefulness ends. It is incapable of throwing any light on the cooling-history of a rock.

Another chemical classification of rocks that is widely used is that of Paul Niggli. In this system the proportions by weight of the various oxides are converted into molecular numbers, as in the Norm system, but they are not subsequently built up again into mineral molecules. Instead, the various oxides are thrown into four groups as follows:—

$Al_2O_3 + Cr_2O_3 +$ rare earths;
$FeO + Fe_2O_3 + MnO + MgO$;
$CaO + BaO + SrO$;
$Na_2O + K_2O + Li_2O$;

The sum of the molecular numbers of all these oxides having been brought to 100, the four groups are represented respectively by the symbols *al*, *fm*, *c*, and *alk*. The molecular number of silica is then reduced in the same proportion as that of alumina and is represented by *si*. Further, the symbol *k* is used for the proportion of K_2O in *alk*, and *mg* for that of MgO in *fm*. The advantage of this procedure is that molecular ratios are retained and used as the basis of comparison instead of percentages by weight.

The two rocks that we have already discussed have the following Niggli symbols :—

	si	*al*	*fm*	*c*	*alk*	*k*	*mg*
Rock A ...	79·5	9·7	65	21	4·1	·05	69
Rock B ...	73	9·7	65	21	3·8	·16	·67

For the purpose of classification, Niggli (1) set up 67 arbitrary "magma-types" to which rocks might be compared; afterwards (2) this number was increased to 183. Searching among these magma-types for one to which rocks A and B may be referred, we find that no less than three come into question, namely :—

	si	*al*	*fm*	*c*	*alk*	*k*	*mg*
Hornblenditic type	80	15	60	20	5	(small)	0·6
Hornblende-peridotitic type ...	80	10	74	14	2		0·7
Pyroxenitic type	90	9	59	28	4		0·7

The value for *c* favours the first type, *al* favours the second, and *alk* the third; the remaining values are indecisive. Having brought us so far by exact quantitative methods, Niggli now abandons methods of precision in favour of bare opinion and leaves us to choose for ourselves. Neither from the "Niggli values" nor from the indicated "magma-type" do we learn anything definite about the composition of the magma, anything about its cooling-history, or anything about the actual mineralogical composition of the rock that is to be classified. In short, Niggli's system is no more successful than the norm system in enabling us to make any useful deductions about the genesis of the rock.

We must therefore conclude that the chemical analysis of rocks does not afford that harvest of "important general propositions" of which we are in search. This does not mean that the chemical composition of rocks has no value in classification. It is hardly possible to study cryptocrystalline and glassy rocks without the aid of chemical analysis, and even in the case of holocrystalline rocks the phases can seldom be completely determined without that aid. But it is evident that chemical analysis alone can never provide a basis for the genetic classification of rocks. Its real value is to supplement the evidence given by the microscope, and to investigate the composition of cryptocrystalline and glassy rocks.

THE USE OF TEXTURE IN CLASSIFICATION

The physical conditions that obtained during the crystallization of a magma are deducible to a certain extent from the texture and the geological environment of the rock. The temperature at the moment of intrusion, or at the beginning of crystallization, can often be judged by the "geological thermometer" method (Chapter IV). The viscosity of the magma and the rate of cooling leave their impress in the texture of the rock; and the order of crystallization is also recorded in the texture. The correct interpretation of texture is therefore a very important factor in classification.

Whitman Cross has shown that texture involves three distinct factors, which he defines as:

(1) Crystallinity, or the degree of crystallization;

(2) Granularity, or the magnitude of the crystals;

(3) Fabric, or the shape and arrangement of the crystals.

As regards crystallinity and granularity, it is well known that some rocks are wholly glassy while many rocks consist partly of crystals and partly of glass. The crystalline rocks differ among themselves as regards the magnitude of their crystals, some being composed of crystals from an inch to a foot in diameter, while others, which are just as completely crystallized, have individual crystals so small as to require a microscope for their recognition. In other cases still the groundmass of a rock appears to be structureless, yet its behaviour towards polarized light shows that it actually consists—in part at least—of sub-microscopic crystals.

Both field and laboratory evidence demonstrate that high crystallinity and high granularity are favoured by slow cooling and by the presence of fluxes, while deficient crystallinity and a low degree of granularity may be due either to rapid cooling or to the loss of fluxes, or to both causes combined. Rocks of coarsest grain are accordingly found in large intrusive masses, but also in smaller masses where fluxes were abundant, such as pegmatite veins. Fine grain is characteristic of small intrusions and marginal facies, and of lava flows, from which the loss of heat and of fluxes has been comparatively rapid. The glassy and partly glassy rocks are found only in lava flows, narrow dykes, and the selvages of dykes, where cooling and loss of fluxes occurred too rapidly for crystallization to set in.

Here we have generalizations of some importance, based on a set of obvious characters, and justifying the long-established separation of the *plutonic* rocks (which crystallized deep down in the crust, and have only reached the surface in consequence of prolonged erosion of overlying strata) from the *volcanic* rocks which were poured out on the surface. H. Rosenbusch went beyond this, and set up a third group of "dyke-rocks" (*Ganggesteine*) characterized by textures intermediate between those of the former groups. But since every known variety of texture, from the most coarsely crystalline to the completely glassy, is to be found among rocks that occur in the form of dykes, it is clear that this procedure cannot be followed without abandoning the correlation between texture and physical conditions which distinguishes the plutonic and volcanic groups. So long as we maintain, as Ferdinand Zirkel did, *two* textural groups only, then we can establish definite and important propositions concerning them; but for *purely petrographical purposes* (as distinct from geological purposes) there is no justification for the erection of a separate group of dyke-rocks which share the characteristics of the former groups but have no features peculiarly their own.

With regard, then, to crystallinity and granularity it is clearly desirable to distinguish two great groups of rocks, the first of which includes all deep-seated rocks and the more coarse-grained dyke-rocks, while the second embraces all microcrystalline, cryptocrystalline, and glassy rocks whether they occur in minor intrusions or in lava-flows. For geological purposes we may speak of these as the Plutonic and the Volcanic groups (*Tiefengesteine* and *Ergussgesteine* of Rosenbusch), but these names introduce a hypothesis of origin which is not always justified. Not all coarse-grained rocks are plutonic, nor all ill-crystallized rocks volcanic in the literal sense. Zirkel described the two groups in terms of texture alone, as

(a) gleichmässig-körnig (uniformly granular).

(b) porphyrisch und glasig (porphyritic and glassy).

This is the logical method. The rather cumbersome German adjectives may be replaced without any change of meaning by *eucrystalline* (well-crystallized) and *dyscrystalline* (ill-crystallized, either relatively or actually.).

A special importance has sometimes been assigned in classification to the *porphyritic* texture, for it has been supposed that the large crystals (insets) in porphyritic rocks were formed when the magma was deep down in the crust and the temperature and pressure were higher than at the time the groundmass crystallized. It is certainly true that lava flowing from the crater of a volcano sometimes contains ready-formed crystals, such as the large leucite crystals in some of the lavas of Vesuvius. There is no need, however, to suppose that the insets were formed under very different conditions from the smaller crystals which grew later, because the rate of growth of some minerals (including leucite) is rapid, and the early formed crystals, being few in number, are most favourably situated for attaining a large size. L. V. Pirsson has brought forward some strong field evidence against the idea of the deep-seated origin of insets, and N. L. Bowen has shown experimentally that perfectly continuous crystallization may give the appearance of crystallization in two generations. The first-formed crystals grow rapidly until a period is reached when the increasing viscosity of the magma, or the degree of undercooling, causes new centres of crystallization to appear in large numbers. When the whole mass has frozen there will be a few large insets and a multitude of small groundmass crystals, although crystallization has been continuous and the pressure has not changed. On the other hand, we have seen that some of the most striking examples of porphyritic texture have been formed by sinking or flotation of early-formed crystals, which are therefore strangers to the environment in which we see them.

The real importance of the porphyritic texture, in relation to classification, appears in the interpretation of the *order of crystallization*. It is in this connection, too, that the study of *fabric*, or the shape and arrangement of the crystals in rocks, acquires its chief significance. We discussed the order of crystallization at length in Chapter VII, and found that the interpretation of the sequence of events in any rock is often a matter of grave doubt and difficulty. There are virtually only three rules for our guidance, which we may summarize as the rule of enclosure, the rule of idiomorphism, and the rule of relative sizes; and all of them are subject to exceptions.

It is a simple matter to determine the order of crystallization in an artificial melt. The procedure is to make a mixture of the desired composition and heat it to such a temperature that it is entirely liquid; then cool it to a prearranged temperature, hold it there for some time, and quench it suddenly. The cold solid, consisting of crystals and glass, is studied under the microscope and the crystalline phases are identified by their optical properties. When this procedure is repeated at various temperatures it becomes possible to fix the precise temperature at which each phase begins to form, and the order in which the various phases succeed one another.

Nature has made a few such quenching experiments for us, but far too few. Rocks composed of glass with only a few crystals are fairly common in the highly siliceous group, but scarcer and scarcer as we pass to groups of lower silica content. A comparison of the chemical composition of 124 glassy or partly glassy rocks cited in H. S. Washington's *Tables* has shown that 85 per cent of these are leucocratic, 88 per cent are oversaturated with silica, and only 11 per cent show a deficiency of silica. Furthermore, the study of natural rock-glasses is complicated by the products of gas-reactions, which yield unstable phases such as tridymite and cristobalite, and also by devitrification of the glass at atmospheric temperature.

As long ago as 1887, A. Lagorio made a study of the composition of the glassy base of a large number of lavas and dyke-rocks. He was able to establish the approximate rule that in rocks of moderate acidity (about 55-56 per cent of SiO_2) the glass base is more acid than the entire rock, whereas in very acid rocks the base is less acid than the rock. In other words, quartz crystallizes early from highly siliceous magmas and later from those of lower acidity. J. H. L. Vogt gives similar data. He finds that in dacites, trachytes, rhyolites and quartz-porphyries with not more than 72 per cent. of total silica, the glass base is invariably more acid than the whole rock. When the percentage of silica is about 73 to 75, the glass base has about the same acidity as the whole rock. A few very acid porphyries, with more than 75 per cent of total silica, have more quartz among the insets than in the ground mass, indicating early crystallization of quartz.

It would seem to be established by the observations of Lagorio and Vogt that the boundary between the quartz field and the field of soda-potash feldspar in these acid lavas lies in the neighbourhood of 74 per cent of silica, varying a little according to the composition of the feldspar. It is easy to test this conclusion. For instance, a porphyritic glass from Fifeshire, Scotland, which was examined by the writer and analyzed by one of his students, shows a few insets of sodic plagioclase but none of quartz. The deduction is that the composition of the magma lay on the feldspar side of the boundary. Analysis showed $SiO_2=68{\cdot}26$ per cent.

Another porphyry, from Robertson, South Africa, has large insets of potassic feldspar and smaller ones of quartz. Apparently feldspar began to crystallize before quartz, so the initial composition of the magma must have lain in the feldspar field. An unpublished analysis by D. L. Scholtz showed $SiO_2=69{\cdot}47$ per cent.

A third porphyry, from Saldanha Bay, South Africa, has insets of feldspar and quartz which are almost equal in number and size. This suggests simultaneous crystallization of the two phases from a magma with about 74 per cent of silica. Analysis showed $SiO_2=74{\cdot}82$ per cent.

Nevertheless, the validity of such deductions about the order of crystallization is none too certain. Minerals differ in rate of growth, so the observation that the feldspar insets are bigger than the quartz insets is not a conclusive demonstration that feldspar began its crystallization before quartz. H. A. Powers records that out of 15 acid lavas, so siliceous that their composition would take them into the "quartz field" of Vogt, twelve contain varying amounts of both quartz and feldspar insets; but in spite of Vogt's rule, three of the rocks hold feldspar insets without any of quartz.

Passing from acid to basic lavas, we find that T. F. W. Barth has tried to establish the position of the cotectic surface between the plagioclase field and the pyroxene field in basalts, and that he has proposed a formula connecting the order of crystallization with the chemical composition of the lava. If the correctness of Barth's formula should be established, it might enable us to distinguish between basalts formed by normal crystallization in the plagioclase field or the pyroxene field, and basalts which have undergone enrichment with plagioclase or pyroxene by gravitational differentiation.

Deductions drawn from the observation of the order of crystallization in lavas and porphyries are not necessarily applicable to coarse-grained rocks. It is well known that quartz, although it crystallizes early from many acid lavas, is generally one of the last phases to crystallize in granites, Idiomorphic crystals of quartz which may have crystallized at a relatively early stage are found especially in the "Rapakiwi" type of granite. Many of these rocks contain more than 74 per cent of silica, but in some of them the proportion of silica falls to 68 per cent and even lower.

It should now be apparent that the order of crystallization is a character of first-rate importance in the interpretation of the cooling-history of a magma, and for that reason most deserving of recognition in the classification of rocks. But it is equally apparent that we are not yet in a position to employ this character confidently and consistently. It is perhaps in this direction more than any other that the classification of rocks is likely to progress in the future.

THE USE OF MINERALOGICAL COMPOSITION IN CLASSIFICATION

We have seen that the chemical analysis of a rock can tell us little or nothing of its cooling-history. The genesis of a rock is not a question of what might have happened but of what *did* happen. Something can indeed be learned from the study of texture, but then texture has no meaning apart from the minerals that show it. Thus there is no escape from the conclusion that *the actual mineralogical development is the only possible basis for the genetic classification of rocks.* Of course rocks may be studied from other points of view than that of genesis—as structural elements of the lithosphere, as storehouses of the chemical elements, or just as building stones—and for every viewpoint there is an appropriate classification; but scientific petrology has no concern with these. The petrologist sees eruptive rocks as the end-products of a long series of chemical reactions and physical transformations, and he tries to elucidate these and express them in a classification which is appropriate to his point of view. The keynote of this classification must be physical chemistry.

Although the classification of a rock must be based on the minerals actually present in it, yet it is not to be supposed that all minerals are equally important for this purpose, or even that the importance of a mineral is in proportion to its abundance in the rock. The importance of any mineral for the purpose of classification depends upon the information we can extract from it about the cooling-history of the system. A mineral with a wide range of stability, such as orthoclase or an intermediate plagioclase, tells us very little about this matter; whereas some of the less common minerals such as leucite, olivine and tridymite are highly informative even when they are present only in trifling quantity. It is certain that too much importance has been attached in the past to the very stable feldspars, and too little to certain other minerals which are more significant of the cooling-history of the magma.

In our study of the geological thermometer we found that quartz, tridymite, leucite, hornblende, olivine, acmite and biotite all tell us something of the temperature of the magma during the crystallization period; and even of fluctuations in the temperature, as when hornblende or biotite is formed and afterwards undergoes dissociation.

From the study of assimilation we learn that melanite, vesuvianite, scapolite, melilite, cancrinite and perhaps nepheline too, are indicative of the contamination of a magma by assimilation of limestone. Similarly corundum (sometimes), andalusite and cordierite indicate contamination by assimilation of aluminous sediments, and so perhaps does hypersthene. It has been said that " if there were not so many slates there would not be so many norites."

We have learned from experiment in the crucible that some minerals are pyrogenetic—that is, they will crystallize readily from anhydrous silicate melts without any flux—whereas others can only be induced to crystallize in the presence of water or some other substance of low melting point which acts as a flux. Many of the hydatogenetic minerals have a constant proportion of hydroxyl groups in their atomic framework, which points to a relatively high concentration of hydroxyl ions in the magma at the time these minerals crystallized. Similarly those minerals that contain fluorine, boron, beryllium, lithium, zirconium, cerium and other rare elements must have crystallized out of a residual magma or solution in which these scarcer elements had become concentrated towards the end of a long crystallization period.

We know, too, that the rock-forming minerals fall into two groups on the basis of their compatibility or incompatibility with excess of silica under magmatic conditions. Thus the presence of unsaturated minerals such as leucite, nepheline or melilite affords information which is indispensable for the understanding of the chemical reactions that went on in the magma.

It follows that from the mineralogical composition of a rock we can deduce something of the temperature of the magma at the beginning or end of the crystallization period; something of the concentration of the fugitive constituents during the same period; something of the ratio of silica to metallic oxides in the magma; also something of the fluctuations in each of these factors; and something of the extent to which the magma has been contaminated by reaction with its walls.

When we have information of this character at our disposal, why should we continue to classify rocks on the basis of an arbitrary ratio of orthoclase to plagioclase, or of albite to anorthite? Whether the plagioclase in a rock contains 45 parts of albite on 55 parts of albite tells us little of the genesis of the rock or of its history. It is much more important to know whether the rock contains augite or hornblende, and whether olivine or feldspathoid is present or not.

It is often said that "there are no natural boundaries between one igneous rock and another." This astonishing misstatement is made even by people who must be fully aware of the incompatibilities between quartz and feldspathoids, quartz and olivine, enstatite and feldspathoids, and other incompatibilities which we discussed in Chapter VIII. What better or more natural boundaries could there be than these mutual exclusions, which are the visible expression of phase-boundaries in the magma-system?

The nature and the significance of a phase-boundary are made clear by *Fig.* 31, which presents diagrammatically the conditions of equilibrium in the system MgO—SiO$_2$ from the data of Bowen and Andersen. The liquidus of this system shows four interruptions, at about 38, 43, 60 and 65 per cent of silica. The first of these breaks corresponds to the eutectic between periclase and forsterite, the fourth to the cristobalite-clinoenstatite eutectic. The break at 43 per cent of silica is a phase-boundary between enstatite and periclase, for on one side of this boundary clinoenstatite cannot be formed and on the other side periclase is not a possible phase. Similarly the break at 60 per cent of silica indicates a phase-boundary between forsterite and cristobalite. These two phase-boundaries divide the field completely into three natural regions, within

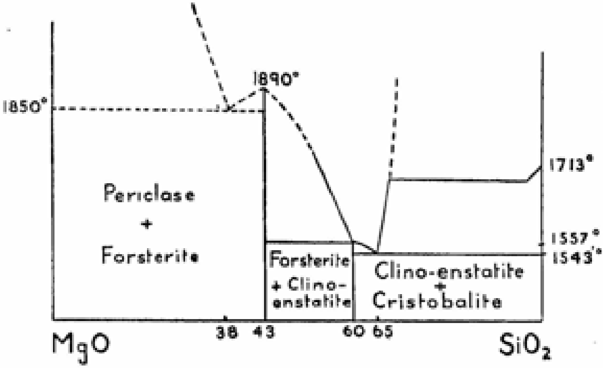

FIG. 31. Phase boundaries and eutectic points in the system MgO—SiO$_2$; after Bowen and Andersen, also Bowen and Schairer.

each of which certain phases are possible and certain other phases impossible when the system is in equilibrium.

The eutectic points also divide the field into three regions, but the middle region (from 38 to 65 per cent of silica) does not have a definite composition in terms of minerals, since periclase, forsterite, enstatite and cristobalite are all possible in some portion of this region. Thus for the purpose of classification the phase-boundaries are more useful than the eutectic points since they enable us to effect a precise correlation between chemical composition and mineralogical development.

In rocks there are probably no eutectics, but there are certain phase-boundaries which form natural dividing lines. A petrology based on the stability or instability of particular phases in particular associations offers a prospect of "important general propositions" connecting mineralogy and chemistry, such as no other kind of classification has yet been able to supply. We may call this "phase petrology." The map of the Cortlandt complex (*Fig*. 26, p. 167) illustrates the use of "phase petrology" in place of the old-fashioned "species petrology."

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The purpose of classification is not to set forth final and indisputable truths but rather to afford stepping-stones toward better understanding.—L. C. GRATON.

CHAPTER XIV

A SYSTEM OF PETROGRAPHY

OUR immediate task, arising out of the discussion in the last chapter, must be to show how rocks can be grouped in such a way that each group will convey the greatest possible amount of information about the physical and chemical history of the magma-system. We must distinguish between characters of fundamental importance from the standpoint of physical chemistry, and characters which have no more than a descriptive value; and we must subordinate the latter to the former. We shall proceed therefore by discussing first the primary or physico-chemical basis of classification, and afterwards the secondary or descriptive basis.

THE PRIMARY OR PHYSICO-CHEMICAL BASIS OF CLASSIFICATION

(1)

The first step should be, as in Zirkel's day, the separation of eucrystalline from dyscrystalline rocks. There is no natural dividing-line between these two groups, but an artificial boundary may be drawn at the limit of visibility with the unaided eye or with a simple hand-lens. Any rock in which the grains of the essential minerals are too small to be seen individually without the help of the microscope is dyscrystalline. In the case of a porphyritic rock the criterion is to be applied to the groundmass. The smallest quantity of a glassy base puts the rock in the dyscrystalline division.

The following general propositions may be stated regarding these two great groups of rocks.

Eucrystalline rocks. The pressure upon the magma was high, but the temperature at the time of intrusion, as judged by the geological thermometer method, was generally less than 800° C. The rate of cooling was slow, hence the crystallization-period was prolonged. Loss of dissolved gases was also slow, so the liquid phase, as crystallization advanced, was progressively enriched with fugitive constituents; thus crystallization was complete and it was concluded at a comparatively low temperature. In the final stages the concentration of water was so high that the period of magmatic crystallization was sometimes followed by a period of hydrothermal (deuteric) crystallization and reaction. Free silica takes the form of quartz, never that of tridymite, and it always crystallizes late, the crystals enclosing minute bubbles of water and gas.

Dyscrystalline rocks. Crystallization took place under low pressure, but the temperature of intrusion or extrusion was high, often exceeding 900° C. Loss of heat to the atmosphere or to surrounding rocks took place rapidly, and on account of the reduction of pressure there was a rapid loss of gases; consequently crystallization often failed to be completed and the last portion of magma solidified as glass. Hydroxyl-bearing silicates such as mica and tourmaline are scarce; and inclusions within crystals are mostly of glass or stony matter. Quartz is less common than in eucrystalline rocks and it has often crystallized at an early stage; tridymite sometimes accompanies or replaces it. Pyroxenes are commoner than amphiboles.

In the pages that follow the letter X will be used as a shorthand symbol for eucrystalline, and D for dyscrystalline.

(2)

The phase-boundaries between free silica (quartz, tridymite, silica-glass) and the various unsaturated minerals that we discussed in Chapter VIII are natural boundaries of great significance, and they lead at once to the following classification of rocks.

Class I—*Oversaturated rocks* or those which contain free silica of primary origin. Symbol O.

Class II—*Saturated rocks* or those which contain neither free silica nor any unsaturated mineral. Symbol S.

Class III—*Undersaturated rocks* or those which consist either wholly or in part of unsaturated minerals. This class may be subdivided as follows:—

(a) *Non-feldspathoidal division*. The alkalis are saturated and the deficiency of silica affects the dark minerals only. Symbol V.

(b) *Feldspathoidal division.* The alkalis are partly or wholly unsaturated. The dark minerals may be either saturated or unsaturated. In the former case the symbol is U, in the latter, W.

This classification makes no great departure from the historic system of Ferdinand Zirkel, in which the quartz-bearing rocks were nominally separated from the feldspathoidal rocks and the olivine-free from the olivine-bearing rocks. Zirkel was not quite consistent in this matter, for he described under the heading of "syenite" certain rocks that contain minor amounts of quartz, nepheline, sodalite or olivine; and under "trachyte" some that hold tridymite and others containing a little leucite, sodalite, haüyne or olivine.[1] The proposed classification differs more widely from that of Rosenbusch, who either ignored chemical considerations altogether or subordinated them to his conception of *Gesteinsfamilien* and the *Ganggefolgschaft.* For instance, the syenite, monzonite, essexite and gabbro families of Rosenbusch's classification all contain both oversaturated and undersaturated members. It was only when the proportion of quartz or feldspathoid became quite large that Rosenbusch would change the name of the rock; that is to say, he preferred an indefinite and chemically meaningless boundary to a precise boundary possessing physico-chemical significance.

Criticism of the proposed basis of classification has come mainly from petrologists of the Rosenbusch school, who claim that it "obscures the natural affinities" between rocks. The phantasy of petrological affinity dies hard! When these "natural affinities" can be defined as clearly and unequivocally as the *natural antipathies* that form the basis of the proposed classification, and can be shown to have an equal significance in terms of physical chemistry, then this criticism will deserve consideration, but not till then.

A. Lacroix has said that "the production of a small quantity of quartz or of a feldspathoid has no real importance from the point of view of classification." G. W. Tyrrell, too, thinks that "little notice should be taken of small quantities

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[1] In the report on the 40th parallel, Zirkel described a rock from Slater's Fork, in the Elkhead Mts., as trachyte, although he had himself identified nepheline in it.

of quartz, olivine, or feldspathoids." Why should one take little notice of facts that bear directly on the cooling-history of the magma and the course of differentiation? The quartz-feldspathoid and quartz-olivine boundaries are *natural dividing-lines*, and as such they have a scientific importance which artificial boundaries can never have.

Both Lacroix and Niggli have objected to the separation of quartz-bearing from olivine-bearing rocks on the ground that some rocks hold both minerals. It is strange that this criticism should be brought up again and again, for it has been shown repeatedly that the association is not a stable one and that its appearance is a mere accident of cooling. That the olivine-silica reaction was not completed in a particular instance is a *local exception* to a general rule. The logical procedure, in such a case, is first to establish the rule and then to deal with the exceptions. The existence of exceptions should not be permitted to blind one to the rule. In dealing with a rock that contains both quartz and olivine, one should estimate the (usually small) proportion of each and decide which is in excess (in round numbers, two parts by volume of olivine will neutralize one part by volume of quartz). This will decide the primary classification of the rock; its abnormal character may be indicated in the process of subdivision.

(3)

After silica, the component next in importance is alumina. In the feldspars the molecular ratio of alumina to soda, potash and lime combined is exactly 1 : 1, and it is so in nepheline and leucite too; consequently any excess or deficiency of alumina with respect to this ratio must show itself in the character of the dark minerals. This consideration leads us directly to a fourfold grouping of rocks, as follows.

I. The *peraluminous* type (Symbol p). In these rocks the molecular proportion of alumina exceeds the molecular proportions of soda, potash and lime combined. The excess of alumina goes into muscovite, biotite, corundum, tourmaline, topaz, or an iron-manganese garnet (almandine or spessartite). These are typical products of low-temperature, hydrous, generally acid magmas and of the later stages in the magmatic reaction-series.

II. The *metaluminous* type (Symbol m). The proportion of alumina exceeds that of soda and potash combined, but is generally less than that of soda, potash and lime combined. Nevertheless, some of the alumina enters the dark minerals, forming either a moderately aluminous mineral such as hornblende, epidote or melilite, or else an association of an aluminous with a non-aluminous mineral, such as biotite with pyroxene, hornblende with olivine. The physical conditions during crystallization were intermediate between those of the preceding and succeeding types.

III. The *subaluminous* type (Symbol s). There is little or no excess of alumina over that required to form feldspars and feldspathoids. The dark silicates are olivine and orthopyroxene, which are strictly non-aluminous, and diopside and augite, in which a small proportion of silicon atoms may be replaced by aluminium. These are all pyrogenetic minerals and are typical of hot, relatively anhydrous magmas.

IV. The *peralkaline* type (Symbol k). The molecular proportion of alumina is less than that of soda and potash combined. The characteristic dark minerals are the soda-pyroxenes and soda-amphiboles, ænigmatite, eudialyte, and other species in which the place of alumina is taken by ferric oxide, zirconia, or titania. Ferric oxide and titania also take the place of alumina in melanite and schorlomite, two lime-rich species which are only found in rocks of peralkaline character. These are all low-temperature minerals formed in the last stage of crystallization of soda-rich magmas, when the residual liquor was strongly alkaline in character.

The four rock-types that we have just distinguished mark successive stages in the cooling-history of a magma. The crystallization of a system containing alumina often begins with the production of olivine or an almost non-aluminous pyroxene, which at a later stage reacts with the remaining liquid and is gradually transformed into hornblende or even further into biotite. These four minerals constitute a reaction-series in the sense of N. L. Bowen, and the corresponding rock-types may be called reaction-types. In a less aluminous magma a different reaction may take place and the original pyroxene may be replaced, or at least mantled, by ægirine or a soda-amphibole.

		p	m	s	k
O	X	1	3	5	7
	D	2	4	6	8
S	X	9	11	13	15
	D	10	12	14	16
V	X	17	19	21	23
	D	18	20	22	24
U	X	25	27	29	31
	D	26	28	30	32
W	X	33	35	37	39
	D	34	36	38	40

TABLE A. THE PRIMARY BASIS OF CLASSIFICATION.

A symbol such as XOp or DWs indicates some of the most outstanding characteristics of the rock described, but it does not go far enough. In the group XOp the feldspar may be all orthoclase or all plagioclase. The symbol XSs represents anything from a potash-syenite to an anorthite-rock or a pyroxenite. DWm may contain either nepheline or leucite and either olivine or melilite. Further subdivision is clearly needed, and it seems desirable that it should have a quantitative basis. But here Nature declines to help us. Apart from the exclusions discussed on earlier pages, and the very doubtful occurrence of "anchi-eutectic" rocks, there are no natural limits to the proportion of any mineral in any rock. It is true that quartz-orthoclase rocks typically hold only a small proportion of ferromagnesian minerals and that the plagioclase rocks are mostly rich in ferromagnesians; but to this statement there are many exceptions, among which the case of the anorthosites is outstanding. If we wish to establish quantitative mineralogical subdivisions, we shall have to do it on an arbitrary basis.

Chemical analysis, too, fails to establish any natural dividing lines. H. S. Washington directed attention to a sympathetic relation that obtains between magnesium and potassium, on the one hand, and iron and sodium on the other hand. (See *Fig.* 32.) But there are rocks rich in potassium yet almost without magnesium, and rocks rich in sodium but almost free from iron; so this relation is not definite enough for the purpose that we have in view. Any subdivisions that we establish upon the proportions of the various oxides must be arbitrary.

THE SECONDARY OR DESCRIPTIVE BASIS OF CLASSIFICATION

(1)

From the genetic standpoint the mere presence of a critical mineral such as nepheline or olivine, which marks the crossing of a phase-boundary, or tourmaline or ægirine, which marks a particular reaction-type, is more important than the actual abundance of that mineral; nevertheless it is desirable, in order that one may form some sort of picture of the rock, to have a statement of the proportions of the various minerals in it. A preliminary statement of this kind may usefully take the form of the *ratio of heavy to light minerals in the rock*. All minerals of density less than 2·8 are free from magnesium and iron and are consequently pale in colour, whereas the majority of minerals denser than 2·8 are distinctly and often strongly coloured. The ratio of heavy to light minerals is therefore nearly the same thing as the ratio of dark-coloured to light-coloured minerals, and since it is this which mainly determines the colour of a rock it may be called the *colour ratio* or *colour index* of the rock. (The colour ratio of a rock is *x per cent;* its colour index is simply *x*. The latter term has been Germanized in the form *Farbzahl;* in French it has become *indice de coloration*).

The light minerals (or those so regarded by the writer, as well as by Lacroix and Tröger) are quartz, tridymite, all feldspars, nepheline, cancrinite, the sodalite group, leucite and analcime. But Johannsen and Niggli have interpreted the adjective "light" more literally, and they include muscovite, scapolite, melilite, topaz, corundum, apatite, andalusite, sillimanite, and fluorspar among the light minerals. If colour alone is to be decisive, then enstatite, olivine, phlogopite, and zircon might all have been added to this list. It seems better to base the distinction of the two groups upon the definite character of density than upon the very indefinite one of colour.

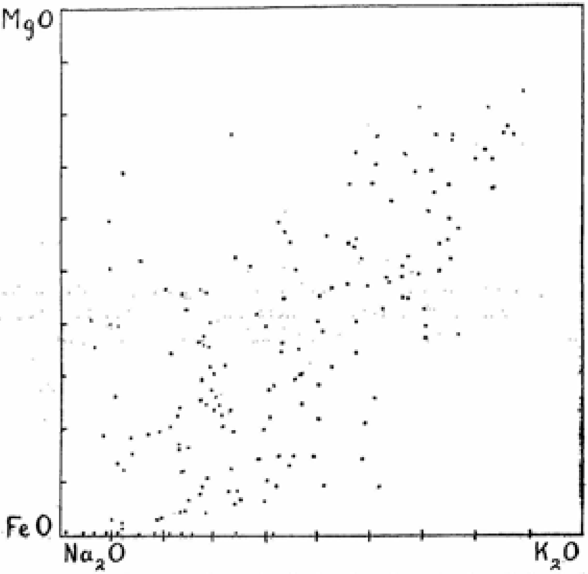

FIG. 32. A plotting of nearly two hundred rock analyses, to show that potash and magnesia on the one hand, and soda and iron oxide on the other, tend to vary together. (H. S. Washington.)

There are no natural boundaries in respect of colour index, so we must fix our boundaries arbitrarily yet with some regard to the tendencies that we observe in nature. For instance, the very abundant granites, granodiorites, tonalites and anorthosites hardly ever carry as much as thirty per cent. of dark minerals, whereas the abundant gabbros, norites, diabases and basalts nearly always carry more than 30 per cent. of these minerals; so a boundary at about 30 per cent (by volume, since it is easier to judge volume than weight) is suggested. Another boundary at 90 per cent cuts off the rocks composed almost wholly of dark minerals (pyroxenites, hornblendites, peridotites) from the rest; and between these limits we may reasonably put another boundary at 60 per cent by volume of dark minerals. The four facies thus distinguished may be called *leucocratic* (colour index 0-30), *mesotype*[1] (colour index 30-60), *melanocratic* (colour index 60-90), and *hypermelanic* (colour index 90-100) respectively. Similar limits have been suggested by Lincoln (33, 67), Lacroix (35, 65, 95), and Niggli (37½, 62½, 87½).

When one writes a symbol to indicate the composition of a rock, it should generally be sufficient to express the leucocratic, mesotype, melanocratic or hypermelanic character by the addition of the capital letters L, M′, M″, or H, as in XSmM′ or XVsH. Alternatively, one may append the exact colour index as in XOpβ(15). The word "melane" (borrowed from biology) is a convenient term for the dark minerals in rocks; thus we may say that a rock contains 15 per cent of melanes. The adjective "melanic" covers both mesotype and melanocratic and saves many syllables. The corresponding symbol M includes both M′ and M″.

(2)

The great majority of rocks contain some kind of feldspar, and this is far the largest constituent of most of them. On this account the composition of the feldspar has been a factor in every mineralogical classification of the past. The usual practice has been to separate the rocks in which alkali-feldspar predominates from those in which plagioclase is dominant,

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[1] The form *mesocratic* is used by some writers, but should be avoided as it has a nonsensical meaning. The middle cannot predominate. Scientific writers who are always ready to jeer when literary men misuse scientific terms should be specially careful not to misuse language.

and to make a third division for rocks in which orthoclase and plagioclase are about equal.

But since "orthoclase" or "alkali-feldspar" includes varieties that are rich in soda, and "plagioclase" means anything from pure soda-feldspar to pure lime-feldspar, the practice referred to is not good chemistry; and since soda appears in both terms of the ratio, it is not even good arithmetic. The weakness of this method of classifying rocks is shown by the contradictions to which it gives rise. For instance, a lava of San Mateo Mountain, New Mexico, which was named "andesite" by J. S. Diller, has the following composition:—

quartz	5·5%
orthoclase molecules	30·0
albite molecules	54·0
anorthite molecules	2·2
melanes	8·8

This rock carries more plagioclase than orthoclase, so the use of the name andesite is justified by current practice; but the proportion of potash feldspar is more than thirteen times that of anorthite.

In H. H. Robinson's admirable description of the lavas of San Francisco Mountain, Arizona, the names dacite and latite are employed. According to the accepted definitions of these names, latite contains roughly equal quantities of orthoclase and plagioclase, while dacite has plagioclase in considerable excess of orthoclase. Yet the chemical analyses of these lavas show that, in the dacites described by Robinson, orthoclase molecules are from one and a quarter to two and one-half times as plentiful as anorthite molecules; while in the latite the reverse is the case.

These examples show that the statement of the ratio of plagioclase to orthoclase which is conveyed by such names as latite, dacite, andesite, *as these names as currently defined*, is worthless as a guide to the relative proporitions of potash, soda and lime in the feldspars. If much albite is present, plagioclase will exceed orthoclase even when the proportion of anorthite is quite small; and if little albite is present, orthoclase may exceed plagioclase although the rock actually contains more anorthite than in the last case. If there is a great excess of albite over anorthite, then the plagioclase approaches the composition of pure albite; in this case it is the practice of those petrographers who follow Rosenbusch to transfer this highly sodic plagioclase to the other side of the equation to swell the total of "alkali-feldspar."

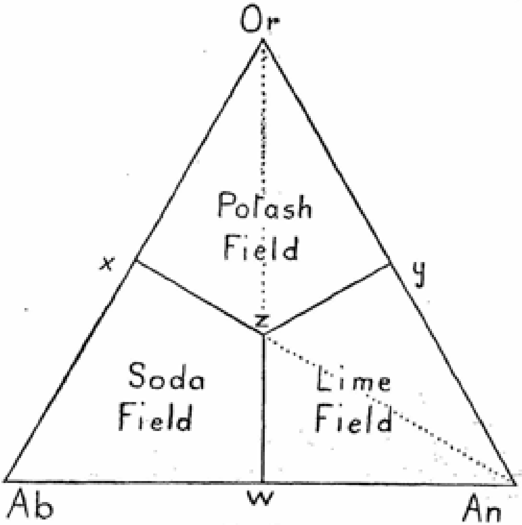

Fig. 33.

The complications which arise from the use of the indefinite terms "orthoclase," "alkali-feldspar," and "plagioclase" can only be avoided by discussing the feldspars in terms of the *pure components* $KAlSi_3O_8$ (Or), $NaAlSi_3O_8$ (Ab), and $CaAl_2Si_2O_8$ (An). In *Fig.* 33 the three corners of the triangular diagram represent Or, Ab, and An respectively. Now the Or-Ab-An field can be divided up in many ways, but the way that immediately and strongly suggests itself is to demarcate a potash-field (Or-xzy), a soda-field (Ab-xzw) and a lime-field (An-yzw), all of them equal in area. Every rock that is represented by a point in the potash field contains more orthoclase than either albite or anorthite, though not necessarily more than albite and anorthite combined. Similar statements can be made for the soda-field and for the lime-field.

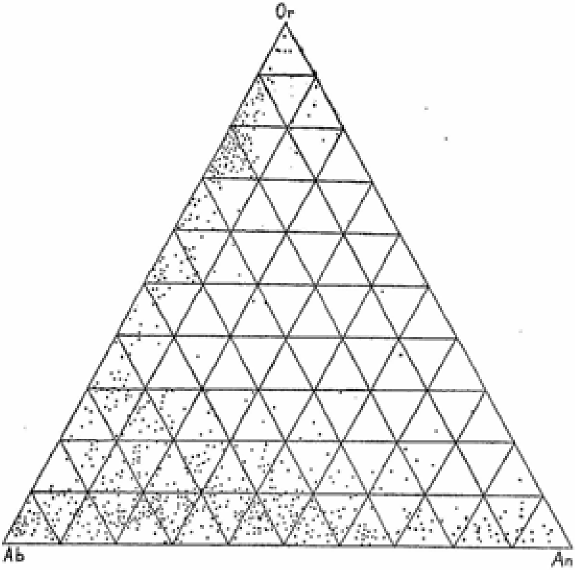

FIG. 34. The composition of 964 natural feldspars expressed in terms of Or, Ab, and An. (After H. L. Alling.)

It is well known that potash-lime feldspars are not found in nature, and that mixtures of orthoclase and anorthite are always diluted by much albite. This means that the area Or-z-An in *Fig*. 33 is almost unrepresented, and that the potash and lime fields are each reduced in practice to about half their theoretical area. The soda-field, however, is fully represented, for rocks are known which carry much albite and orthoclase with almost no anorthite, and others that have albite and anorthite with only a minimal quantity of orthoclase; there are also many rocks in which all three components are present in considerable amount. (In further illustration of this point see *Fig.* 34, which gives the composition of over 900 natural feldspars, as computed by H. L. Alling.) It will be advantageous, therefore, to subdivide the large sodic field into a soda-potash section and a soda-lime section, as in *Fig.* 35.

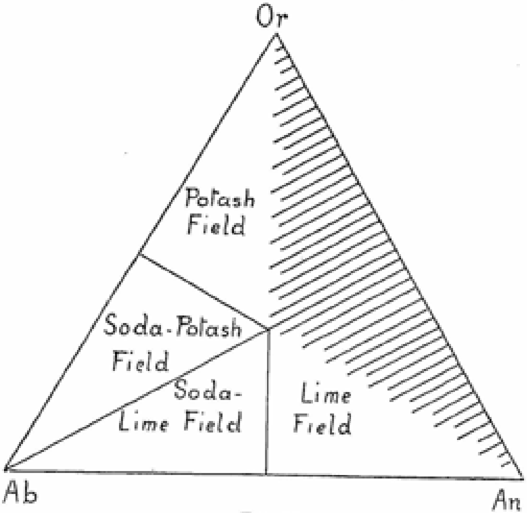

FIG. 35.

We have now divided the useful part of the triangle into four approximately equal fields, which are characterized mineralogically by potash-soda feldspar, soda-potash feldspar, soda-lime feldspar, and lime-soda feldspar respectively. The fields are defined in quantitative terms as follows:—

Or>An { 1. Or>Ab (potash-soda field). Symbol α
Or>An { 2. Or<Ab (soda-potash field). Symbol β
Or<An { 3. Ab>An (soda-lime field). Symbol γ
Or<An { 4. Ab<An (lime-soda field). Symbol δ

In the case of rocks containing leucite, Or is replaced by Or + Lc.

To classify rocks on the above basis is not much more difficult than under any existing system, for all good petrographic descriptions include either optical or chemical determination of the composition of the feldspars present. The first step will be to compare orthoclase to anorthite; in most rocks there will be a clear excess of one or the other, and as soon as that has been ascertained, a provisional name can be given to the rock. The ratio of orthoclase to albite will also be afforded by mere inspection in some cases; but in doubtful instances it will be necessary to undertake a closer investigation of the composition of the feldspar, using either gravitational, optical, or chemical methods, all of which are already part of the routine of petrography.

A. Johannsen has criticized the foregoing proposal on the ground that it introduces a calculated or average feldspar in place of the kind of feldspar actually present in the rock. It would, for instance, put a rock containing anorthoclase and a little labradorite in the same compartment as one with andesine and a less sodic orthoclase. But the customary method of using the feldspars in classification may lead to graver complications, as we showed on p. 236. In dealing with zoned plagioclase it has always been the practice of petrographers to estimate the average composition of the crystals, and Johannsen himself uses an average in the case of microperthite. Again, in the case of hypocrystalline and glassy rocks it is inevitable that we should discuss their composition in terms of theoretical feldspar. It is thought that the method proposed above leads to a more consistent and precise use of the feldspar factor in classification than has been customary hitherto.

We may combine the four main divisions based on feldspar with the four based on colour-index, as in Table B.

In the leucocratic and mesotype divisions it is necessary to recognize at least four subdivisions based on the kind of feldspar, but in the melanocratic division two subdivisions should be sufficient. In the hypermelanic rocks feldspar is only an accessory mineral and it should not be allowed a prominent place in classification. We may reckon, therefore, with a

	α	β	γ	δ
Leucocratic (L)	1	2	3	4
Mesotype (M′)	5	6	7	8
Melanocratic (M″)	9		10	
Hypermelanic (H)	11			

Table B. The Secondary or Descriptive Basis of Classification.

framework of eleven compartments for our secondary basis of classification; and these, when combined with the forty compartments of the primary scheme, give us ample accommodation for all eruptive rocks. The arithmetical product is 440, but in practice this number would be reduced to about 300 compartments.

The Question of Further Subdivision

It may be thought that the classification proposed above goes far enough; yet within certain groups of rocks it is clear that still further subdivision is necessary. In the undersaturated class, for instance, the leucite rocks must be separated from the nepheline rocks and the feldspathic rocks from the non-feldspathic ones. Such requirements as these must be met by partition of the appropriate compartments of the classification; but in the writer's opinion it is not advisable to attempt any further subdivision of the system *as a whole.* With two grades of crystallinity, five degrees of silication, four reaction-types, four divisions based on the colour-index, and four on the composition of the feldspar, we have already made it possible to indicate the salient mineralogical and chemical characters of a rock and something of its cooling-history. If further subdivision is needed within a particular rock-group, the matter may be dealt with when that group comes up for discussion.

In certain systems of classification, the relative proportions of the principal mineral-groups are used to establish arbitrary numerical boundaries. Thus, in the quantitative mineralogical classification of A. Johannsen there are four classes based upon the ratio of light-coloured to dark-coloured minerals (see back, p. 234), with limits at five, fifty and ninety-five per cent; within each class there are four orders based (in the first three classes) upon the composition of the plagioclase and (in the fourth class) upon the ratio of iron-ores to silicates. The orders in the first three classes are further subdivided into families according to the ratio of quartz to feldspar or feldspar to feldspathoid; in the fourth class the families are determined by the numerical relations between olivine, pyroxene, amphibole and biotite. This system gives a total of 364 "families" of rocks.

Paul Niggli has produced an amended version of Johannsen's system which some regard as an improvement on the original. In place of the dividing ratios of 5, 50, and 95 per cent which Johannsen used to define the classes, orders and families in his system, Niggli introduced the ratios that were used by the authors of the Norm system, namely 7 : 1, 5 : 3, 3 : 5, 1 : 7; thus obtaining a more equal division of the field. Niggli also differs from Johannsen on the question of what is to be understood by "alkali-feldspar," the point at issue being the same that we referred to on p. 236. The twelve "orders" of rocks in Niggli's system are just the larger groups of the Rosenbusch system, with arithmetical limits.

We may say of the systems of Johannsen and Niggli that they are logical and consistent and that they afford a convenient means of cataloguing rocks according to their mineralogical composition. But they fail to satisfy the requirements of a scientific classification, for no correlation is effected between the mineralogical composition of a rock and any of its other properties, or its cooling-history. Either of these classifications would be perfectly suitable for a series of mechanical mixtures, but they fail to indicate the operation of the physical and chemical laws which govern the crystallization of rocks. In short, they ignore the natural history of rocks.

ON NAMES AND SYMBOLS

In an ideal classification the names would arise out of the system; but to bring this about in petrology it would be necessary either to abandon all our present rock-names and create an entirely new set, or else to re-define the old names to fit the new classification. The experience of the last quarter of a century has shown that there is a steadily growing opposition among petrologists to the introduction of new names, and a scarcely less marked reluctance to submit to any curtailment of the right to use old, familiar names in the way they have always been used. In view of this resistance to change one must just resolve to put up for some time longer with the well-known names of the Rosenbusch era, and their "well-known" (that is, ill-defined) connotations. But it must be recognized that they are only *sack-names*, having no permanent validity and indeed no precise meaning in terms of either chemistry or mineralogy. To these names we must try to impart some degree of precision by means of adjectives and symbols such as those introduced in this chapter.

To take an example from the commonest group of rocks, the granite of St. Austell, Cornwall, is more accurately described as a peraluminous potash-granite with a colour-index of about 13; all this information may be expressed concisely by the symbol XOpα(13). The granite of Boulder, Montana, is a metaluminous soda-granite with some 22 per cent of dark minerals, or XOmβ(22). The granite of Rockall, or rather the variety called "rockallite," is a mesotype peralkaline soda-granite, XOkβ(43). The granophyre of Carrock Fell, England, is a dyscrystalline dyke-granite holding 9 per cent of augite, or DOsβ(9). The Plauen syenite, in spite of its name, is a distinctly oversaturated rock holding some 8 or 10 per cent of quartz. This slighter degree of oversaturation might be indicated by writing a small letter "o" in place of the usual capital, thus Xomα(22).

Symbols have already been used in addition to names in the Norm system and the systems of Johannsen, Niggli and Lacroix. Until petrologists are ready to consider a complete revision of nomenclature, no better course is open to us than to supplement indefinite names with definite symbols. The alternative is to continue the witless practice of giving a new name to every rock that is slightly different from any rock seen before and thus to extend indefinitely the list of "specific" names which already contains over six hundred items and includes such gems as katzenbückelite and leeuwfonteinite, anabohitsite and sviatoynossite, bogusite and bugite. There is indeed another alternative which has found a few advocates: it is to return to the language of babyhood and coin words by putting together unrelated syllables, as in neapite (nepheline-apatite rock), apaneite (apatite-nepheline rock), olpybinemelite (olivine-pyroxene-biotite-nepheline-melilite rock). There may be still more dreadful alternatives, but the writer has not been able to imagine them.

It is the thesis of this book that characters which are capable of interpretation in terms of physical chemistry are more valuable for systematic use than any others, and that rock-names should be based as far as possible on these characters. The writer believes that all the requirements of a scientific petrology can be met by a series of not more and probably less than a hundred names. It is impossible to express every character of a rock in a single word; something must be left out; and those who require more information than the name conveys will find it either in the symbol or in the detailed description of the rock.

SUMMARY OF THE SYMBOLS INTRODUCED IN THIS CHAPTER

Crystallinity

X = eucrystalline
D = dyscrystalline

Degree of Saturation with Silica

O = oversaturated (o if excess silica not more than 10 per cent)
S = saturated
V = undersaturated, non-feldspathoidal
U = undersaturated, feldspathoidal
W = undersaturated, both light and dark minerals unsaturated

Reaction Type

p = peraluminous
m = metaluminous
s = subaluminous
k = peralkaline

Colour Index

L = leucocratic

$$\left.\begin{array}{l} M' = \text{mesotype} \\ M'' = \text{melanocratic} \end{array}\right\} M = \text{melanic}$$

H = hypermelanic

Character of Feldspar

(Classes O, S, V)

α = Or > An, Or > Ab
β = Or > An, Ab > Or
γ = An > Or, Ab > An
δ = An > Or, An > Ab

(Classes U, W)

α = Or + Lc > An; leucite present
β = Or > An; leucite absent
γ = An > Or + Lc; leucite present
δ = An > Or; leucite absent

The symbols are intended to be written in the above order, except that the colour-index symbol should come last, thus XSmβL or XSmβ(14).

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The origin of granite is to be sought in the depths of time as well as of space.—R. A. DALY.

CHAPTER XV

THE OVERSATURATED ROCKS

OCCURRENCE AND GENESIS

THE oversaturated rocks are those that hold free silica in the form of quartz, tridymite, or siliceous glass. Silica of hydrothermal origin (opal, chalcedony, agate, etc.) is to be disregarded. In a glassy rock, the presence of an excess of silica may be deduced from the density and the refractivity of the glass, or from data furnished by chemical analysis.

Most rocks that hold much quartz have also a good deal of feldspar, and from the earliest days of geology such rocks, when coarse in grain, have been called granite. Fine-grained dyke-rocks and lavas of similar composition have been named quartz-porphyry, rhyolite, felsite; and the vitreous ones obsidian and pitchstone.

After the introduction of the microscope into petrology (1858 to 1870) it was recognized that both orthoclase and plagioclase feldspars occur in granite, sometimes one and sometimes the other predominating. Ferdinand Zirkel defined granite as a rock in which alkali-feldspar, quartz and plagioclase take part, the plagioclase being as a rule subordinate to orthoclase. For rocks in which plagioclase exceeds orthoclase Zirkel used the name quartz-diorite; but his contemporary, G. vom Rath, gave the specific name of tonalite to the quartz-diorite of Tonale, and this name is now widely used for any plagioclase-rich granite. At a later date Cathrein added the name adamellite for rocks intermediate between granite and tonalite, and Becker introduced granodiorite with the very same meaning. Disregarding this choice of names, some petrologists have continued to use the name granite even for rocks in which all the feldspar is plagioclase. The consequence is that when the name granite is used without qualification it has no more value than a field name. In the discussion that follows we shall use certain of the above names in the customary loose way, and afterwards an attempt will be made to define them more precisely.

It is generally believed that oversaturated rocks are much more abundant, in the outer layers of the earth-crust at least, than any other class. In most countries the data for testing this belief statistically have not been gathered together, but we owe to R. A. Daly a valuable computation of the areas occupied by outcrops of different kinds of eruptive rocks in the United States of America. For obvious reasons, no comparison of volumes is possible. Out of 6,607 square miles occupied by plutonic rocks, 5,932 square miles, or ninety per cent, consist of oversaturated rocks, mostly granite. In contrast with this enormous development of oversaturated plutonic rocks, the comparative scarcity of oversaturated lavas is surprising. Daly's figures show that out of 9,715 square miles of lava in the United States, only twenty-three per cent are oversaturated. This is a point of great significance for the theory of petrogenesis; it suggests, and indeed goes far to prove, that the extrusion of a particular type of lava is not necessarily connected with the intrusion of a plutonic rock of the same chemical and mineralogical character.

OCCURRENCE IN THE FIELD

Granite, granodiorite and tonalite occur in the form of enormous dome-like intrusions—so-called batholiths—whose upper surfaces may be exposed over hundreds or even thousands of square miles. The Coast Range batholith of British Columbia, according to T. C. Phemister, is "probably the largest plutonic body in the world," having a continuous outcrop at least 1,250 miles long and up to 125 miles wide. To the Idaho batholith in Idaho and Montana C. M. Langton ascribes an area of 16,000 square miles, although only a small part of this is massive granite. The Vermilion batholith in Minnesota covers between 2,000 and 3,000 square miles, and the Boulder batholith, Montana, about 1,200 square miles. The principal granite bodies in the south-western region of the Cape of Good Hope occur within an area of 120×60 miles. If they are all connected underground, which is likely in view of their petrologic similarity, they must form a batholith approaching 7,000 square miles in area.

The thickness or downward extension of a batholith is quite unknown, but since erosion has never exposed the floor of a really large intrusion, the thickness is certainly to be expressed in miles. It is most unlikely that any of the greater intrusive bodies rests on a floor of sedimentary rocks, but "what underlies the granites" is one of the problems that geology has still to solve.

The margin of a large granite intrusion is sometimes sharply defined and more or less plane, and one can lay a knife-blade along the junction so that all rock on one side of the blade is eruptive, all on the other side sedimentary. The contact between granite and Silurian limestone at Konerudkollen, Norway, is said by W. C. Brögger to be "hair-sharp." W. D. Johnston and E. Cloos say of the granodiorite of Grass Valley, California, that "the contacts of the granodiorite with the adjoining rocks are commonly so sharp that both rocks, in their normal facies, can be represented by a single hand-specimen." But a different type of junction is also seen, especially in regions of intense metamorphism; it is characterized by the absence of any sharp bounding surface. The appearance in such a case is as if the country rock had soaked up the liquid magma like a sponge, being itself at the same time partly assimilated and partly incorporated in the magma. In crossing such a junction one passes insensibly from pure granite to granite holding dark bands and streaks of half-assimilated sediment; from this to a composite rock made up of granitic and sedimentary laminæ; then to sedimentary rock containing veins of granite or isolated, large feldspar crystals similar to those in the granite; and eventually to sediment which has been affected by the physical conditions prevailing at the contact but which has received no visible addition of material from the magma. The streaky or banded rocks which result from this intermingling of eruptive with sedimentary material are known in general as "gneiss," and the various degrees of admixture are indicated roughly by the use of such terms as "gneissose granite" or "granite-gneiss" for those in which granitic material predominates, and "injection-gneiss" or "migmatite" for those consisting to a large extent of material of sedimentary derivation. Many of the oldest and largest granite masses in all parts of the world have this gneissose character in part, and it is usual to call them either granite or gneiss indifferently.

A. M. Macgregor has pointed out that there are two types of batholith, which he calls "Suess-batholiths" and "Daly-batholiths" respectively. The former are composed of rocks of highly variable character; enclosures of country rock are numerous and often large; and there is evidence of assimilation about the margins, which are of the intricate type just described. The "Daly-batholiths" are composed of fairly uniform rock; they show sharp contacts, comparatively few enclosures, and little evidence of assimilation. While it is true that many batholiths conform fairly closely to either the Suess or the Daly type, there are others which have an intermediate character. The rock may be massive in one part of the batholith and gneissose in another; one junction may be sharp and another intricate. It is likely that the difference between the Suess and the Daly batholith depends on the rate of movement during the emplacement of the magma, the former representing a slow upwelling of magma and the latter a comparatively rapid injection. E. Blackwelder and E. R. Baddley show that 67 per cent of the batholiths of North America have a transgressive relation towards the schistosity of country rocks, and only ten per cent are conformable to the schistosity. They suggest that cross-cutting is characteristic of higher levels in the earth-crust, and that at great depths the invaded rocks become plastic and flow, so developing a conformable schistosity.

We have seen that in the Suess type of batholith the rock is very variable in composition. As a rule the variation is more or less progressive from the centre towards the margin of the batholith. We may take as an example the Vermilion granite of Minnesota which is described by F. F. Grout as a leucocratic biotite-granite. Along its northern margin the granite develops a gneissic border-phase, there being within a mile "all gradations from gneissic to massive granite." Along the southern boundary the granite is bordered for many miles by a belt of hornblende-syenite and hornblendite, such that "a series of intermediate rocks can be selected showing scarcely a break from granite to hornblendite. No single outcrop shows a good gradation, but in a broad way the relation of the several rocks to a single period of batholithic injection can hardly be doubted." The composition of some of the rocks of this batholith is shown diagrammatically in the table below, and a map will be found on p. 270.

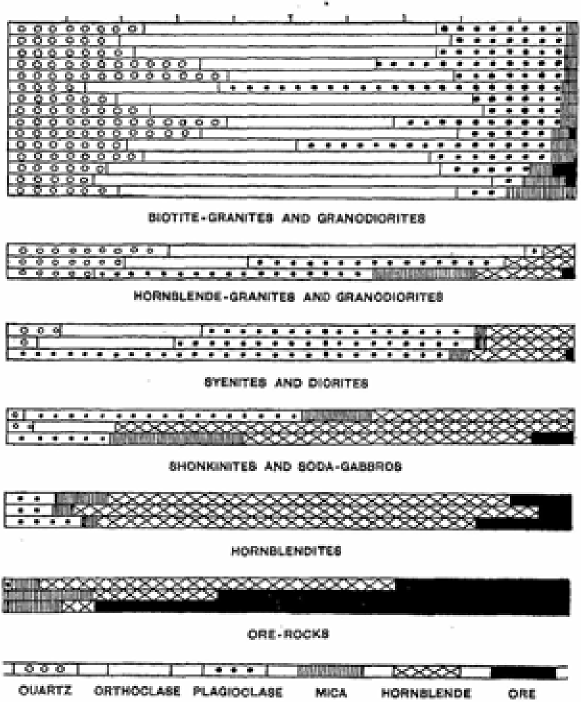


Diagram showing the composition of various rocks of the Vermilion granite batholith. Each horizontal bar represents the composition of one rock. The biotite granites and granodiorites make up nine-tenths of the whole. After F. F. Grout, with slight changes.

In other instances, variation takes place in the vertical direction. The Dedham granodiorite, in Essex County, Massachusetts, is a leucocratic rock containing both microcline and plagioclase, with 36 per cent of quartz and about 12 per cent of dark minerals. It passes downward into quartz-diorite, by decrease of quartz and microcline and increase of hornblende and augite. This again passes down into gabbrodiorite, a rock composed of plagioclase and dark minerals in nearly equal quantity. According to C. H. Clapp there appears to be continuous transition between granodiorite and gabbrodiorite. "The rocks are almost certainly magmatically related and from their structural relations apparently make up one large batholith. Most of the rock of this batholith is gabbrodiorite."

In laccoliths and sills the transition from granite (above) to gabbro or another type of basic rock (below) is often observed. In the Bushveld complex of the Transvaal, which is a giant example of the basin-shaped intrusive or lopolith, a very acid red granite overlies norite and other "basic" rocks; and similar conditions appear in the Sudbury lopolith, Ontario. In each of these cases the passage from the acid to the basic rock, which was at one time supposed to be gradational, is now known to take place through a very narrow transition-zone, and it is claimed in consequence that the acid magma came into place slightly later than the basic one. In other instances of the granite-gabbro association the same sequence has been established. The great Breven dyke in Sweden was formerly supposed to show a continuous passage from olivine-diabase into granite, but later study by T. Krokström indicates that the dyke was probably formed by successive intrusions of olivine-diabase, diabase, granophyre, and again olivine-diabase. These observations make it rather likely that the apparently continuous transition between granodiorite and gabbrodiorite, in Essex County, would also prove to be interrupted if better exposures were available.

H. H. Thomas, reviewing the evidence presented by the composite dykes and sills of the western islands of Scotland, observes that "generally the fissure was first filled by a basic magma that partially solidified," and this was followed immediately by an acid magma which as a rule "does not chill against its more basic associate but merges rapidly into it."

THE ORIGIN OF GRANITIC ROCKS

Is granitic magma "primary" in the sense of being one of the fundamental materials of the earth, or is it derived from something else? The author of the term batholith, E. Suess, defined a batholith as "a stock-shaped or shield-shaped mass intruded as the result of the *fusion of older formations*"; and again he says "batholiths have reached their present position by *melting and absorbing the adjacent rock*."

The concept of granite as the product of an igneous or molten magma has prevailed since the beginning of the 19th century; it survived even H. C. Sorby's demonstration, about the middle of the century, that some granite has completed its crystallization at such surprisingly low temperatures as 89° to 356° C., and his claim that "proof of the operation of water" in the formation of granite "is quite as strong as that of heat." It is only since the application of physical chemistry to petrological problems, during the present century, that the realization has come that granitic rocks have been formed at comparatively low temperatures, from magmas containing water and other fugitive constituents.

Speculation about the origin of granitic magma has followed three distinct lines. One group of investigators has followed Suess to the extent of regarding granite as a "fused" or completely liquefied sediment; another group, starting out from the work of Michel-Lévy and Lacroix, claims that granite is formed by the addition of certain alkaline "emanations" to sedimentary rocks; the third group follows J. H. L. Vogt and pictures the granitic magma as a residual liquid from the crystallization of much larger bodies of gabbroic magma.

Among the representatives of the first group, J. J. Sederholm may be cited. In his volume (1) on the origin of granite and gneiss in Fenno-Scandia, Sederholm says "the phenomena of refusion or resolution have occurred here so extensively that it must be assumed that *the whole area has been in a melting condition*, when it was once sunk to so great a depth that it approached the bottomless magma-ocean or tectosphere of the earth." Following Hutton, Sederholm pictures the destruction and gradual renewal of the solid earth-crust as a circulation. "The granitic magma, once solidified and in part decomposed, undergoes again, when brought into the deeper parts of the earth, a resurrection or palingenesis."

Although, as we pointed out in Chapter V, Sederholm sometimes used the words melting and fusion when his meaning was really what we should now call solution or assimilation, yet it seems clear that in the above sentences he had actual melting in view. He gives a remarkable account of the "palingenesis or refusion" of a conglomerate near Helsingfors, which was "intruded by and partly dissolved in" the post-Bothnian granite. At its junction with mica-schist the conglomerate is changed into a more or less massive rock which behaves like an eruptive, for it penetrates the schist as veins, even when it retains distinct vestiges of its former structure. "Thus we have illustrated the gradual transformation of a conglomerate into a new magma." Chemical analyses are given which show that the intrusive veins have almost the same composition as the best preserved facies of the conglomerate, so again it would seem that actual melting is indicated.

P. J. Holmquist writes of an *ultrametamorphism* of schistose rocks, by which they are converted into pegmatitic gneiss and granite. He says "I am convinced that a general segregation of the quartz-feldspar material from the rocks took place at this high degree of the regional metamorphism, and that it marks the first steps of remelting of the crust" . . . "There seems to be no possibility of bringing this pegmatization into connection with any granite." H. von Eckermann claims that the Loos-Hamra region of Sweden presents a "typical case of a regional subsidence of the earth-crust and a palingenetic re-melting of its base."

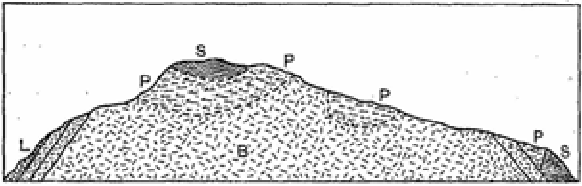
W. Wahl shows that the older granites of Finland are oligoclase-granite or granodiorite, the younger ones are microcline-granite. During the primary stage of the Svecofennian folding, the granodiorite magma was intruded into the supracrustal rocks over a considerable part of the total area of Finland. The younger granites, which are rich in potash and silica and often pegmatitic in character, are thought to have been formed by palingenesis of the older.

Sederholm modified his view in later years, for in 1926 he wrote "the word refusion does not imply that a dry fusion should have taken place." He introduced the term *anatexis* and defined it as "refusion, although in some cases re-solution would be more appropriate." Then he said specifically that "the anatexis of the crustal granitic masses has been caused . . . by the influence of emanations from the abyssal magma, whatever may have been its composition." In 1933 he wrote with more assurance about these emanations. "Granitic *ichor*, a magma diluted with water and containing other mineralizers, has played a great part both in the last stages of the consolidation of plutonic rock masses and in the formation of migmatites. This ichor percolates through the older rock-masses and dissolves some of their minerals and replaces them with others, gradually giving the rocks a more granitic composition."

Starting from the "juvenile gases" of Suess, which were supposed to be given out by the core of the earth in cooling and to be so hot that they would "melt and stope their way through the overlying rocks," we have in recent years reduced our demands to a dilute, watery magma containing the constituents which are necessary to convert siliceous and aluminous sediments into granite; in other words, a solution of potash and soda, at a temperature not much above that of a hot spring.

Since 1933 there has been a stream of publications describing the "granitization" of sedimentary rocks by emanations from adjacent bodies of granitic or granodioritic magma. To cite only a few examples, we have the description by A. L. Anderson of the Cassia granodiorite in Idaho, which has converted the surrounding quartzite into quartz-microcline granite; H. Ebert's account of the hybrid granite of Lausitz, Saxony, and the formation of aplite veins by liquation; C. E. Wegmann's study of migmatite; H. C. Horwood's study of granitization in the Cross Lake region of Manitoba; M. MacGregor's account of the evolution of quartz-diorite in Galloway, Scotland; and the description by G. H. Anderson of the granites of the Northern Inyo Range in California.

The Inyo batholith is composed of two kinds of granite, the Pellisier granite overlying the Boundary Peak granite. The last-named is a fairly uniform, light grey biotite-granite with very few enclosures. The Pellisier granite is dark grey, like the argillites and phyllites with which it is in contact; and it is decidedly variable both in composition and in texture. It is sometimes porphyritic, and it has a mottled appearance owing to the uneven distribution of hornblende and biotite; enclosures of the country rock are abundant. It is demonstrated that the Pellisier granite was formed in place, partly by recrystallization and partly by replacement of the country rocks, which include both sediments and interbedded volcanics. The necessary heat as well as the transforming solutions are believed to have come from the Boundary Peak granite magma.



Limestone, L; Schist and Argillite, S; Pellisier Granite, P; Boundary Peak Granite, B.
FIG. 36. Generalized section across the northern Inyo Range, California. (G. H. Anderson.)

H. C. Horwood has shown that in the Cross Lake region of Manitoba, rocks as different as arkose and andesite have been granitized, and that in each case the end product is a granitic gneiss which approaches in composition to the younger granite which invades both rocks. Horwood does not assume (as others have done in similar cases) that the younger granite is consequently the ultimate product of granitization, but makes only the reasonable claim that "the normal aqueous end products of a granitic magma will under favorable conditions replace the rocks they flow through and produce rocks of granitic composition."

Such studies as these do indeed demonstrate that rocks of granitic aspect and composition have sometimes been formed from sediments, by the action of solutions emanating from an adjacent body of acid magma. But the product is not granite: it is a mixed rock which *resembles* granite more or less closely. F. F. Grout (2) has discussed the formation of igneous-looking rocks by metasomatism, and shows that many writers have failed to distinguish clearly between the generation of granite and the generation of rocks that look like granite but have a totally different history. The loosely used term "granitization" requires to be clearly defined, and Grout offers the following definition: "Granitization includes a group of processes by which a solid rock (without enough liquidity at any time to make it mobile or rheomorphic) is made more like granite than it was before."

A rock produced by the recrystallization (without melting) of an arkose should not receive the same name as a rock formed by the slow crystallization of a rhyolitic magma, although the two rocks may be almost identical in mineralogy, in chemical composition, and in texture. Only one of them is granite; the other is a "pseudo-granite," or a "metamorphic granite" as it was called during last century. The recognition that some granitic rocks are metamorphic should not stampede us into the weak conclusion that no granite is eruptive.

In a further group of studies of granitization there is not even an acid or basic magma within sight; we are asked to believe that emanations *of unknown origin* have converted sediment into granite. In this group we must notice the remarkable case of the Killarney gneiss of Ontario, as described by T. T. Quirke and W. H. Collins.

North of Lake Huron a belt of Huronian sediments comprising quartzite, conglomerate, greywacke and limestone, is closely folded along an E-W axis and is invaded by small batholiths of Killarney granite. At the eastern end, where the thickness of the sediments is 23,000 feet, they suddenly vanish and their place is taken by granite and gneiss. The line of separation is fairly straight and cuts diagonally across the strike of the Huronian. In part this is due to faulting, but in other parts there is evidence of extensive granitization of the sediments, which are said to grade into pink and gray gneiss and eventually into normal granite. In the neighbourhood of

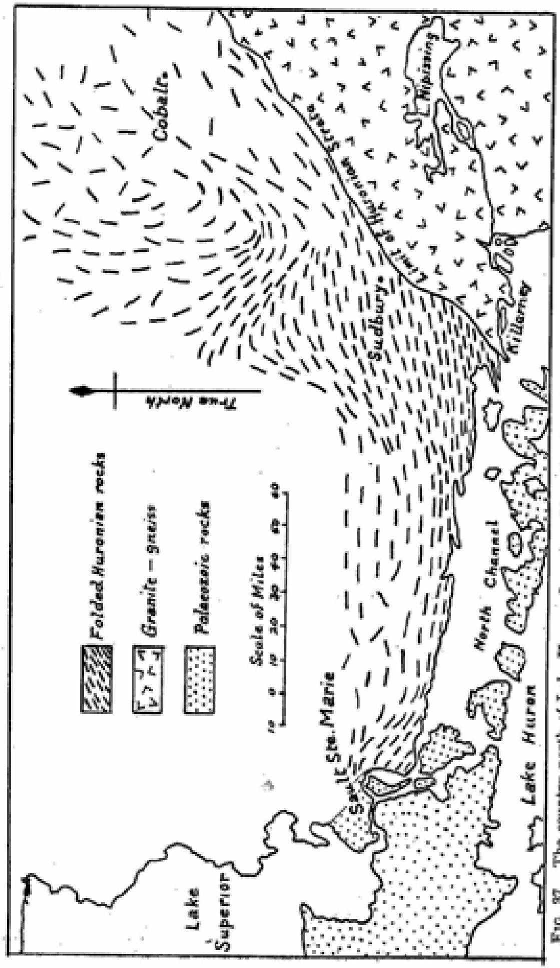


FIG. 37. The country north of Lake Huron, Ontario. The direction of folding in the Huronian sediments is expressed by the course of the lines and the intensity of folding by the thickness and closeness of the lines.

Killarney the Lorrain quartzite, which is entirely composed of quartz and white mica, begins to develop red patches and changes gradually into a dark red quartz-feldspar-sericite gneiss which retains the bedding of the quartzite. In a further stage the gneiss becomes more coarse-grained and develops insets of orthoclase; the banding of the gneiss becomes obscured and the texture becomes more and more granitic. "There is an unbroken gradation from quartzite at one end to syenite and granite at the other." Quirke points out that there are no pegmatite "feeders" in the Killarney gneiss, so it is not a migmatite. In that case we are left without any explanation of the source of the alkalis which have converted hundreds of square miles of sediments into banded gneiss and granite.

W. A. Jones has made a petrographic and chemical study of part of the Killarney area and does not accept Quirke's claims regarding granitization of the Lorrain quartzite. Jones shows that the unaltered quartzite contains 4·38% of K_2O in the form of muscovite; enough to furnish some 25% of feldspar without any addition from outside sources. His microscopic examination "did not reveal proof of the origin of the fine-grained gneiss by the metasomatic replacement of quartz by feldspar." Nevertheless Quirke (1940) insists that the field evidence is more reliable than Jones' chemical and microscopical data, and he cannot abandon the conclusion that the fine-grained porphyritic gneiss is a highly metamorphosed quartzitic rock.

An interesting divergence of views appears in the interpretation, by G. E. Goodspeed and K. B. Krauskopf respectively, of the Wallowa batholith in Oregon. The very uniform rock that composes this batholith may be simply described as a plagioclase-granite; it has been intruded into a "heterogeneous assortment" of hornstones, greenstones, limestones, limy shales, and quartzites. Goodspeed and Krauskopf are agreed that granitization has taken place at many points about the borders of the batholith, and that "all varieties of metamorphic rocks which make direct contact with the batholith can be shown to undergo locally an extreme alteration that gives them a texture and composition similar to the intrusive." Goodspeed suggests therefore that "most of the granitic rocks of this region were formed in place from sedimentary rocks by the gradual additive and subtractive action of hydrothermal solutions," and he derives these from "the deeper portions of the lithosphere." But Krauskopf argues against granitization in place as the sole means of emplacement of the batholith, on the ground of the remarkable uniformity of the granite in texture and composition, and the absence of evidence of assimilation along considerable stretches of the contact. He claims that "some movement of the batholith as a whole seems necessary to explain its simple structural pattern, its prevailingly sharp contacts, and its petrographic uniformity." Krauskopf seems inclined to substitute palingenesis in depth for granitization in place, as the origin of the batholithic rock.

In Norway and Sweden the idea of granitization by emanations which are not derived from a magmatic source has found several advocates. H. G. Backlund discusses granitization in terms of *rheomorphism,* which he defines as "the sum of the processes of partial or complete thermal liquefication of a preexisting rock with addition of smaller or larger quantities of new material which has entered by diffusion"; but he gives no indication of the source of this new material. T. F. W. Barth contends that the Telemark granite of southern Norway, which covers an area of 15,000 square miles, was formed by "an ichor of granitic composition" which rose almost vertically from the depths and granitized all rocks in its course. This ichor is thought to have been formed by "differential refusion" of the oldest pre-Cambrian rocks, followed by squeezing out of the pore-liquor thus formed. "This pore-liquor or ichor originated in the interstices of the mineral grains and extracted from the rocks the ingredients that were easily dissolved and easily fused. . . . Driven by gravity and orogenic movements . . . this liquor slowly rose to higher levels, dissolving, assimilating, or altering metasomatically every molecule of the subjacent rock masses. Thus was formed the Telemark granite."

In Chapter X we discussed the nature and composition of the residual fluid from crystallizing magma; this is the only "ichor" of which we have any knowledge, and its amount is limited by the known solubility of water in molten granite, The "partial or complete thermal liquefaction" which Backlund postulates, and the "differential refusion" pictured by Barth, would not yield such an ichor. It is possible to study the partial melting of granite in some instances. Larsen and Switzer described a mass of granodiorite which was enclosed in andesitic magma and almost half melted; and other writers have recorded similar phenomena. In no case do the two liquids appear to have mixed together. The writer has studied granitic inclusions from the lava of Paricutin, the Mexican volcano which broke out in 1943. The inclusions have been melted to a considerable extent, forming a colourless interstitial glass. The olivine-basalt which envelops them has a base of deep brown glass. Where the two kinds of glass meet there is no sign of mixing, even on a microscopic scale, and the white glass has rough, jagged edges like a solid body. By no imaginable means could a "liquid" of such extreme viscosity be separated from the associated crystals, or assume the characters of a "granitizing ichor."

Some petrologists now dispense altogether with ichors and solutions and claim that metasomatic reactions, even on a large scale, have been effected by the migration of ions through the crystal lattices in completely solid rocks. The basis for this opinion has been presented by R. Perrin and M. Roubault (1, 2). In the first of these publications these writers describe recent experimental work on reactions in the solid state; in the second they discuss the granite problem in the light of this work. They do not deny that water has played a part in the genesis of granite, but they attach much more importance to solid diffusion. J. A. W. Bugge has discussed the thermodynamics of solid diffusion and has applied his conclusions to perthitic feldspars, coronas, and the granite problem. Although he strongly believes in the geological importance of solid diffusion, he admits that a certain amount of diffusion, especially of the larger ions, must take place through intergranular films. Healthy criticism of the hypothesis of solid diffusion has come from the pen of N. L. Bowen.

We come at last to the third hypothesis of the origin of granite, which is associated especially with the name of J. H. L. Vogt. In various publications since 1908, Vogt has maintained that granitic magma is of the nature of an eutectic residue, or at any rate a "rest-magma," from the crystallization of the basic magma. In 1931 he was still convinced that "granitic magma is the final product of a long series of differentiations in the rest-magma direction," but he offered no opinion as to the mechanism by which the rest-magma became separated.

The presence of interstitial micropegmatite, composed of quartz and alkali-feldspar, in many diabases has been held to support Vogt's contention. As long ago as 1892, A. C. Lane asserted that the interstitial micropegmatite in diabase dykes in Michigan is the last product of the basic magma, and not a secondary material as others had claimed. In diabases of the Tholei type the place of micropegmatite is taken by a glassy or cryptocrystalline material. F. Walker was able to separate enough of this interstitial glass, from a Scottish "tholeiite," to have a complete chemical analysis made. The result is expressed below in the form of the calculated mineral composition.

	Entire Rock.	*Glassy Residue.*
Quartz	7·6	30·7
Orthoclase	12·2	25·0
Albite	17·8	20·4
Anorthite	17·5	10·0
Pyroxene	25·5	2·8
Ores, etc.	15·6	2·9
Water	3·6	8·7

Walker's results amount to a complete demonstration that the residue from the crystallization of this very common type of rock has actually the chemical composition of a granite. The high content of water should be noted.*

P. Eskola, starting from the logical postulate that "granite could not be formed by re-fusion of older material before the existence of something to be re-fused," maintains that the earliest method of formation of granite must have been differentiation of the *sima* or basaltic layer of the earth crust. He is satisfied that the interstitial micropegmatite of diabasic rocks is the final product of the crystallizing magma, and he thinks that a certain amount of the residual liquor might be *squeezed out*, from the mush of crystals and liquid, by earth-movements. But he keeps a foot in the Sederholm camp too, claiming that there may be some degree of differential fusion

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*See also E. A. Vincent, *Min. Mag.* 29, 1950, p. 46.

of crustal rocks when they are depressed in geosynclines. He thinks both processes have been active in forming the outer, granitic shell of the earth.

H. H. Thomas and E. B. Bailey have brought to notice eleven instances of density-stratification in dyke-rocks on the island of Mull. Each dyke consists of gabbro passing up into granophyre. In the dyke which they have chosen as their typical example there is an upward transition from gabbro. consisting essentially of plagioclase and pyroxene, through a dioritic rock containing two kinds of feldspar, to a pink rock composed of alkali-feldspar with subordinate quartz. The specific gravity of the gabbro is 3, that of the granophyre about 2·5. Thomas and Bailey contend that "In such a high column of magma as these dykes represent, the gravitational pressure of the early crystalline phases on the still liquid residue would be considerable, and would in our opinion be quite sufficient to cause an upward filtration of the still fluid portion of the mass, and its ultimate intrusion as an entity into the uppermost part of the dyke-fissure."

Both A. Holmes and C. N. Fenner have vigorously criticized this conclusion. They point out that the hypothetical granitic residual liquor could hardly have amounted to more than 5 or 10 per cent of the original magma, and they claim that a much greater deforming force than gravitation would be required to squeeze this residue out. They also point to the absence of rocks of intermediate composition and ask why the expulsion of residual magma should invariably be delayed until the residuum has reached a granitic composition?

To this rhetorical question one can only make a suggestive reply. F. Walker's analysis of the glassy residue of a tholeiite. already cited, shows a very high content of water; if 8·7 per cent of water was retained in the glass, it is likely that the residual magma held no less than this quantity of water. This would give it not only a low viscosity but also an extremely high vapour pressure; so high that actual boiling might be expected under certain conditions. The process of boiling or "gas-streaming" (as we called it on an earlier page) might cause the expulsion of some of the liquid residue from the interstices of the crystalline rock. If this be admitted as a possibility, it would explain the absence of rocks of intermediate composition, for the expulsion of the interstitial liquid would only begin when boiling set in; that is, at a very late stage in the crystallization of the system.

In further criticism of the theory of Thomas and Bailey. Holmes asserts that there is no known example of an oversaturated lava on any of the basaltic islands of the Pacific. (Actually there are several records of rhyolite or quartz-trachyte on oceanic islands, as on the Marquesas, Tonga, Hawaii, Gambier, Easter, Samoa, St. Paul, St. Helena, Ascension, but the amounts are relatively insignificant.) However, W. Q. Kennedy retorts that the tholeiite type of diabase or basalt, which is the type that gives granitic residua, is equally scarce on oceanic islands. The basalt of these islands is typically olivine-bearing and often olivine-rich, whereas tholeiite is either free from olivine or else it holds only a little olivine by reason of the non-completion of the Bowen-Andersen reaction.

Now, although rhyolite is rare on oceanic islands, trachyte is commonly found there in small amounts, as on Ascension. St. Helena, Mauritius, Hawaii, Samoa; and it has always seemed impossible to explain this association except by the separation of residual liquid from crystallizing olivine-basalt. If such an idea is acceptable in the case of the basalt-trachyte association, why should it be unacceptable in the case of diabase and granophyre? The writer suggested some years ago that gas-streaming (*i.e.*, boiling) may have been the cause of the separation of trachyte from basalt on the island of Mauritius. He would extend that suggestion to the case of gabbro and granophyre. The process pictured is a *purely mechanical* expulsion of residual fluid by bubbles formed in the interstices of the crystal network when the vapour pressure reaches its highest value. It is not the same as the transference of material in the gaseous state, which has been discussed by C. N. Fenner.

Out of this review of the evidence bearing upon the origin of granite, two facts emerge. One is that granitization, or the conversion of sedimentary rocks into something that resembles granite, has taken place on an enormous scale in many parts of the world. The other is that the residual liquor from the crystallization of basaltic magma may approach granite in composition. Furthermore, it appears that this granitic residue may on occasion become separated from basic rock and form semi-independent bodies of granophyre and red granite. The question arises whether it is possible to identify the feldspathizing "ichor" of one school of petrology with the "residual liquor" of the other? The answer must be *no*, because the ichor that can (by hypothesis) granitize hundreds of square miles of sediments must be enormously more plentiful than the residual liquor of all the known basic intrusives in the neighbourhood.

To throw all responsibility for the widespread granitization of sediments upon a hypothetical ichor is not a satisfactory solution of the problem. Perhaps some of the phenomena that are attributed to the ichor are capable of a more prosaic solution; namely, the *redistribution of alkalis by connate water* contained in the pores of the sediments. It has been demonstrated experimentally by P. W. Bridgman that a pressure equal to the weight of 17 miles of rock is not able to close all the pores in rocks. Some proportion of connate water must therefore be retained even in the deepest sediments. It may be heated and set into circulation by intrusive magma, and to a certain extent it may be activated by the fugitive constituents of such a magma; but perhaps the main effect of the circulating fluid may be *the redistribution of matter that is already there*. rather than the introduction of new matter.

It seems to the writer that we should do well to keep an open mind for a little longer, as regards the origin of granite. While freely admitting the reality of granitization (in the sense of Grout's definition), we should beware of attaching a miraculous potency to "ichors" and "emanations." We do not explain one mystery by substituting another for it. Palingenesis remains a hypothesis, but rhyolitic magma is a fact. For the present, the wisest procedure is to follow the petrologists of last century and admit the existence of two kinds of "granite," respectively eruptive and metamorphic. Concerning these we quote the words of Ferdinand Zirkel, written in 1876:—

> "The granites of the Fortieth Parallel are doubtless partly eruptive rocks, which have broken through sedimentary strata of a different geological age, and in part dependencies of the old crystalline schists, alternating with gneisses, etc., and showing no sign of eruptive character. Whatever may be the origin of the crystalline schists, that of the accompanying granites must be the same. According to the most favorite theory, these granites that are not eruptive, and are at the same time generally stratified, should be called metamorphic granites."

Even that staunch protagonist of granitization, H. H. Read, admits at the end of his meditations that there may be "granite *and* granite."

The Occurrence and Genesis of Rhyolite

We have already quoted a well-known computation by R. A. Daly, which shows that oversaturated plutonic rocks form something like ninety per cent of the area of all the deep seated rocks of the United States of America, whereas oversaturated lavas make up only twenty-three per cent of the volcanics. For gabbro and basalt (including both olivine-free and olivine-bearing types) the position is reversed, because gabbro, anorthosite and diabase together account for no more than ten per cent of the plutonic rocks, while seventy-five per cent of the volcanic rocks are either basalt or pyroxene-andesite. Geological considerations make it easy to understand why a particular kind of lava should cover a greater area than the corresponding kind of plutonic rock, but hard to conceive why it should cover a smaller area. If all granitic rocks were pre-Cambrian and all basalts Tertiary, an explanation might be found in terms of erosion; but the Jurassic granites of the Sierra Nevada and the Coast Range, and the pre-Cambrian anorthosites of Quebec and New England, make such an explanation untenable.

The data quoted above contain a very definite suggestion that the extrusion of rhyolitic lava is not necessarily connected with the intrusion of a body of granite. We have already discussed the possibility that some granite may be derived from the residual fraction of a body of crystallizing *sima* or basaltic magma; if so, then some rhyolite may have a similar origin.

There is abundant evidence of a close connection in time as well as in space between rhyolite, andesite and basalt. H. S. Washington has furnished several instances of this association; notably (1) at Monte Arci, Sardinia, where rhyolite was followed by andesite and basalt, and (2) in the Lipari islands. where the earliest eruptions were basaltic, the later ones andesitic, and the youngest rhyolitic. At Steens Mountain, Oregon, it appears from the work of R. E. Fuller that basalt was followed by rhyolite, dacite and andesite, and the series ended with a recurrence of olivine-basalt. In the Newberry volcano, Oregon, H. Williams established the following succession of flows:—

6.	Rhyolite (in south wall only)	up to	1000	feet	thick.
5.	Basalt (in north and east walls)	,,	450	,,	,,
4 and 3.	Basalt and andesite	,,	500	,,	,,
2.	Rhyolite	,,	400	,,	,,
1.	Basalt	,,	2000	,,	,,

M. C. Bandy has lately described the petrology of Easter Island, in the east-central Pacific. The youngest volcanic product of this typically basaltic island is an acid obsidian, which Bandy estimates to form only 0·02 per cent of the amount of basalt.

Such evidence as this compels one to admit that basalt and rhyolite may have a common origin, in which case it must follow that the rhyolite was derived from a basaltic parent, since the converse is clearly impossible. Nevertheless, some rhyolite must be connected with underlying granite. P. Lapadu-Hargues has recently described a granite pluton in the Western Sahara which has a roof of rhyolite into which it passes by continuous transition.

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It is just as important to eliminate this theory or that as it is to support this theory or that.—R. A. MILLIKAN.

CHAPTER XVI

THE SATURATED (AND SLIGHTLY OVER-SATURATED) ROCKS

OCCURRENCE AND GENESIS

A SATURATED rock is one which contains neither free silica nor any unsaturated mineral such as nepheline or olivine. Since saturation with silica is just an intermediate stage between oversaturation and undersaturation, saturated rocks ought to be relatively scarce. Many rocks have been described as syenite, monzonite, or diorite which actually contain ten to twenty per cent of quartz; and many gabbros, diabases and basalts have either a little interstitial quartz or a good deal of olivine.

A small quantity of an unsaturated mineral is easy to recognize, for olivine stands out in high relief in thin sections and the presence of feldspathoids is most easily demonstrated by staining (S. J. S., 1); but quartz in small quantity is less easy to recognize and it may be present to the extent of several units per cent without being detected under the microscope. In a hypocrystalline rock any excess of silica tends to be retained in the glassy base, where it cannot be detected without chemical analysis; and even when a complete analysis is available the calculation of the mineralogical composition presents difficulties. The norm is no safe guide to the presence or absence of quartz, for in consequence of the method of computation quartz may appear in the norm, although it is entirely absent from the rock.

For these reasons it is difficult to apply the saturation criterion strictly as regards the presence or absence of quartz. A further consideration is that we could not apply it strictly without redefining certain names such as syenite, monzonite and diorite, and it is clear that the majority of petrologists are not prepared for such a change. Under these circumstances we can only continue to use these names in a loose way, but we should try to avoid using them so loosely that they have no definite mineralogical or chemical significance at all. There can be no justification for giving the name syenite to a rock with 25 per cent of quartz, or even with 15 per cent of quartz. When the proportion falls as low as 10 per cent, the quartz is interstitial and not easy to see without the microscope; under these conditions the name quartz-syenite may be acceptable. It is proposed, then, to discuss along with the saturated rocks those transitional varieties in which the proportion of quartz or free silica does not exceed 10 per cent. (The presence of this small amount of quartz in any particular rock may be indicated by writing a small letter o in place of the capital letter in the symbol of the rock.)

But as regards the lower boundary of the class there should be no concession to a careless custom. The state of exact saturation with silica marks a phase-boundary in the system, at which one characteristic phase has just disappeared and another is just about to appear; the presence of the smallest recognizable quantity of any unsaturated mineral such as nepheline or olivine indicates the crossing of the boundary and justifies, or even demands, a change in the name of the rock.

THE OCCURRENCE AND GENESIS OF SYENITE

The majority of syenites are either marginal modifications of great bodies of granite, or else they rise in the form of dykes, stocks or cupolas from the roof of a concealed granite batholith. Thus the Plauen and Meissen syenites of Saxony are local phases of the Meissen granite; the granite of Upsala, Sweden, passes locally into syenite; the Beverly syenite in Massachusetts is a facies of the Quincy granite; the Vermilion granite in Minnesota has a syenitic border facies; and the syenite of Bulawayo, Rhodesia, forms a lobe of the Matopo granite. In some instances, however, syenite forms a marginal facies about a foyaite intrusion; this is the case at Umptek, in the Kola Peninsula, and Pilansberg, Transvaal. In the Haliburton-Bancroft district of Ontario, syenite is associated with granite and foyaite and is transitional into both rocks. A transition from granite through syenite into foyaite has been reported at French River, Ontario; at Ditró, Roumania; and at Ilimausak, Greenland.

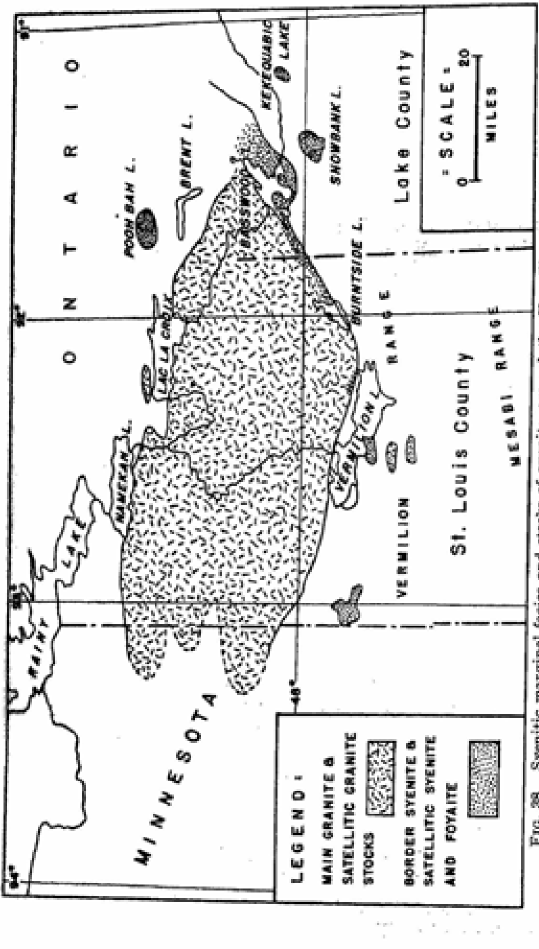

FIG. 38. Syenitic marginal facies and stocks of syenite around the Vermilion batholith, Minnesota—Ontario. (F. F. Grout.)

The syenite of the Adirondack Mountains, New York State, as associated with anorthosite and R. Balk formerly regarded it as the "mother liquor" left behind by the removal of labradorite crystals from a gabbroic magma when in the state of a "crystal mush." But W. J. Miller (1) says emphatically that he has never found any proof that either syenite or granite is a differentiate of anorthosite in this region, and he claims that where apparent transition is seen it is due to assimilation of anorthosite by a younger granitic-syenitic magma. A. F. Buddington also takes this view of the matter.

Small stocks and bosses of syenite are not uncommon in the neighbourhood of larger granite intrusions. In a belt about 10 miles wide around the Vermilion batholith, F. F. Grout indicates at least half a dozen such stocks at Rainy Lake, Poohbah Lake, Snowbank Lake, Kekequabic Lake, and Vermilion Lake (*Fig.* 38). The average diameter of these stocks is not more than five miles, and some of them include shonkinitic and foyaitic rocks as well as syenite. In the Transvaal the consolidation of the red granite of the Bushveld Complex was followed by the eruption of syenitic magma through vents and fissures, as at Leeuwfontein, near Pretoria.

The Monteregian Hills in Quebec are formed by nine small stocks or laccoliths which are in the main syenitic in composition. Brome Mountain, the largest of the group, is a laccolith covering 30 square miles; Shefford Mountain covers 9 square miles; the others are bosses of 3 square miles or less. Intrusives of the same geological type and similar petrographic character reappear in New England; J. E. Wolff enumerates five in New Hampshire, three in Vermont, and two in Maine. The syenite is associated with foyaite and other rocks, and in no instance does the syenitic facies occupy more than six square miles.

In most of these cases it is impossible to doubt that the syenite existed as a magma; at Leeuwfontein, for instance, the syenite is associated with lavas of its own composition. But H. Ebert claims that the well-known syenite of the Plauenscher Grund, near Dresden, was formed by an accumulation of feldspar crystals from the adjoining Meissen granite magma, and that only the cementing matter between the feldspar crystals represents the original fluid. F. L. Amm has formed a similar opinion regarding the genesis of the syenite at Bulawayo, Rhodesia, in which the proportion of large feldspars to groundmass is excessive. It is not unlikely that the mechanical accumulation of crystals may have played a part in the genesis of certain other syenites, particularly those with prominent banded structure and parallel texture.

In a remarkable instance described by O. H. Erdmannsdörffer in the Harz Mountains it is claimed that pyroxene-syenite has been formed by partial refusion of trachytic tuff, the necessary heat being supplied by an intrusion of gabbro.

The evidence cited above regarding the occurrence of syenite should be enough to prove that there can be no independent body of syenitic magma in the lithosphere. Syenite has generally been formed from granitic magma that has undergone desilication and contamination, though in some cases it may perhaps represent a residual product from the crystallization of gabbroic magma. The problem of the origin of syenite by desilication of granitic magma will be taken up in a later chapter, in connection with the genesis of feldspathoidal rocks. The whole amount of syenite in the lithosphere is insignificant in comparison with granite or gabbro.

The name *shonkinite* was given by W. H. Weed and L. V. Pirsson to a melanocratic syenite at Shonkin Sag, Montana, which was described by them as consisting of "nearly equal volumes of augite and orthoclase." A little nepheline or sodalite is present in some facies of this rock, but the name should be used with the meaning given to it by Weed and Pirsson. Other rocks of this character have been described under the name shonkinite in Montana, Minnesota, British Columbia, the Ural Mountains and the Transvaal; so we may regard shonkinite as the accepted name for mesotype and melanocratic syenites. Shonkinite is typically a transitional facies between syenite and pyroxenite and it is probably due to a purely local enrichment of the syenite magma with pyroxene crystals. There are no lavas of shonkinitic composition. The chilled marginal phase of the shonkinite at Shonkin Sag is a dense rock of basaltic appearance; it contains more soda and potash than the shonkinite, and less lime, magnesia and iron,

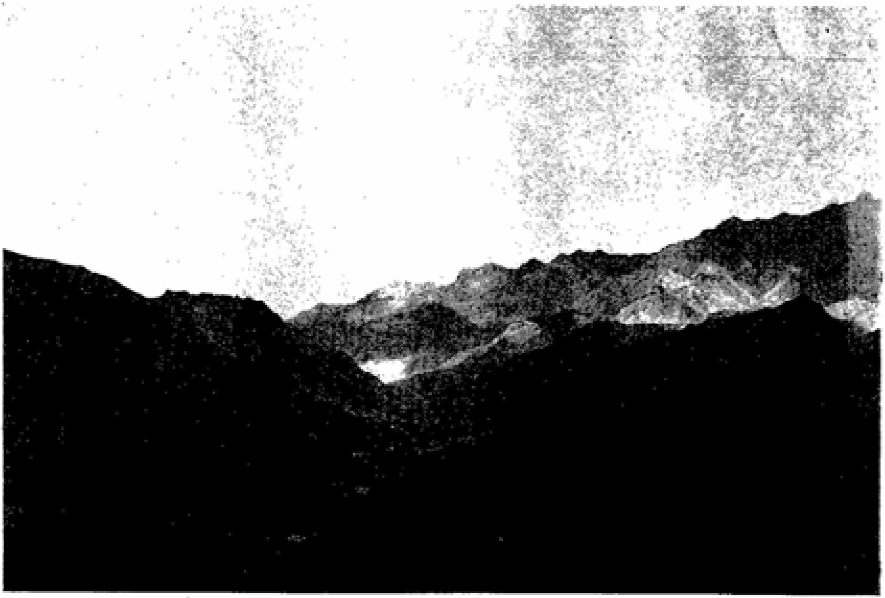

PLATE VI.—Trachyte plugs (light coloured) in olivine-basalt. St. Helena. [*To face p.* 272.

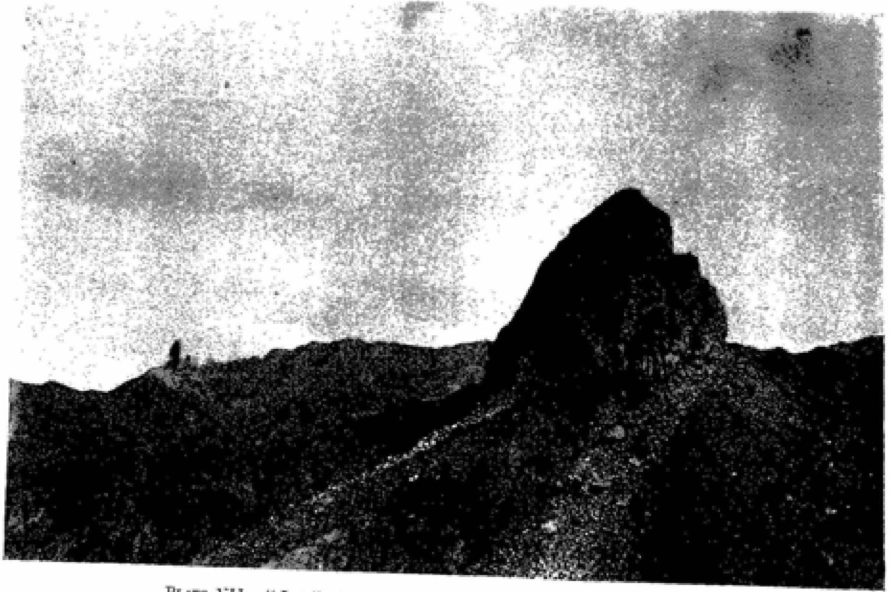

PLATE VII.—"Lot," St. Helena—a trachyte plug pushed up through olivine-basalts. In the distance (*left*) is "Lot's Wife," a similar plug of trachyte.

[*To face p.* 273.

and it formed leucite in cooling. This is a clear demonstration that the shonkinite was not formed from a magma of its own composition.

THE OCCURRENCE AND GENESIS OF TRACHYTE

This is a general name given to lavas of syenitic composition. Trachyte is certainly commoner than syenite, but it is difficult to form a numerical estimate of its importance in the lava-fields of the world. Quartz-bearing trachytes and their tuffs cover a very large area in Kenya, where they have a visible thickness of at least 800 feet in the walls of the Great Rift Valley. They are associated with phonolite and olivine-basalt. Erosion has sometimes exposed underlying plugs of syenite-porphyry which occupy the site of vents from which the trachyte was extruded (S. J. S., 2). Trachyte is widely distributed among the volcanic rocks of the Rocky Mountains and

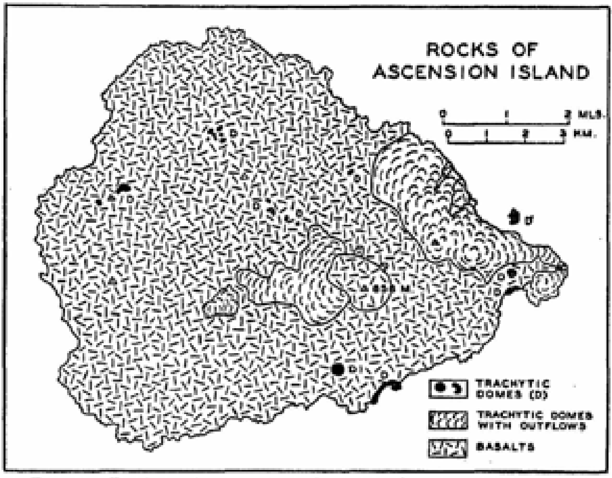

FIG. 39. Trachyte plugs and flows in the dominantly basaltic cone of Ascension Island. (R. A. Daly.)

the Great Basin, especially in Montana, Wyoming, Utah and Nevada, but little can be said regarding its genesis there. Of greater interest are those localities where trachyte is associated, in quite small bulk, with enormous effusions of basalt. This association is well known in the Auvergne district, France; in the western islands of Scotland; and in Iceland. It is still more striking to find the association of trachyte with olivine-basalt on oceanic islands such as Samoa, Tahiti, Hawaii; Mauritius, Réunion, Kerguelen; St. Helena, Ascension, Cape Verde Islands; because in these cases it is hardly possible to doubt R. A. Daly's contention that trachyte is a derivative of basaltic magma. Daly (1) points out that trachyte, in these circumstances, characteristically fills narrow pipes or subsidiary vents which rise through thick piles of the dominant basalts; that the trachyte magma must have been fully liquid, since it has sometimes given rise to trachytic obsidian; that the trachyte appeared at a late stage in the history of the volcano; and that many effusions of trachytic magma have been immediately preceded by more or less violent expulsion of gas. It has seemed to the writer that the last of these observations may afford a clue to the meaning of the basalt-trachyte association. He has supposed (S. J. S., 3) that violent evolution of gas may have driven the residual fluid out of the interstices of the largely crystallized basaltic magma beneath the volcanic vent, and caused it to be extruded independently. Since the residual fluid from tholeiitic (olivine-free) basalt has a rhyolitic composition (see page 261), it is not unreasonable to assume that the "rest magma" of olivine-basalt will be distinctly less siliceous. We conclude that trachyte may either be connected with syenite underground, as in Kenya and the Transvaal, or it may be derived directly from underlying olivine-basalt.

The trachyte domes of Ascension and St. Helena were ably described by R. A. Daly; those of Maui and Samoa have been studied recently by Stearns and Macdonald.

THE OCCURRENCE AND GENESIS OF MONZONITE AND DIORITE

The names monzonite and diorite have been used so loosely that it is difficult to establish any generalization about rocks so named. Monzonite was originally a collective name used by de Lapparent for all the plagioclase-bearing rocks found at Predazzo and Monzoni, and one cannot expect a geographical name to have a precise mineralogical significance. Some so-called monzonite is syenite, some is diorite, and some is tonalite. Of the so-called diorites some hold 30 or 40 per cent of quartz and some hold two to four times as much orthoclase as plagioclase; these are no diorites at all, but granites, granodiorites and tonalites. It is certain that monzonite and diorite, according to any strict mineralogical definition, are much less common than one would gather from the frequency with which these names are used in geological publications.

As regards the genesis of monzonite and diorite, we turn again to the observations of H. H. Thomas and W. Campbell Smith on the inclusions in the granite of Trégastel, Brittany, which we discussed in Chapter V. This granite has broken through a body of olivine-norite, enclosing pieces of it which have reacted with the granitic magma. The olivine and hypersthene have been replaced step by step by clinopyroxene, hornblende, and biotite, while the plagioclase has been progressively enriched with albite, with the consequence that the enclosures have every possible mineralogical composition between that of norite and that of granite. The intermediate types are indistinguishable from diorite and monzonite. The possibility is definitely indicated that some bodies of monzonite and diorite may be hybrid rocks.

Observations in every way similar to those just mentioned have been made by A. K. Wells and S. W. Wooldridge in the island of Jersey. A granitic magma has incorporated blocks of gabbro and has reacted with them, transforming them into apparent diorite. More complete evidence is available in Glen Dubh, Arran, where A. Harker (in 1903) and G. W. Tyrrell (in 1928) have described a zone of dioritic " mixture-rocks " between gabbro and granite. This zone includes rock types that are clearly xenolithic and hybrid, also homogeneous diorite and quartz-diorite in which no trace of mixture or hybridism can be detected. According to Tyrrell " the petrographic evidence renders it probable that the pure, homogeneous diorite and quartz-diorite are the final results of the complete solution of the gabbroidal rocks within the granite magma. All stages in the process can be studied in the field and under the microscope."

M. MacGregor claims that a body of homogeneous quartz-diorite, near Dalbeattie, Scotland, was formed by the "granitization" of sedimentary hornstones. Some degree of granitization (see Grout's definition on p. 256) may readily be admitted, but the evidence that all the quartz-diorite was formed in this way fails to carry conviction. The possibility is not excluded that the quartz-diorite is a contaminated facies of the granodiorite magma that followed it.

The monzonite of the type locality in the Tyrol is a very variable rock, including syenitic, dioritic and gabbroidal facies. All are pyroxene-bearing, but the more siliceous types contain hornblende and biotite in addition. These rocks fall into place in a series of intrusions and extrusions which W. C. Brögger summarizes as follows :—

(1) basic dykes and flows (diabase and basalt);
(2) basic plutonics (pyroxenite, passing into gabbro-diabase and monzonite);
(3) monzonite with dioritic and syenitic facies;
(4) biotite-granite;
(5) dykes of camptonite and liebenerite (pseudo- nepheline) porphyry.

The rocks are intrusive into Triassic crystalline limestone, and since the youngest dykes are undersaturated with silica, the possibility that the granite magma has reacted with limestone must be admitted.

F. Walker and C. F. Davidson have described diorite and monzonite, ranging from leucocratic to highly melanic varieties, which form marginal facies of the Dorback granite in the central Highlands of Scotland. The rocks are intimately associated with a body of crystalline limestone, and it is suggested that the melanocratic diorite and monzonite are due to contamination of the granite magma by assimilation of limestone. The circumstances are thus very similar to those at Monzoni.

Considerable interest attaches to the description by D. L. Reynolds of diorite and monzonite of hybrid origin at the eastern end of the Newry granodiorite mass, in northern Ireland, as described in Chapter IX. These hybrid rocks surround and grade into a body of biotite-pyroxenite, from which they have undoubtedly been formed, although there is a difference of opinion as to the details of the process. The evidence is open to more than one interpretation; but the description of the hybrid rocks is admirably complete and the phenomena have a direct bearing on the origin of diorite and monzonite in other regions.

In the Cortlandt complex of New York State, hornblende diorite forms a marginal facies between norite and the acid Manhattan schist, and holds many enclosures of the latter (G. S. Rogers; S. J. S., 4). Some assimilation of schist has taken place, but the chemistry of the reaction still awaits study.

These instances should serve to show that an origin by assimilation or hybridism is more than likely for all occurrences of diorite or monzonite. This is indeed to be expected, because all occurrences of these rock are of small dimensions. The entire monzonite syenite-diorite complex at Monzoni, which has played such a large part in petrographic literature, actually has an area of only two square miles. The diorite of the Cortlandt complex covers about the same area, forming no more than one-tenth of the intrusive body to which it belongs. The diorite of Garabal Hill occupies two and one-half square miles; the granite with which it is associated covers five times this area (Dakyns and Teall). The diorite stock at Mount Ascutney, Vermont, in which hornblende-mica-diorite is connected by all gradations with typical gabbro or diabase, occupies only 4 square miles (R. A. Daly, 2). One cannot assume an independent origin for such insignificant bodies of monzonite and diorite, any more than for similar bodies of syenite.

Three distinct modes of origin are suggested by the instances described above: one by reaction between acid magma and basic eruptive rock; another by reaction between acid magma and limestone; and a third by reaction between basic magma and acid country rocks. Still another possibility is suggested by the observations of J. L. Gillson in the Pioche district of Nevada. Gillson shows that a magma which began to consolidate as norite was attacked by solutions (gases according to Gillson) which brought about a replacement of pyroxene by amphibole and of plagioclase by orthoclase. resulting in the formation of quartz-monzonite. Perhaps these solutions were the forerunners of an underlying body of granite; if so, this mode of origin is not really different from the one first mentioned.

THE OCCURRENCE AND GENESIS OF ANDESITE

We have seen that rhyolite at the surface of the earth is not necessarily connected with granite underground, or trachyte with syenite. We may safely add that the appearance of andesite on the surface is no proof of the existence of diorite in depth. It is difficult to discuss the genesis of andesite on account of the very loose way in which the name has been used by different writers. The andesite of Zirkel was a rock containing any kind of plagioclase together with either hornblende or biotite. The andesite of Rosenbusch contained any kind of dark mineral, together with a plagioclase not more calcic than andesine. Some petrographers have followed Zirkel, some Rosenbusch, and some have distinguished andesite from basalt by the silica percentage only.

The well-known computation by R. A. Daly (3), of the areas occupied by different kinds of rock in the United States of America, shows pyroxene-andesite to be more than a hundred times as abundant as the hornblende and mica-andesites. It is actually the commonest extrusive rock in the country, only basalt being comparable with it. The hornblende-mica andesites are nearly as scarce as trachyte, with which they are commonly associated in the field. These points suggest that Zirkel, in separating the hornblende-mica rocks from the pyroxene rocks, came nearer to a natural classification than Rosenbusch did.

If trachyte may be a differentiate from basaltic magma, one cannot deny the same possibility in the case of andesite. Yet Daly makes the significant observation that sills which exhibit gravitational differentiation of basaltic magma seldom (perhaps never) show any andesitic facies. This is not hard to understand. If andesite (in the Rosenbusch sense) were to be formed from basaltic magma by crystal settling, it would be necessary for the early-formed plagioclase crystals to sink to the bottom along with some of the augite, leaving the liquid relatively richer in albite. Now the sinking of augite has been demonstrated beyond doubt — for instance, in the Palisade sill (page 147)—but the sinking of plagioclase has never been conclusively demonstrated in any intrusive rock. If the plagioclase floated instead of sinking, the result would not be andesite but only a more feldspathic basalt, as at Kijabe (page 143). The separation of crystals from basaltic magma is therefore not a likely way for andesite to be generated. The expulsion of residual fluid from partly crystallized basalt is not any more likely; we have seen that trachyte and even rhyolite may have been formed in this way, but there is no field evidence which suggests such a mode of origin for andesite. The probable conclusion, therefore, is that the production of pyroxene-andesite depends upon the acidification of basaltic magma by reaction with the siliceous rocks of the continental regions. But hornblende-andesite and biotite-andesite may result from contamination of acid magma, just as some diorite does. It is not without significance that cordierite, a typical product of contamination, has been found in certain andesites; for instance, by H. A. Brouwer in the island of Ambon, East Indies.

THE OCCURRENCE AND GENESIS OF GABBRO, NORITE, ANORTHOSITE AND DIABASE

Gabbro, norite and anorthosite are sometimes found in enormous intrusions almost comparable with the great granite masses. The norite of the Bushveld complex, with an area of nearly 20,000 square miles, is one of the largest intrusions in the world. The anorthosite intrusions of eastern Canada occupy about 60,000 square miles of the surface of Quebec and Labrador. F. D. Adams has estimated the area of the Saguenay mass to be 5,800 square miles, and the Morin mass 1,000 square miles. The Sudbury gabbro in Ontario covers 500 square miles; and K. S. Fowler has described "four hundred square miles of labradorite rock" in the Laramie Mountains of Wyoming The gabbro and norite intrusions of north-east Scotland have a total area of about 1,000 square miles, but if they form one huge sill, as H. H. Read supposes, then the area may be fully ten times as great.

Although their extent is so considerable, it has been definitely established that many of these intrusions have the form of laccoliths or lopoliths, not that of batholiths; consequently the volume of rock contained in any of them is much less than that in a batholith of the same area. The Bushveld norite is a lopolith with a probable thickness of three miles, the Sudbury gabbro is also a lopolith and about two miles thick. Even the Canadian anorthosites are considered to be laccolithic in form, and the Adirondack intrusion to be a floored sheet. The numerous masses of norite in the charnockite series of India are described by H. H. Holland as lenses and thick dykes.

In striking contrast to many granite batholiths (Daly batholiths) which are almost constant in composition and texture over large areas, these great intrusions of plagioclase-rocks often show extreme variability. By diminution of the dark minerals leucocratic facies are developed, which pass into the almost wholly feldspathic anorthosite; the converse process gives melanocratic gabbros and pyroxenes, formed almost entirely of pyroxene or pyroxene and ore. Or, again, the pyroxene may change its character from monoclinic augite or diallage to ortho-pyroxene, giving rise to norite. Hornblendic facies and facies containing olivine may also be developed, as well as relatively acid, dioritic or syenitic bands.

But gabbro also makes its appearance in a more fine-grained form which is as remarkable for its uniformity as the coarse variety for variability. This is the so-called diabase or dolerite which has a world-wide distribution in the form of dykes and sills. Although this variety of gabbro is common in every country from Spitzbergen to Antarctica, it is displayed on the most spectacular scale in the Permo-Triassic rocks of South Africa, where A. L. du Toit has said that "one can never get out of immediate sight of dolerite" within an area of more than 220,000 square miles. Individual sills are often 100 to 300 feet thick, and an exceptional sheet in East Griqualand, which is so coarse-grained as to be largely gabbro, reaches 3,000 feet in thickness. Du Toit considers that the volume of rock in this sill alone may have exceeded 300 cubic miles. Rocks in every way similar to these South African dolerites are developed under identical geological conditions in South America, India, Tasmania and South Victoria Land (Antarctica). The average composition of all such rocks is about: labradorite 40-50, augite 30-40, orthoclase 0-5, quartz 0-5, olivine 0-5, ore 2-10 per cent.

The remarkable chemical and mineralogical resemblance between the dolerites and diabases of widely separated localities suggests a common parentage for all of them and invites us to believe in the existence of a primary world-magma of gabbroic composition. We shall return to this question after we have given some consideration to the basalts, which are the extrusive members of this great group of rocks.

Within the gabbro group, the norites and the anorthosites present their own special problems. The name *norite* was originally given to a Norwegian hornblende-gabbro, but in the course of time it has been transferred to gabbros containing rhombic instead of monoclinic pyroxene. Many large intrusives consist partly of gabbro and partly of norite; this is the case, for instance, in the Bushveld complex, where monoclinic pyroxene predominates in the upper part of the "norite" sheet and rhombic pyroxene in the lower. In other instances a body of gabbro or olivine-gabbro passes peripherally into norite. Such a case was described by A. Lacroix at Le Pallet, France. The gabbro mass, measuring five miles by three, is intrusive in a series of mica schists. Towards its margins the gabbro becomes finer in grain and turns into a micaceous norite holding more or less quartz and cordierite. In part this cordierite-norite holds enclosures of schist which show every stage of transformation into norite. Lacroix claims, accordingly, that the norite is a modification of the gabbroic magma, produced in consequence of assimilation of schist.

The chemistry of the process is simple. When alumina is added to a magma containing the components of plagioclase and diopside ($CaMgSi_2O_6$) it combines with calcium silicate to form anorthite which goes into the plagioclase. Magnesium silicate is therefore left over, and crystallizes independently as orthopyroxene. When all the lime has been precipitated as anorthite, any excess of alumina is free to combine with magnesia and form biotite or cordierite.

The observations made by Lacroix were soon repeated by A. N. Winchell at Snowbank Lake, Minnesota, and led Winchell to the same conclusion regarding the generation of the norite facies. In north-east Scotland, W. R. Watt and H. H. Read found other instances of gabbro bordered by cordierite norite, and Read has described the evidence in full detail. There can be no doubt that these cordierite-bearing norites have been produced by contamination of normal gabbroic magma; whether this is true of *all* norite is a more difficult question. Read says "if there were not so many slates there would not be so many norites"; but Bowen (3) is more cautious and will admit only that "the formation of norite and of pyroxenites characterized by orthopyroxene . . . *may be facilitated*" by reaction between gabbroic magma and aluminous sediments. It is certainly difficult to believe that all the vast amount of orthopyroxene contained in the pyroxenites of the Bushveld complex and the Great Dyke of Rhodesia can have been produced by this reaction.

The problem of the genesis of *anorthosite* is also a baffling one. According to the original definition, anorthosite consists "almost entirely of plagioclase." Some anorthosite has as little as 3 per cent of dark minerals, but the name has gradually been extended to rocks with 10, 20, and even 30 per cent of pyroxene, hornblende, ilmenite, epidote or garnet. Every known body of anorthosite seems to be associated with ordinary gabbro or norite, and there is a complete gradation from the former into the latter. Yet it is remarkable that although gabbro and norite are freely intruded in the form of dykes among other rocks, and are represented in the extrusive form by the very abundant basalts, yet dykes of anorthosite are rare and lavas of similar composition are almost unknown. N. L. Bowen (3) contends that no such thing as an anorthosite magma ever did or could exist, basing his argument on the high melting point of the lime-rich plagioclases, and that all anorthosite rocks must have been formed by the separation of plagioclase crystals from gabbroic magma. Bowen conceives that in such a magma the heavy phases such as pyroxene will first sink to the bottom of the magma-chamber, forming a basal sheet of pyroxenite, and that when the residual magma is sufficiently lightened by this process the plagioclase crystals will also sink and give rise to a body of anorthosite, leaving the uppermost portion of the magma with syenitic or granitic composition. N. H. Winchell, too, believed in the formation of anorthosite by the local accumulation of plagioclase crystals, but he considered the gabbro magma at Duluth to have been denser than the crystals, which accordingly floated upward and formed bodies of anorthosite near the roof of the intrusive. F. F. Grout formed the same opinion at Pigeon Point, Minnesota; and W. Larsson maintains that in the Nygard pluton, in Sweden, hypersthene crystals sank while plagioclase crystals floated. R. Balk has pictured the formation of anorthosite in the Adirondacks as a consequence of friction between the magma, the suspended labradorite crystals, and the walls of the channels through which the magma rose. Increasing friction is supposed to have retarded the movement of the suspended crystals, causing them to gather together in clots which gradually grew to be bodies of anorthosite.

On the other hand we have the contention of W. J. Miller that the sinking of pyroxene crystals, in the gabbro magma of the Adirondacks, left a viscous but still fluid plagioclase magma which crystallized as anorthosite. Miller claims that dykes of anorthosite are to be found in the region, and he has since described dykes of "nearly pure plagioclase-anorthosite" in Los Angeles County, California. F. L. Lessing is also convinced of the existence of anorthosite magma and claims that the occurrence of gabbro with anorthosite and pyroxenite bands makes it necessary to admite the existence of anorthosite and pyroxenite in the fluid state. K. S. Fowler maintains that the anorthosite of the Laramie Mountains in its earliest stages, "consisted of a viscous magma in the process of crystallization and intrusion." J. M. Harrison has found inclusions of crystalline limestone, surrounded by reaction-rims, in the Wilkinson anorthosite of Ontario. J. B. Mawdsley has described a body of andesine-anorthosite in the St. Urbain district of Quebec which surrounds a central body of labradorite-anorthosite and encloses isolated blocks of it. This evidence clearly indicates fluidity in the andesine-anorthosite; yet as Bowen (2) points out, it is not necessary that it should have been entirely or even largely liquid. It might have been a mush of andesine crystals with 20 or 30 per cent of residual liquor from which orthoclase and hypersthene (of which the rock contains 10-15 per cent) crystallized at a later stage.

Bowen certainly seems to have the better of this argument, and until somebody discovers a lava with the composition of anorthosite it will be difficult to believe in the existence of a completely liquid plagioclase magma. The foliation of many anorthosites, and the protoclastic microtexture which is so characteristic of many of them, are more easily explained by movement of a mush of liquid and crystals.

THE OCCURRENCE AND GENESIS OF BASALT

We have seen that the extrusion of rhyolite, trachyte or even andesite is not necessarily connected with the intrusion of plutonic rocks of the same composition. But the extrusion of basalt is so clearly connected with the intrusion of sills and dykes of diabase, with which it is virtually identical in composition, that there can be no doubt of their common origin. Where the Parana basalts have been removed by erosion, in southern Brazil, the underlying sediments are cut by extraordinary numbers of dykes and sills of diabase (C. L. Baker). The same relation is seen in South Africa, the magma that formed the Stormberg basalts having also given rise to the Karoo dolerites. In the same geological period, extrusion of basalt and intrusion of diabase went on together in India and in the eastern United States, where the Watchung basalts are the surface expression of the same magma that formed the Palisade diabase underground.

The quantity of lava emitted in fissure eruptions is enormous. In the region that is now the Deccan, in India, 200,000 square miles of the surface were flooded with basalt to a maximum depth of 6,000 feet. The Columbia River basalts of Idaho and Oregon are spread over an even greater area and range from two thousand to four thousand feet in thickness. According to C. L. Baker, the area covered by basaltic lavas in the Parana basin, South America, may be as great as the combined areas of the Deccan and Columbia River lava fields. Assuming an average thickness of 1,000 feet, Baker computes the volume of basalt extruded in this region to have exceeded 50,000 cubic miles.

These "plateau basalts" are in the main saturated rocks, although individual flows may contain either olivine or free silica. According H. S. Washington the Deccan basalts are typically free from olivine and hold a small excess of silica. C. L. Baker reports that the lavas of the Parana basin range in composition from andesite and olivine-free basalt to limburgite with abundant olivine. The mineralogical composition of the Deccan basalts, calculated from Washington's data, is roughly as follows:—quartz 2-5, orthoclase 5-7, labradorite 40-50, pyroxene 30-40, iron ores up to 12 per cent. Quite similar data are given for the Columbia River basalts and for the basalts of Iceland and the Faroe Islands (A. Holmes).

The question arises whether a magma that is so abundant, so widespread, and so uniform in composition is a primary one, or whether it has itself been derived from something still more fundamental. The only other magma that could possibly come into question is the olivine-basalt magma that appears to underlie the great oceans. This magma has a wider distribution than any other, for the ocean basins cover two-thirds of the earth's surface, and olivine rocks are widely developed on the continents too. W. Q. Kennedy has discussed this question and has reached the conclusion that the plateau-basalt magma and the olivine-basalt magma exist independently, neither being a derivative of the other. Yet he admits that the plateau-basalt magma is absent from the ocean basins and that it "seems to be connected in some manner with the presence of the granitic crust." Thus there is at least a suspicion that the plateau-basalt magma may be derived from the olivine-basalt magma by assimilation of silica and alkalis from the granitic crust. This view has been maintained by T. Tomita (2), but H. Kuno thinks the same result may be reached without assimilation, in consequence of the sinking of crystals of olivine and lime-rich plagioclase from a magma of the composition of eucrite or olivine-anorthite gabbro. It is permissible to suspect that olivine-free basalt has been derived from the same magma as olivine-basalt, but the details of the process are not yet understood.

THE OCCURRENCE AND GENESIS OF PERKNITE

This name was introduced by H. W. Turner to describe rocks composed almost wholly of pyroxenes and amphiboles, with iron ores and accessory minerals. The main subdivisions are pyroxenite and hornblendite, which explain themselves. Among the pyroxenites it is useful to distinguish the *ortho-pyroxenites,* which contain enstatite, bronzite or hypersthene, from the *clino-pyroxenites* which contain diopside, augite or diallage as their principal pyroxene. It will be shown that these two kinds of pyroxenite have generally been formed in different ways.

Among the *clino-pyroxenites* we may notice those described by W. C. Brögger in the parish of Gran, Norway. There are in this parish several small bosses of essexite or orthoclase-gabbro, a rock composed essentially of orthoclase, labradorite and augite, with subordinate hornblende, olivine and biotite. About the margins of these bosses the augite has become concentrated at the expense of the other minerals, giving rise to a rock that consists to the extent of 95 per cent of augite, with only trifling amounts of hornblende, biotite and labradorite. Comparing the chemical composition of the pyroxenite with that of the essexite, it is apparent that there has been a diffusion of magnesia and lime towards the periphery and of alumina and alkalis in the opposite direction. The pyroxenite was not formed from a magma of its own composition.

The association of pyroxenite with essexite and syenite has been described by G. A. Young at Mount Yamaska, Quebec. All three rocks contain the same dark minerals, but in the pyroxenite clino-pyroxene and hornblende have been concentrated to such a degree that feldspar is no more than an accessory constituent. Clino-pyroxenite is associated with diorite and monzonite at Monzoni, Tyrol, and there, too, it is evident that the pyroxenite was not formed from an independent magma.

Pyroxenites formed mainly of diopside are not uncommon where a siliceous magma has come into contact with limestone. Numerous instances are to be seen at granite-limestone contacts in Ontario. A very interesting example has been described by

Cross and Larsen at Iron Hill, Colorado; in addition to diopside the pyroxenite contains perovskite and frequently much apatite and biotite. A similar occurrence of diopside-pyroxenite containing much apatite was described from a granite-limestone contact zone at Palabora, Transvaal (S. J. S., 4). In each of these instances it seems that the pyroxenite was formed either by direct reaction between magma and limestone, or else by the soaking of magmatic solutions containing silica and phosphoric acid into the limestone.

Many pyroxenites contain a good deal of hornblende and show a transition to the *hornblendites*. J. Phemister describes a diopside-pyroxenite near Loch Ailsh, Scotland, which contains both hornblende and biotite. In this rock the crystals of hornblende typically enclose diopside; and in the same neighbourhood there are bodies of massive hornblendite composed of the very same hornblende, but the crystals contains cores or remnants of diopside, showing that the rock was originally pyroxenite. This is probably a case of deuteric action. Similar cases were discussed in Chapter X, which deals with post-magmatic reactions.

Between the clino-pyroxenites and the ortho-pyroxenites there are intermediate types containing both kinds of pyroxene. G. H. Williams has recorded such rocks in association with the gabbros of Maryland, either diallage or diopside appearing in conjunction with hypersthene. Williams makes the observation that in these rocks "the cohesion between the grains is so slight that perfectly fresh specimens may sometimes be rubbed to powder in the fingers." The absence of interlocking among the grains is suggestive of the origin of the rock by accumulation of crystals. Another rock of this kind occurs near Webster, North Carolina, and Williams introduced the name *websterite* for the whole group.

The group of *ortho-pyroxenites* has a much greater quantitative importance than the clino-pyroxenites. Varieties of "bronzitite," composed of bronzite with only trifling amounts of diopside and chromite, play an important part in the lower half of the great lopolith of the Bushveld, Transvaal. P. A. Wagner and A. L. Hall have described thick sheets of bronzitite, interstratified with every possible variety of norite

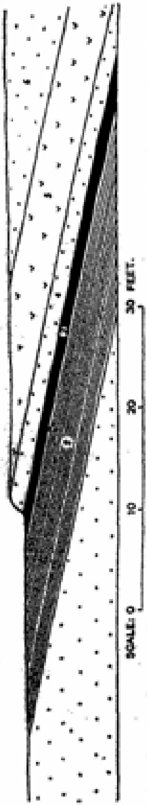

FIG. 40. Section showing the differentiation of the Bushveld norite in the eastern Transvaal. (P. A Wagner.)

1. Medium-grained spotted Norite.
2. Feldpathic pyroxenite, merging upward into
3. Pyroxenitic norite, with sulphides.
4. Light coloured spotted Norite.
5. Mottled Anorthosite.
6. Coarse-grained spotted Norite.

and anorthosite and containing seams of chromite. This alternation of feldspathic and pyroxenitic rocks persists through more than 6,000 feet, and single sheets of bronzitite may be several hundred feet thick. To explain this remarkable sheeted structure, which closely simulates the bedding of sedimentary rocks, Wagner has supposed that the crystals that were earliest to form in the noritic magma, namely those of olivine and chromite, sank on account of their great density and accumulated at lower levels, where the olivine was converted partly or wholly into bronzite. B. V. Lombaard believes rather in the direct settling of pyroxene crystals from the magma. He shows that the Bushveld "norite" contains ortho-pyroxene in its lower part, but clino-pyroxene in the upper part; also that the pyroxenites in the lower part are all of the ortho-type, whereas the few pyroxenites known in the upper or gabbroic zone are clino-pyroxenites.

This view, that the ortho-pyroxenites were formed by crystal settling, is strongly supported by experimental chemistry. Bowen and Schairer have shown that bronzite in the ordinary orthorhombic form is only stable below 1100° C., whereas a melt of the same composition crystallizes in the form of clinobronzite in the neighbourhood of 1500° C.; consequently common bronzite cannot possibly have crystallized from a magma of its own composition.

Ortho-pyroxenite similar to that of the Bushveld complex occurs in great thickness in the Great Dyke of Rhodesia, in the Stillwater complex, Montana, and in the Bay of Islands region, Newfoundland, always in a very similar petrological association.

Some pyroxenites contain garnet. A. Lacroix described dyke-rocks consisting of diopside and garnet, diallage and garnet, and diallage with hornblende and garnet, at Lherz in the Pyrenees. The name "rodingite" was given by P. Marshall to a dyke-rock in Nelson, New Zealand, which consists of grossularite and diallage. This rock was at first thought to be a primary eruptive product, but L. I. Grange has since demonstrated that it is an altered gabbro. It may be that the rocks described by Lacroix are also metamorphic products.

Much has been written about the nature of *eclogite*. This very striking rock consists typically of green omphacite (a more or less jadeitic diopside), together with pink garnet intermediate in composition between almandine and pyrope. These rocks occur as bands and lenses among crystalline schists, especially in association with hornblende-gneiss and amphibolite. P. Eskola has claimed an eruptive origin for the eclogites of Norway, but most students of the subject consider eclogite to be a metamorphic rock. G. Switzer showed that the eclogite of California is associated with basalts from which the eclogite was formed by hydrothermal contact metamorphism caused by serpentine intrusions. Even ortho-pyroxenite sometimes contains garnet, as in the "griquaite" of the Kimberley diamond mines, which is largely composed by enstatite and pyrope. The origin of this rock is obscure.

A special problem is presented by rocks which consist almost entirely of oxides of iron, titanium and chromium. Among the gabbros and norites in many parts of the world there are dyke-like and sheet-like bodies of nearly pure ore, which behave as if they were independent intrusions. C. F. Kolderup has described an ilmenite-rock at Ekersund, Norway:

W. Petersson, magnetite-spinel rocks in Sweden; A. L. Hall, thick sheets of titaniferous magnetite in the Transvaal; and Watson and Taber, dykes of ilmenite, rutile and apatite in Virginia. The difficulty about regarding these as eruptive rocks is that the melting points of magnetite, ilmenite, and spinel lie far above those of silicate rocks. If the ores are sufficiently diluted with silicates, the difficulty disappears on account of the mutual lowering of freezing points. E. Sampson has studied especially the chromite deposits of the Bushveld complex and other localities, and favours the view that some of the chromite was formed from residual magmatic solutions at quite low temperatures. The writer (S. J. S., 5) thinks it probable that some of these iron ores were formed by self-oxidation of ferrous hydroxide contained in residual solutions of magmatic origin.

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> *The fashion in which we think changes like the fashion of our clothes, and it is difficult, if not impossible, for most people to think otherwise than in the fashion of their own period.*—G. B. Shaw.

CHAPTER XVII

THE UNDERSATURATED ROCKS (NON-FELDSPATHOIDAL DIVISION)

OCCURRENCE AND GENESIS

When a magma contains too little silica to convert all the metallic oxides into saturated compounds, then one or more of these oxides must go short. In the commonest case it is magnesia that is affected, and olivine is formed instead of pyroxene. The resulting rocks are olivine-gabbro and olivine-basalt, which are perhaps the most abundant rocks in the outer one-tenth of the earth's radius; as well as peridotite and less common types. Much less frequently the deficiency of silica falls upon the soda, and nepheline or sodalite may be formed, as in foyaite and phonolite. Potash is rarely affected by any deficiency of silica, so leucite-bearing rocks have a very limited distribution. Lime is still more rarely found in unsaturated compounds. Melilite and its constant companion, perovskite, are only formed in rocks of extremely low silica content, in which most of the magnesia and soda have been reduced to olivine and nepheline, and even potash has gone into leucite or mica. Melanite (lime-ferric garnet) is another lime-mineral that is occasionally developed in undersaturated rocks, mostly in association with nepheline or leucite, though the condition for its formation is perhaps a deficiency of alumina rather than silica. Finally, if the deficiency of silica falls upon alumina, we get the very rare group of corundum pegmatites.

The Occurrence and Genesis of Olivine Rocks

The largest bodies of undersaturated magnesian rocks are those in which feldspar predominates and olivine plays quite a minor part. One of the largest intrusive bodies of this character is the lopolith of olivine-gabbro at Duluth, Minnesota, which has an area of 2,400 square miles and a probable thickness of 20,000 feet. This great mass of rock varies from normal gabbro and anorthosite to olivine-gabbro, troctolite and peridotite, but according to F. F. Grout, over two-thirds of the mass is olivine-gabbro. Another great body of olivine-gabbro and olivine-norite forms the peninsula of Sierra Leone, West Africa. The area occupied by this rock in the peninsula alone is nearly 200 square miles, and the same kind of rock reappears in the Banana Islands, showing that the norite continues far out to sea. From the orientation of the primary banding, N. R. Junner concludes that this intrusion, too, has the form of a lopolith or basin-shaped mass. A full petrographic description of the norite of Sierra Leone has been given by F. Dixey.

The East Griqualand sheet in South Africa, to which we referred on page 280, is mainly composed of olivine-gabbro, but it has an acid roof-facies which D. L. Scholtz describes as quartz-monzonite. This passes down into gabbro, which soon becomes olivine-bearing; the content of olivine increases rapidly towards the base of the sheet, and the basal zone consists of alternations of picrite, troctolite, and olivine-rich hypersthene gabbro (*Fig*. 41).

In each of these three instances the bulk of the intrusive is made up of olivine-gabbro or olivine-norite of fairly constant character, but there is a tendency towards the formation of bands of more feldspathic and more melanic character respectively, which in extreme instances become distinct sheets or lenses of anorthosite on the one hand and peridotite on the other. This banded character is very strongly developed in the Stillwater complex, Montana, another lopolith with an outcrop thirty miles long and an exposed thickness of 11,000 feet. The upper part of the mass is mainly norite and anorthosite; the lower part contains a great thickness of bronzitite, peridotite and dunite in bands that range from a few inches to hundreds of feet in thickness. One section was measured in which there were 77 bands within a range of 200 feet (Howland, Peoples and Sampson).

In the Bay of Islands district, Newfoundland, four layered intrusions of the same character as the last lie in a belt 65 miles long and ten miles wide. The southern pair were described by J. R. Cooper, the northern pair by E. Ingerson. There is an unfortunate difference of opinion about the geological structure of the belt, but all four masses show the

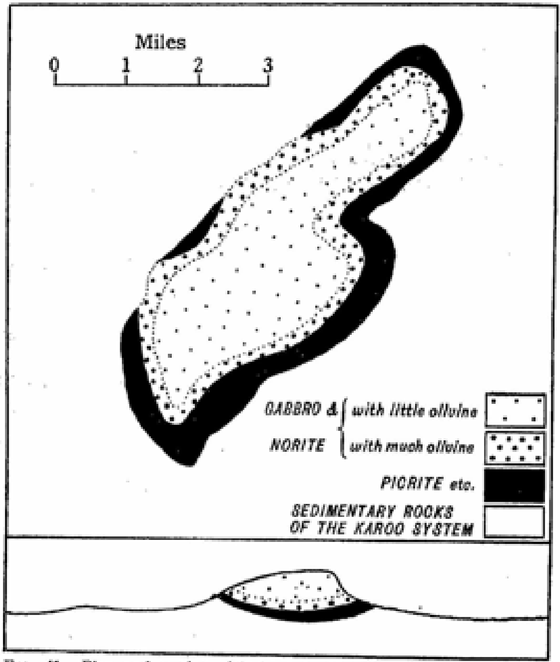

FIG. 41. Plan and section of Tabankulu Mountain, Pondoland, South Africa. (After A. L. du Toit, modified by D. L. Scholtz.)

same remarkable alternation of bands that vary in composition from gabbro to peridotite. According to Ingerson the Table Mountain mass contains some 7,000 feet of olivine-gabbro, anorthositic gabbro, and bytownite-peridotite, overlying 14,000 feet of bronzite-dunite and bronzite-peridotite.

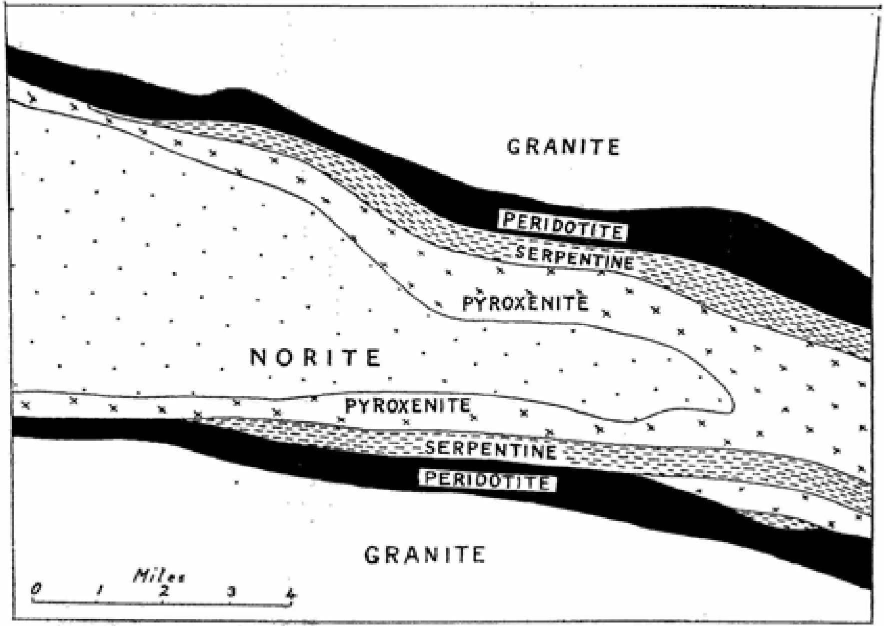

FIG. 42. Plan of a portion of the Great Dyke of Rhodesia. (After B. Lightfoot.)

The remarkable intrusion known as the Great Dyke in Rhodesia has an outcrop more than 300 miles long, with an average width of only five miles. From the form of the outcrop one would consider the intrusion to be a vertical or highly inclined sheet; but the strongly banded rocks which make up the sheet are not arranged parallel to the walls; they are described by B. Lightfoot as forming "a series of large discontinuous cakes, each of which has a gentle synclinal structure." The rocks vary in composition from norite to bronzitite, harzburgite, and serpentinized peridotite, sometimes one and sometimes another predominating in the outcrop of the Dyke (*Fig.* 42).

A classic instance of a composite intrusion consisting of alternate feldspathic and peridotitic layers was described by A. Harker in the island of Rum, Scotland. In Allival Hill there are nineteen nearly horizontal layers of peridotite and allivalite alternately, the former consisting of olivine with only a little anorthite, and the latter of anorthite and olivine in roughly equal amount. Harker thinks that the sheets, which are 50 to 150 feet thick, represent distinct intrusions of heterogeneous magma, consisting of more feldspathic and more peridotitic portions which did not mingle freely. Later students of banded intrusions have tried to explain the phenomena in terms of crystal settling, but neither explanation gives complete satisfaction. If the bands were always horizontal or nearly so, then crystal settling might provide a general solution, but there are certain intrusions in which the differentiation takes place from margin to centre.

At several localities in the Ural Mountains, bodies of peridotite and dunite occur within larger bodies of gabbro or norite (*Fig.* 43). For instance, at Koniakowski Hill in the district of Nicolai-Pawda an elliptical body of dunite measuring 1 x ¾ km. lies within an elliptical pyroxenite mass which is itself completely surrounded by melanocratic gabbro, passing out into normal gabbro (Duparc and Grosset). Duparc maintains that the rocks of these concentric zones, from dunite to gabbro, are all differentiation-products of a single magma. In describing a similar occurrence at Solowiewa Hill, in the Tagil district, R. Spring seems to favour another view of the matter, for he writes that "the appearance is as if the peridotite was a younger, stock-like mass which was intruded after the consolidation of the diorite" (or gabbro).

In support of the view expressed by Duparc, we may note that at Susimäki, Finland, M. K. Palmunen has described a hornblende-gabbro which passes centrally into a peridotite composed of olivine, hornblende, pyroxene and ore, and this again into an ore-bearing dunite containing 60 per cent of olivine and 36 per cent of iron ore. An occurrence similar to this has long been known at Taberg, Sweden.

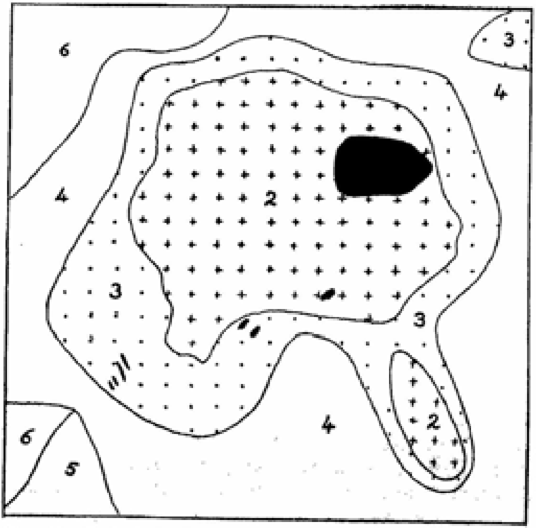

FIG. 43. Geological plan of the olivine rocks of Koniakowsky Hill, district of Nicolai-Pawda, Russia; after Duparc and Grosset. Black, dunite; 2, pyroxenic peridotites; 3, melanocratic olivine-gabbro; 4, normal olivine gabbro; 5, norite; 6, diabase. Scale, 1 : 100,000.

It is interesting to find, in the Thetford area, Quebec, examples of both types of differentiation. A series of basic intrusive masses, comprising the usual association of gabbros and peridotites, forms a belt about 70 miles long. The intrusives include both sills and stocks and within them the rocks are arranged in order of decreasing density; in the sills from the base upwards, in the stocks from the centre outwards (J. A. Dresser). One suspects that gravity must be the cause of the differentiation in both cases, but the structure of the "stocks" does not seem to be known. C. Camsell says that the stock at Tulameen, B.C., "resembles a hard-boiled egg, the peridotite being equivalent to the yolk and the pyroxenite to the white of the egg." His explanation of the structure is that pyroxene crystallized before olivine, hence pyroxenite was formed along the walls of the intrusion, where cooling was rapid, leaving a concentration of olivine in the central magma.

From these great lopolithic masses and belts in which peridotites are associated with feldspathic gabbros and anorthosites, it is only a step to another series of intrusions composed almost wholly of hypermelanic rocks.

The Great Serpentine Belt of New South Wales is a body of olivine-rocks and serpentine which has been intruded along the plane of a powerful reversed fault (W. N. Benson). The outcrop is not continuous, but many intrusive bodies from a hundred yards to thirty miles long crop out at intervals along the fault line through a total distance of more than two hundred and fifty miles. The width of the belt varies from a few inches to nearly two miles. There can be no doubt that all these detached outcrops are connected underground, and that the whole is essentially one great dyke. The dominant rock is harzburgite (olivine-enstatite rock), but this passes locally into olivine-rock and enstatite-rock, while amphibolite, anorthosite and olivine-gabbro play quite a minor part in the complex. All the magnesian rocks are more or less serpentinized.

It is now recognized that belts of serpentine and related rocks follow the flanks of most of the great mountain chains and island arcs. In eastern North America a chain of these rocks runs parallel to the Appalachian axis from Newfoundland to Alabama. Another great serpentine belt extends across Guatemala into Cuba, Hispaniola and Puerto Rico; here it swings southward and H. H. Hess (1) thinks that it makes a complete loop and joins the serpentine belt in the Cordillera of Venezuela and Colombia. In Cuba, according to Hess (2), a single mass of peridotite has an outcrop of probably 5,000 square miles. This may be the largest body of peridotite in the world, unless it is surpassed by the astonishing development of similar rocks in New Caledonia. The most extensive chain of peridotite and serpentine is the one that follows the entire course of the Alpine folds across Europe into Asia Minor, and reappears in Burma to follow the East Indian arc through the Andaman Islands, Sumatra, Timor, Celebes, Borneo and the Philippines. A most valuable discussion ot the serpentine belts of the world and their tectonic significance was given by W. N. Benson in 1926, and H. H. Hess (1) has carried the matter to a further stage by linking the serpentine belts to belts of negative gravity anomalies.

Before leaving this subject we may just mention the serpentine belt of Dun Mountain, New Zealand, which has given us the name "dunite" for a rock composed almost entirely of olivine. This is a belt about 20 miles long and up to three miles wide, composed of serpentine with subordinate dunite and harzburgite; it follows the strike of a series of highly tilted limestones and shales, and it is separated from the limestone by a marginal zone of diabase. These rocks were mapped and described by Bell, Clarke and Marshall (*Fig.* 44).

Still one more mode of occurrence of olivine-bearing rocks must be mentioned. In this case the olivine rock occupies more or less vertical pipes which apparently represent the throats of former volcanoes. The best known examples are the "kimberlite" pipes of South and Central Africa. One of the largest of these pipes, the Premier Diamond Mine, near Pretoria, is 2,800 feet in greatest diameter. Every pipe has the form of a more or less symmetrical, steep-sided funnel, there being a steady decrease in size with increasing depth. Although a connection between individual pipes is not often apparent at the surface, yet mining operations have revealed that in many cases the pipes are arranged along dykes, being sometimes mere enlargements or "blows" in the course of the dyke, while in other cases they represent points where an old channel of eruption, marked by a dyke, has been reopened during a younger phase of explosive volcanic activity. The Newlands Diamond Mine, for example, consists of four small pipes arranged along a narrow dyke; the Voorspoed mine was discovered by following up a dyke; and the linear arrangement of the pipes in the Belgian Congo is evident on P. A. Wagner's map.

The typical rock of these diamond pipes, the "blue ground" or kimberlite, is a breccia consisting of fragments of many kinds of rocks enclosed in a matrix of serpentine and carbonates. While it is certain that much or all of this serpentine was formed from olivine, it is not correct to describe the rock as a serpentinized peridotite. There is a certain amount of evidence which suggests that it may have been a melilite-basalt. This view, originally put forward by H. C. Lewis, has been revived by M. S. Taljaard and the writer. Dykes of serpentine that closely resemble kimberlite occur in New York and Pennsylvania, and C. H. Smyth was able to identify melilite in one of them.

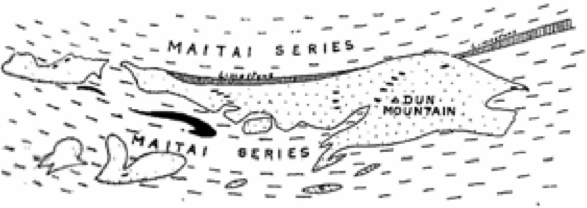

FIG. 44. The peridotite intrusions of Dun Mtn., New Zealand. (After Bell, Clarke and Marshall.) The surrounding rocks are shales and greywackes of the Maitai Series (Trias-Jura), with a thick limestone, from which the peridotite is separated by a marginal zone of diabase. Black areas are rodingite (grossularite-diallage rock). Scale, 4 miles to 1 inch.

Before we discuss the significance of these observations it is necessary to devote a few lines to the *extrusive* olivine-rocks. The first point to be noted is the entire absence of any lavas corresponding in composition to dunite or any of the varieties of peridotite. The only important type of olivine-bearing lava is olivine-basalt, which corresponds in composition to olivine-gabbro and olivine-diabase. The second point is the universality of olivine-basalt, which is not only one of the commonest of lavas on land but is virtually the only kind of lava developed in the vast area of the great oceans. One may reasonably conclude that olivine-basalt is not only the commonest of all extrusive rocks, but that it is more abundant than all other lavas put together. The third point to which we would direct attention is the extreme range of variation in the olivine-content of these lavas. The average olivine-basalt holds some 10 to 15 per cent of this mineral, but in the oceanic regions particularly one finds varieties with 50 and sometimes as much as 75 per cent of olivine. In these "oceanites" or picrite-basalts the olivine crystals are exceptionally large in comparison with the associated minerals, and there can be no doubt that these lavas represent portions of magma that have been enriched with olivine crystals either by crystal settling or by convection currents.

We must now put all the evidence together and see what generalizations we can make regarding the olivine-basalt magma. From the world-wide distribution and the great abundance of olivine-bearing rocks we are entitled to conclude that if there is such a thing as a fundamental world-magma from which all other magmas and eruptive rocks have been derived, it must have the composition of olivine-basalt. The evidence cited in this chapter and in Chapter IX compels us to believe that the differentiation of such a magma, whether effected by the settling and flotation of crystals or by the expulsion of residual fluid during freezing, is capable of furnishing most kinds of rock from dacite and trachyte to anorthosite and dunite. The potassic rocks alone seem to stand apart from the olivine-basalt line of descent.

The absence of lavas of peridotitic composition points to the non-existence of peridotite magma. The manner of formation of peridotite has generally been the settling of crystals of olivine and pyroxene from a body of olivine-basalt magma. The peridotites are *accumulative* rocks rather than magmatic rocks, just as the anorthosites and many perknites are. Picrite-basalt on the surface, and picrite underground, represent stages in the accumulative process; and it is clear that the farther the sinking of olivine has proceeded, the less likely it is that the olivine-enriched magma will ever reach the surface; hence the scarcity of picrite-basalt and the entire lack of lavas consisting wholly of olivine or olivine and pyroxene.

But what are we to say of the great masses of serpentine and peridotite that are not associated with gabbro or any other feldspathic rock? H. H. Hess (2) has shown that some of these masses actually exhibit chilled margins against the country rocks, so they must be intrusive. An important point about these intrusive bodies, which is generally obscured by the habit of calling them "peridotite," is that they are always more or less deeply serpentinized. A content of more than 10 per cent of water is common. Pure serpentine holds only 13 per cent of water. Another significant point is that even the greatest masses of serpentine have effected very little contact metamorphism of the rocks among which they were intruded, consequently they were not very hot. Consideration of the melting points of forsterite and fayalite will show that an olivine-magma, even if its freezing point were lowered considerably by small amounts of other components, could scarcely exist at a temperature lower than 1200° C. To reconcile these observations one must believe that the serpentine intrusions were not formed from an olivine magma, but from a serpentine magma, that is to say, a serpentine gel containing crystals and fragments of olivine. Such a gel may be formed by hydration of a peridotite of the accumulative type, and it may be set in movement by the disturbances of orogeny. The concept of a serpentine magma is favoured by H. H. Hess (2), R. A. Daly, and W. F. Gisolf. R. B. Sosman speaks of "a mass of crystals lubricated by very thin films of liquid which is not a silicate melt, but is simply water or a water solution"; and T. du Rietz thinks the peridotites of northern Sweden were intruded "in a semi-crystalline condition."

If a serpentine magma can be intruded underground, would it not sometimes make its way to the surface? Hess has considered this point and thinks that the high water content of such a magma would cause it to be erupted explosively, without forming lava-flows. The invariably brecciated character of kimberlite certainly falls into line with this suggestion.

A. N. Aleshkoff differs from all other students of the subject and expresses the opinion that "pure molten olivine formed dunites." He has described an "uninterrupted wall" of peridotite and dunite which extends for 125 miles between latitude 65° 30′ C. and latitude 67° C. in the Polar Ural. His map shows that the peridotite is bordered by zones of pyroxenite, gabbro, and diorite, just as so many other peridotites are in the Ural and elsewhere. Since the rocks are the same and the association is the same, one cannot help feeling that the argument against the existence of an anhydrous olivine magma must apply to this region just as to all others. The latest student of Uralian peridotites, N. V. Lodochnikov, believes the magnesian magma was a hydrous one and that the olivine crystals which separated from it at an early stage were serpentinized at a later stage by water concentrated in the residual magma.

French petrographers have claimed that peridotite may be developed in consequence of the assimilation of magnesian limestone by granitic magma. Among the products of reaction between granite and limestone in the Pyrenees, A. Lacroix describes a complete series of rocks ranging in composition from hornblende-granite to hornblende-peridotite; the last a rock consisting of large crystals of hornblende which enclose numerous rounded or irregular grains of olivine, closely resembling the well known peridotite of Schriesheim, Germany. This view was supported by M. Longchambon. W. N. Benson has pointed out that if all peridotites were formed in the way Lacroix describes, it would be difficult to account for the relatively high proportions of nickel and chromium oxides which they contain, for these oxides are almost never found either in granite or in limestone.

THE OCCURRENCE AND GENESIS OF ROCKS CONTAINING UNSATURATED LIME MINERALS

Both melilite and melanite are so frequently associated with nepheline or leucite that one suspects a common factor in the genesis of all four minerals.

Melilite-basalt is among the rarest of rocks. Melilite itself is rich in lime and poor in silica, so it is not surprising to find among the melilite-basalts both the highest recorded percentage of lime (27 per cent) and almost the lowest recorded percentage of silica (30·5 per cent) in any eruptive rock. This suggests that assimilation of limestone may have played a part in the genesis of these rocks. From the experimental side this idea is supported by the common occurrence of melilite in basic slags, and by the demonstration by A. S. Ginsberg and K. S. Nikogosyan that the addition of calcium carbonate to melted diabase causes melilite to crystallize out.

There is in addition much field evidence to connect melilite rocks with limestone. The original alnöite (biotite-melilite-basalt) takes its name from Alnö, Sweden, where there is a most intimate mingling of silicate and carbonate rocks (see on, page 324). E. Becker has shown that the distribution of melilite-basalt in south-western Germany is related to the distribution of a thick Jurassic limestone. The melilite rocks of the Polzen district, Bohemia, are also in a region of thick limestones, and K. H. Scheumann is definitely of the opinion that assimilation of limestone played a part in their formation. The uncompahgrite of Iron Hill, Colorado, and the okaite of Quebec, two rocks which are made up to the extent of two-thirds of melilite, are both found in contact with crystalline limestone (Cross and Larsen; J. Stansfield). J. T. Singewald and C. Milton have described a pipe of alnöite near Avon, Missouri, which cuts dolomite and penetrates it in the most intimate fashion, forming dark streaks with indefinite boundaries that fade away in the dolomite. The turjaite of the Kola Peninsula is associated with "carbonatite" dykes, and E. H. Kranck indicates the necessity of "a supply of CaO from outside" in order to explain the abnormal composition of this rock. To complete the argument, C. E. Tilley has described the production of small bodies of nearly pure melilite at a contact between olivine-dolerite and chalk in Antrim, Ireland (1), and a similar occurrence between gabbro and limestone on the island of Muck, Scotland (2).

On the other hand, there are certain occurrences of melilite-basalt in regions where no limestone is known. C. S. Ross has described an alnöite at Winnett, Montana, far away from any known limestone. The melilite-basalts of South Africa are all in regions where there is little known limestone.

In the district of Sutherland the melilite-basalts are intrusive in the shales and sandstones of the Karroo system, but it is not possible to say what underlies these strata. In the district of Namaqualand there is perhaps the greatest development of melilite-basalt in the world, and it lies upon the old gneiss which locally contains only insignificant lenses of limestone. The melilite-basalt here forms hills up to a hundred feet high and half a mile long, which have been interpreted as the outcrops of vertical pipes (A. W. Rogers), but also as remnants of sills (M. S. Taljaard), the rubbly nature of the outcrops making it difficult to decide between the two alternatives. In view of the evidence described above, from other localities, and in view also of the utterly abnormal composition of melilite-basalt, the writer thinks it justifiable to assume the existence of a deep-seated body of limestone in this region. Only the addition of lime or the subtraction of silica can convert ordinary basaltic magma into melilite-basalt, and reaction with limestone is the only obvious way in which both of these changes can be effected.

There are rare and quantitatively insignificant occurrences of melilite-basalt on some of the oceanic islands, for instance, Oahu, Hawaii (W. Cross). The only possible source of limestone in this case is coral reef, and it is a curious coincidence that the melilite-basalt at Moiliili rests on reef limestone. There is no indication of reaction at this point, nevertheless it seems impossible to account for the appearance of melilite in such an environment without the participation of limestone at some stage in the process.

N. L. Bowen has found that melts of nepheline and diopside, in the crucible, give olivine and melilite at temperatures between 1100° and 1300° C. He argues therefore that the presence of melilite is not necessarily an indication of contamination of the magma by limestone. But if the nepheline-diopside magmas in nature are themselves produced in consequence of assimilation of limestone, as we shall see in the next chapter, then Bowen's contention loses its force.

As regards *melanite* and *andradite*, the lime-ferric garnets, the evidence is still more complete that assimilation of limestone is an essential condition for their formation. These two names indicate the same species, but andradite is brown in colour, may contain some alumina, and is nearly free from titania; whereas melanite is black, is nearly or quite free from alumina, and commonly holds five or ten per cent of titania. In the amorphous variety called schorlomite the content of oxides of titanium may exceed 20 per cent. Of these varieties melanite is commonly found in the eruptive rock, andradite either in the eruptive or on the limestone side of the contact.

At the following localities melanite is developed in an eruptive rock at or about its contact with limestone:—Loch Borolan, Scotland; Fen, Norway; Alnö, Sweden; Kuolajärvi, Finland; Turja, Kola; Ditró, Rumania; Monte Somma, Italy; Ice River, B.C.; Franklin, B.C.; Magnet Cove, Arkansas; Potash Springs, Arkansas; Iron Hill, Colorado; Bearpaw Mts., Montana; Libby, Montana; Judith Mts., Montana; Kishengarh, India; Mount Elgon, Kenya; Karamoja, Uganda; Nyassaland; Sekukuniland, Transvaal; Ambavatoby, Madagascar; Port Cygnet, Tasmania. Both melanite and andradite occur at Mount Royal, Quebec. (See Plate IV.)

At the following localities the garnet is andradite rather than melanite:—Sviatoy Noss, Transbaikalia; Haliburton-Bancroft, Ontario; Marble Bay, B.C.; Stockton, Utah; Pouzac, Pyrenees; Hook batholith, Rhodesia; Zeekoegat, Transvaal; San Carlos Mts., Mexico.

Melanite is also found in phonolite or nephelinite at a few places, of which the following are the best known:—Frascati, Alban Hills, and Monte Somma, Italy; Kaiserstuhl and other localities in Baden; Rieden, Olbrück, and other localities in the Eifel. Each of these is a locality where thick limestone is present. Melanite has also been found in phonolite and nephelinite of Mount Elgon, East Africa, and in quantity on Rusinga Island, Lake Victoria, each time in the neighbourhood of a contact-metamorphosed limestone.

We have named above more than forty localities where the occurrence of a lime-garnet is connected with the presence of limestone; and the list could be extended. On the other hand, the writer knows of only five localities where melanite is found in rocks which cut crystalline schists, no limestone being known to occur near. One of these is Iivaara, Finland, where the surrounding country is drift-covered. The highly titaniferous black garnet occurs in a melanic rock composed mainly of diopside and nepheline—the original "ijolite"—and the garnet is intergrown with calcite. At Cevadaes, Portugal, a small lens of foyaite containing melanite lies concordantly in para-gneiss; but in other parts this gneiss contains lenses of "lime-silicate hornfels." A little melanite occurs in the foyaite of Cabo Frio, Brazil; an island formed of red gneiss. Again, A. Lacroix has recorded melanite quite abundantly in a micromonzonite on the island of Kassa, Los archipelago. He says the occurrence is unique in this archipelago, and adds that there is no limestone either in the islands or on the mainland of French Guinea. In a newly described area in Tanganyika (Eades and Reeve), melanite-syenite forms a few small tors in granite country; but a junction with the Basement Complex is near and the granite is full of xenoliths.

All that these few exceptions demonstrate is just our ignorance of the environment in each case. An eruptive magma does not produce new minerals as a conjuror produces rabbits, to amuse and bewilder us. Melanite is never formed without a definite chemical reason, and that reason, it is abundantly clear, is the assimilation of limestone by the magma. In the opinion of the writer, the necessary condition for the generation of melanite is so clearly established that this mineral may be used to prove assimilation of limestone just as surely as cordierite and sillimanite prove assimilation of aluminous sediment.

THE OCCURRENCE AND GENESIS OF CORUNDUM ROCKS

That some corundum may be developed in consequence of the assimilation of aluminous sediments by eruptive magma has been admitted by several writers. Thus K. Busz showed clearly that corundum had been formed in this way in certain dyke-rocks in Devonshire, and H. H. Thomas has studied the various stages in the transformation of aluminous rock-fragments into anorthite-corundum-spinel aggregates, within dolerite sills on the island of Mull, Scotland. L. V. Pirsson assumed that the sapphires in the lamprophyre dyke at Yogo Gulch, Montana, were formed from the clay slates through which the dyke has broken, and the sporadic occurrence of sapphire in the basalts of the Rhineland has generally been explained in the same way. But it is not clear that any such explanation is applicable to the widespread occurrence of corundum in North America and South Africa.

Corundum is found in small quantity in some of the syenites and nepheline rocks of Ontario, but it is especially abundant in the syenite-pegmatites, which form veins and dykes sometimes as much as eighteen feet wide. These very coarse-grained pegmatites are made up of pink, perthitic feldspar (soda-orthoclase) with corundum and a little biotite, muscovite, calcite, and exceptionally a very small amount of quartz. Corundum may make up as much as thirty-five per cent of this rock. The feldspar is not always an alkaline one, but varies from microperthite through albite, oligoclase and andesine to bytownite. A. E. Barlow particularly emphasizes the point that "the nepheline and related alkaline syenites with which the principal occurrences of corundum are associated, show an extreme and rapid variation in composition," although they form a single geological unit.

The important corundum reefs of the Transvaal are coarse pegmatites, which carry quartz and mica when they cut granite, and only become corundum-bearing when they enter a region of basic magnesian rocks (A. L. Hall). The reefs then consist of plagioclase (oligoclase or andesine) with corundum and biotite; orthoclase is rare; and Hall remarks that "not a single example was observed where a corundiferous rock also carried quartz." The surrounding rocks have their magnesian constituents changed into talc in the neighbourhood of such a reef, indicating an addition of silica.

The conditions in North Carolina, Georgia, Massachusetts, and in Plumas County, California, are almost identical with those in the Transvaal. In each case the mineral occurs in pegmatite veins about the junction between granite and a magnesian rock which is generally some variety of peridotite.

At Kyschtym, in the Ural Mountains, similar conditions recur; in this instance a large dyke of anorthite-corundum-biotite rock intervenes between granite and serpentine. In the same district there are also stock-like bodies of corundum syenite in the granite; in these the feldspar is microperthite, and J. Morozewicz notes that a few grains of quartz are very rarely to be seen. This locality is in the neighbourhood of Miask, where nepheline rocks have long been known to occur, so there would seem to be a close general resemblance between the corundum rocks of the Ural and those of Ontario.

In India, the corundum-bearing rock at Coimbatore, Madras, is once more a pegmatite, having microperthite and biotite as its silicate constituents. The geological relations are obscured by surface deposits but T. H. Holland shows that the distribution of corundum is definitely related to outcrops of nepheline-syenite.

In brief, the typical corundum-bearing rock in all parts of the world is a more or less coarse-grained pegmatite, which is typically associated either with nepheline rocks or with peridotite and serpentine. In the former case the corundum rock is a pegmatitic facies of the nepheline rock; in the latter case the corundum-bearing rock is intrusive in the peridotite and has reacted with it, forming biotite, actinolite and talc in the contact zone. This reaction is generally described as "desilication" of the pegmatite, and it does indeed involve the transfer of silica from the dyke into the peridotite; but it is impossible that desilication alone could produce a plagioclase-corundum rock from an average granite-pegmatite. K. A. Vlasov considers that fluorine and boron derived from the pegmatite magma act as catalysts and redistribute the cations between the acid magma and the basic rocks. E. S. Larsen rejects the desilication theory altogether (although it is true in the sense that the magnesian rocks have undoubtedly taken up silica) and claims the corundum-pegmatites as high temperature hydrothermal deposits. That the products in the contact zone (talc, chlorite, vermiculite) are hydrothermal is undeniable; yet neither the hydrothermal theory nor the desilication theory goes to the root of the matter, which is that the magma or solution that gave rise to the corundum rock must have held an altogether abnormal quantity of alumina, and for this there is no obvious source either in acid magma or in basic rock. In short, the problem of the corundum pegmatites is part of the problem of mineralized pegmatites in general, and is still far from solution.

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A theory is a tool and not a creed.—J. J. THOMSON.

CHAPTER XVIII

THE UNDERSATURATED ROCKS (FELDSPATHOIDAL DIVISION)

OCCURRENCE AND GENESIS

NEPHELINE-BEARING and leucite-bearing rocks are very scarce in comparison with other kinds, and most of the occurrences are relatively small. The largest known bodies of eucrystalline nepheline-bearing rocks are found in the Kola Peninsula, in Brazil, in Greenland and in the Transvaal. In Kola there are two masses of nearly circular outline which are so close together that they are probably connected beneath the waters of Lake Umpjaur (*Fig.* 45). Indeed, B. M. Kupletsky claims that the two together form "a synclinal intrusion approaching a lopolith in its form." The larger mass, which is known both by the Lapp name Umptek and the Russian name Chibina, has an area of 512 square miles; the smaller mass, Lujaur-Urt or Lovosera, 250 square miles; but in spite of its great area the complex seems to have no great thickness, possibly no more than 3,000 feet. The Umptek mass is built of concentric zones of various rocks, from syenite on the outside through many mineralogical and textural varieties of nepheline-syenite (foyaite) to ijolite and nepheline-apatite rocks. On the whole the content of silica appears to decrease, and the content of lime to increase, from the margin inwards. This great alkaline complex is intruded into and rests upon pre-Cambrian and early Paleozoic rocks, and it lies on the line of a powerful dislocation. The original description by W. Ramsay (in German) has been amplified in recent years by a number of papers (in Russian with brief English summaries) by A. Fersman and B. M. Kupletksky on Umptek, and N. A. Eliseev and others on Lujaur-Urt. The fullest account in English is that by A. A. Polkanov.

The two massifs of Itatiaia and Pocos de Caldas, in Brazil, have been briefly described by A. R. Lamego and O. Barbosa respectively. The former is estimated to cover 560 square miles, the latter 420 square miles, but the rocks are not all feldspathoidal and no detailed maps are available. A third occurrence, on the island of Sao Sebastiaio, occupies 115 square miles (R.O. de Freitas).

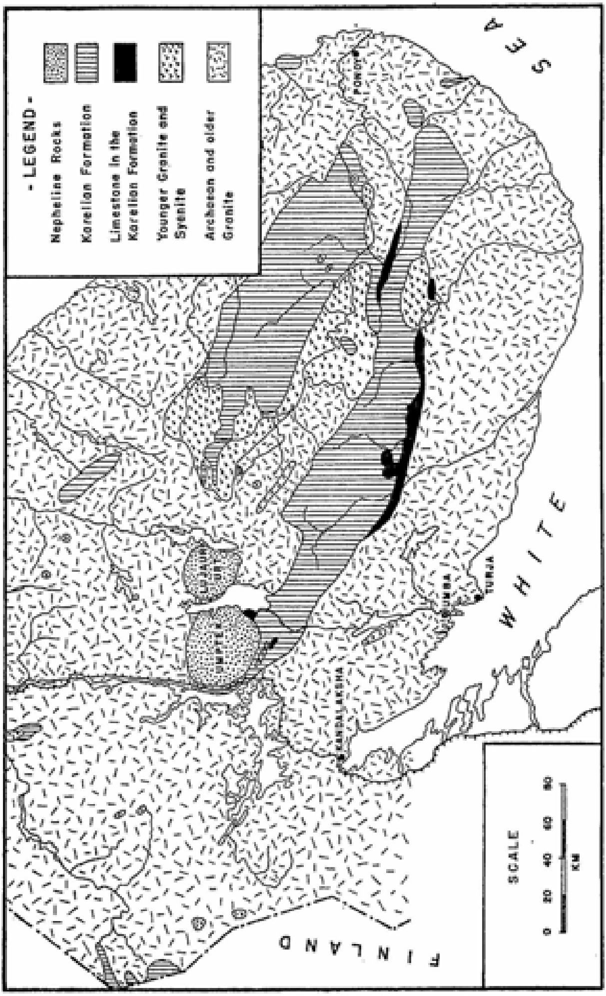

FIG. 45. Geological map of the Kola Peninsula. (After A. A. Polkanov.)

In the western Transvaal a circular group of hills known collectively as Pilansberg (Pilaan's Mountains) is composed of the same kinds of rocks and has the same concentric structure as Umptek. The resemblance extends even to the less common minerals such as pectolite, eudialyte, astrophyllite, lamprophyllite, lovchorrite. This complex, which is superimposed upon the Bushveld Complex, covers 200 square miles and has a probable thickness of 2,000 to 3,000 feet. It has been interpreted as a laccolith (S. J. S., 1).

Another very large body of nepheline rocks is that in the south of Greenland, which was described by N. V. Ussing. The Ilimausak mass contains some 50 square miles of nepheline-bearing rocks and the Igaliko mass perhaps a similar area, with an unknown extension beneath the inland ice-cap. The structure is not well known, but the complex contains the same remarkable banded rocks and the same uncommon minerals as Kola and Pilansberg.

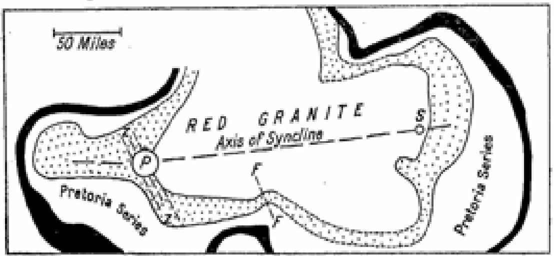

FIG. 46. Geological map of the Bushveld complex, Transvaal. The outcrop of the Dolomite formation is shown in black. The norite is stippled. P—Pilansberg. ZZ—dyke-swarm. FF—Franspoort line of syenitic and foyaitic intrusions. S—the foyaite-ijolite stock of Sekukuniland.

No other body of nepheline rocks is known which comes near the dimensions of those named. The Ditró complex in Rumania has an area of possibly 70 square miles, but little more than half of this is nepheline-bearing, the remainder including syenite and quartz-syenite (A. Streckeisen). The Serra de Monchique stock, in Portugal, covers about 24 square miles; Fourche Mountain, Arkansas, about fifteen; and the nepheline rocks of the Christiania (Oslo) district, Norway, which bulk so largely in petrographic literature, do not exceed ten square miles in all.

It is likely, then, that there are no reservoirs of undersaturated sodic magma in the lithosphere, and that when nepheline-bearing rocks appear they are due to special physical or chemical conditions which have affected granitic or basaltic magma in its ascent through the earth-crust. Observations that seem specially significant, in relation to the genesis of nepheline-bearing rocks, are:—

(1) their relatively high content of such uncommon elements as fluorine, chlorine, phosphorus, zirconium, titanium, and the cerium-yttrium group of metals;
(2) their very frequent occurrence in association with thick limestone formations.

The first point was emphasized especially by C. H. Smyth, who pointed out that the elements with which these rocks tend to be enriched are the very ones which are characteristic of pegmatite veins. Since the segregation of pegmatite is generally believed to have been accomplished with the aid of water and other volatile compounds of low molecular weight (the so-called "mineralizing agents"), Smyth claimed that the same agents, especially water, must play an essential part in the genesis of nepheline-bearing rocks. This contention may be freely granted, but it fails to explain the deficiency of silica which caused nepheline to appear in place of the usual albite and quartz of pegmatite veins.

R. A. Daly (1) lays stress on the second point and maintains that most feldspathoidal rocks have been produced from basaltic or granitic magma by reaction with limestone. He supposes that some of the silica which would normally have entered feldspar molecules combined instead with lime, and that the heavy calcium and magnesium silicates so formed sank through the magma, leaving the higher levels of the magma impoverished as regards silica. Daly also ascribes a part to the escaping carbon dioxide, and to magmatic gases, in carrying alkalis toward the top of the reservoir. The magma is supposed to become differentiated by this double agency into a lower pyroxenic portion and an upper, highly alkalic portion from which, by further differentiation, all possible types of alkali-rich rocks may be derived.

Evidence favouring Daly's hypothesis finds an impressive amount of support in the field. Some of the evidence has already been cited in Chapter V, but we must now look further into the matter. Daly compiled a list of 234 districts in the world which were known (in 1910) to contain alkaline eruptives. Of these districts, 163 show the eruptive to be closely associated with carbonate rocks, 63 are indecisive, and in 8 districts only do the geological conditions appear to be unfavourable to his theory. That is, out of 171 occurrences where the geological environment is known, 95 per cent are in regions where limestone is present.

It must be granted that mere association means little unless there is direct evidence of reaction. At the following localities there is direct field evidence of reaction between magma and carbonate rock, leading to the production of melanocratic rocks containing large amounts of diopside and other lime-rich minerals such as melanite, sphene, apatite, calcite, pectolite, melilite, perovskite, scapolite; with or without nepheline and cancrinite.

Bancroft, Ontario.
Mount Royal, Quebec.
Oka Hills, Quebec.
Ice River, B.C.
Magnet Cove, Arkansas.
Iron Hill, Colorado.
Jacupiranga, Brazil.
Loch Borolan, Scotland.
Fen district, Norway.
Alnö, Sweden.
Kuolajärvi, Finland.
Sviatoy Noss, Transbaikalia.
Turja, Kola Peninsula.
Kaindy River, Turkestan.
Erie-su River, Turkestan.
Tagobi-Sobak River, Turkestan.
Patin Mtn., W. Siberia.
Botogol Hill, Siberia.
Granitberg, S.W. Africa.
Sekukuniland, Transvaal.
Palabora, Transvaal.
Karamoja, Uganda.
Bukusu Hill, Uganda.

We shall discuss some of the more striking examples from the list, choosing first those in which the sedimentary origin of the limestone is clearly demonstrable.

In the Haliburton-Bancroft district of Ontario, which was described and mapped by Adams and Barlow, the immensely thick limestone of the Grenville series has been invaded by batholiths of gneissose granite. In the southern portion of the

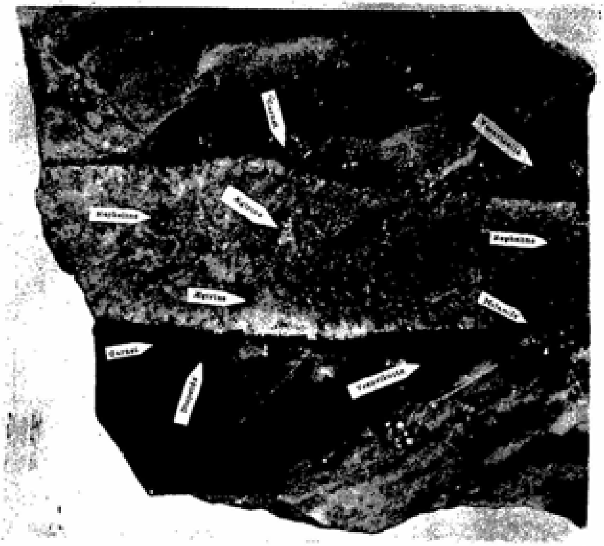

PLATE VIII.—A dike of foyaite cutting limestone and forming a selvedge of calcium silicates along each contact. The abstraction of silica led inevitably to the formation of nepheline in the dike-rock. Mount Royal, Quebec. About three-fifths of actual size. Specimen given to Columbia University by Dr. J. A. Bancroft.

[*To face p.* 316.

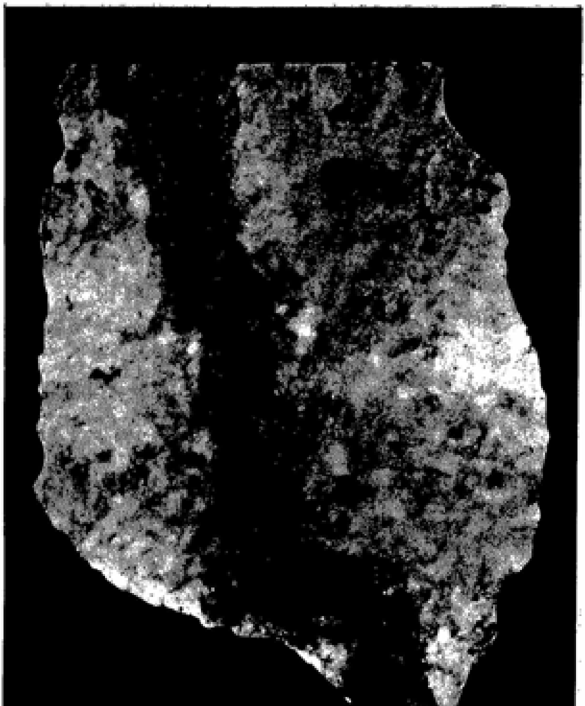

PLATE IX.—A vein of syenite cutting crystalline limestone and developing selvedges of shonkinitic composition. The syenitic portion consists of microcline (grey) and a little quartz; the selvedges are enriched with diopside (black), sphene, and apatite. Palabora, Transvaal (see p. 81). About three-fifths of actual size. [*To face p.* 317.

district the limestone is a fine-grained blue-gray rock of normal character; it is interbedded with sands, clays and tuffs, so its sedimentary origin is beyond question. Followed northward toward the granite batholiths, the gray limestone becomes coarser in grain and is associated with white crystalline marble which often contains grains of pyroxene, phlogopite and apatite. The gray and the white rocks are closely intermingled, and cores of the gray rock are sometimes preserved within the white. Some of the marble shows structures due to compression, and occasionally one finds veins and tongues of crystalline limestone which have been injected in a plastic condition into surrounding rocks (F. Chayes).

The nepheline rocks of this district invariably occur either at the junction between granite and limestone or else in the limestone near such a junction. Adams and Barlow mapped more than twenty separate intrusions of nepheline rock, from bodies of several square miles to quite small sills, and their relation to the granite margin is so constant that one can hardly suppose it to be accidental. The foyaitic magma has often given plasticity and mobility to the adjoining limestone, with the result that foyaite and " mobilized " limestone moved along together and now form composite intrusions as strongly banded as any gneiss. About the foyaite masses the limestone is sometimes charged with silicates or even converted entirely into pyroxenite or amphibolite. Near Tory Hill, W. G. Foye studied a body of feldspar-nepheline pegmatite which passes at its footwall contact into garnet-pyroxenite; the same contact rock is formed by some of the pegmatites along the York River.

Adams and Barlow regarded the nepheline rocks as a "peripheral differentiation facies" of the granite, and indicated that a gradual transition could be traced from granite through syenite into foyaite. Such a transition would have a great theoretical interest, but the writer finds himself unable to confirm it in detail. Both at Bancroft and at Tory Hill the foyaite seems to be younger than the granite, and there is no constant syenitic facies between them. Instead of a simple transition, there is frequently an oscillation of quartz-bearing, quartz-free, and nepheline-bearing facies without sharp boundaries between them, as if due to differential movement in the semi-solid magma. A narrow strip of limestone often separates the foyaite from the granite.

Another occurrence of nepheline-bearing rocks that seems highly favourable to Daly's hypothesis is the laccolithic intrusion at Loch Borolan in the north-west of Scotland. This is one of the few localities where there seems to be a complete gradation from a quartz-bearing to a nepheline-bearing rock. The hill called Cnoc-na-Sroine is formed, in its upper part, of a fairly coarse granitic rock with abundant quartz, which different writers have called quartz-syenite or nordmarkite. This passes down by simple diminution of quartz content into syenite, and after the last trace of quartz has disappeared the syenitic rock develops a little sodalite and scattered grains of melanite. At the foot of the hill the rock holds much melanite

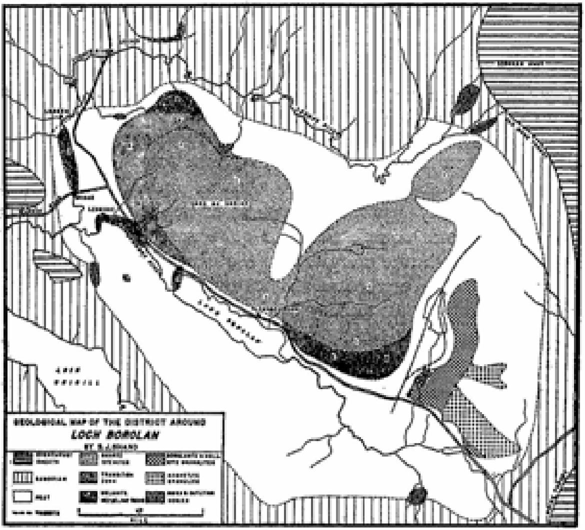

Fig. 47.

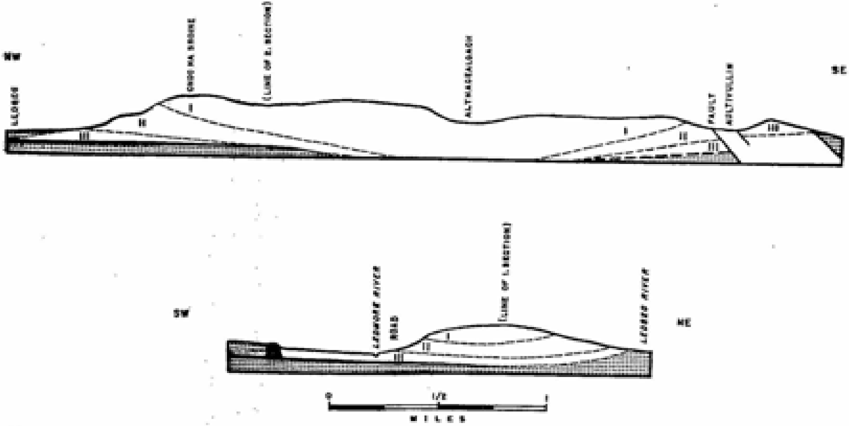

FIG. 48. Two sections through the laccolith of Loch Borolan, Scotland. *I*, subacid granite, average content of lime 1-1·5 per cent; *II*, syenite, average content of lime 2-3 per cent; *III*, diopside, melanite, and nepheline rocks, average content of lime 8-14 per cent; *IV* (black), melanite pyroxenite, content of lime 22 per cent; *V* (dotted), Cambrian limestone.

and sodic diopside, as well as micaceous pseudomorphs having the form of nepheline crystals. The hill is almost completely surrounded by white crystalline limestone, and dikes of melanite-nepheline syenite (so-called borolanite) cut the limestone at several points. A small body of melanite-pyroxenite, exposed at one point only, has the extremely high content of 23 per cent of CaO. Younger dike-rocks include ægirine-microgranite and foyaite with fresh nepheline and sodalite. Descending from the top of the hill one observes a steady decrease in the silica content of the rocks, coupled with a rapid increase of lime and magnesia near the margin or floor of the intrusion (S. J. S., 2). The evidence connecting the appearance of nepheline with the assimilation of limestone seems quite complete.

The intrusion described by J. A. Allan at Ice River, British Columbia, has a structure like that of the Borolan laccolith and exhibits rocks of similar character, although they are generally richer in nepheline. No oversaturated phase is present, but the rocks vary from leucrocratic sodalite-syenite at the top to ijolite, theralite, borolanite, jacupirangite, and other melanocratic types about the base. The intrusion is laccolithic in form; it has a limestone floor and the magma had to rise through 10,000 feet of Cambrian limestones to reach its present position. Rocks rich in lime-iron garnet are also developed at the granite-limestone contact at Sviatoy Noss, Transbaikalia. P. Eskola remarks upon the close resemblance between the pyroxene-syenite and garnet-pyroxene syenite of this locality and those of the Borolan laccolith.

At all localities in Ontario, as well as at Ice River and Loch Borolan, the sedimentary origin of the limestone is unquestionable. P. Eskola recognizes the sedimentary origin of the graphitic limestone at Sviatoy Noss, remarking that after granite, graphitic limestone is the most widely distributed rock in the Transbaikalian highlands. We pass now to other localities in which the source of the limestone is less obvious.

In the district of Sekukuniland, eastern Transvaal, a stock of nepheline rocks having an area of about six square miles cuts the red granite of the Bushveld complex (*Figs.* 49, 50). In the heart of the stock there is a great block of crystalline limestone fully half a mile square. The characteristic rock composing the stock is a schistose foyaite containing alkali-feldspar, nepheline and ægirine; but immediately round the limestone this rock gives way to a coarse-grained ijolite composed of nepheline (without any feldspar), diopside, melanite, sphene, and apatite, with cancrinite and calcite. A comparison of chemical analyses of the two extreme types shows that in consequence of assimilation of limestone the foyaitic magma lost fourteen per cent of silica and three per cent of ferric oxide, gaining instead ten per cent of lime, two of magnesia, three of phosphoric oxide, and one of carbonic acid; the proportion of nepheline rising in consequence from 27 to 52 per cent (S. J. S., 8). The limestone has been recrystallized and contains magnetite, apatite and the fibrous blue soda-amphibole, crocidolite.

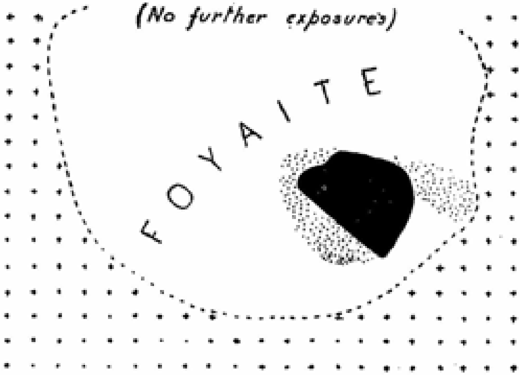

FIG. 49. Plan of foyaite stock cutting granite; Sekukuniland, South Africa. The black area is an enclosed mass of limestone, measuring 1 mile x ½ mile, about which the foyaite has been changed into an ijolite rich in lime (dotted area).

The source of this great limestone inclusion is not obvious until one realizes that the Bushveld granite is part of a lopolith, not a batholith, and that beneath its floor, perhaps three miles underground, there lies the thick dolomite series of the Transvaal System. This great sheet of more or less magnesian limestone comes to the surface some fifty-five miles east of the Sekukuniland stock, where its thickness is stated by A. L. Hall to be 970 feet. The magma that formed the stock *must* have intersected the Dolomite in its ascent, so it is unnecessary to look any further for the source of the great inclusion. Interesting confirmation is provided by the presence, in the limestone inclusion, of fibres and tufts of blue crocidolite. This mineral has *not* been recorded as a contact product anywhere in the world; but it occurs persistently in South Africa at a fixed horizon about the base of the Pretoria Series and the top of the Dolomite. The evidence is fairly strong, then, that the mass of thoroughly crystalline limestone enclosed in the Sekukuniland stock is a recrystallized portion of the Dolomite formation.

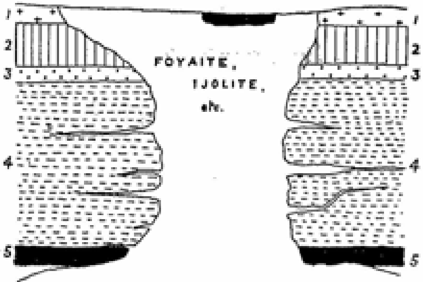

FIG. 50. A hypothetical section through the Sekukuniland foyaite stock, showing the source of the great limestone enclosure. 1, granite; 2, norite; 3, quartzite; 4, Pretoria series; 5 (black), limestone. Vertical and horizontal scales approximately the same.

The complex of nepheline rocks at Magnet Cove, Arkansas, resembles the Sekukuniland stock very closely. It consists of foyaitic rocks which are succeeded towards the interior of the Cove by ijolites and jacupirangites—rocks rich in diopside and melanite, and having a very high content of lime—and in the heart of the Cove there is the mass of crystalline limestone that one has learned to expect in such an environment. The complex, which was first described by J. F.

Williams and afterwards by H. S. Washington, is intrusive in Carboniferous beds, beneath which there is a great thickness of older Paleozoic formations containing abundant limestone.

The Kaiserstuhl, in Baden, is just another Magnet Cove except that it has been less highly eroded and consequently the rocks are mainly lavas and breccias. Lumps of more or less altered limestone, containing melanite and apatite, are to be found, and in the centre of the complex there is a great mass of crystalline limestone measuring nearly a square mile (F. Graeff). Mesozoic limestones are present underground.

The geological relations of the little stock of nepheline rocks in the Fen district, Norway, are almost identical with those of the Sekukuniland stock; but the Fen intrusion is much smaller and the enclosed block of limestone relatively larger, occupying nearly half the area of the stock. The eruptive rocks range in composition from foyaite to ijolite and pyroxenite, but the melanocratic types predominate greatly and many of them contain a great deal of calcite which forms crystals among the silicates. The name "carbonatite" is applied to intrusive rocks in which calcite is more abundant than silicates. W. C. Brögger, who has described this complex in great detail, expressed the opinion that the carbonatite was formed from "melted or assimilated limestone," and indicated as a possible source the limestone of the Telemark formation, the nearest outcrop of which is now some 25 miles away.

N. L. Bowen (1) has formed a radically different opinion about the origin of the carbonatites in the Fen intrusive. He claims that they have been formed by hydrothermal replacement of silicates by calcite, and that every stage of the replacement can be traced under the microscope. This would seem to annul the argument for a sedimentary origin of the limestone—but does it? Where did the hydrothermal solutions get the enormous quantity of calcium carbonate needed to effect this replacement? Not from the eruptive rocks, for Bowen shows that the alkali-feldspars were the first minerals to be replaced; and lime-rich minerals such as diopside, hornblende, melanite and apatite survive unchanged in most of the rocks. To find a source of lime-rich solutions, in a country of acid granitic gneiss, is no easier than to find a source of massive limestone.

Another area which is remarkable for the intimate association of nepheline rocks with carbonates is the island of Alnö,

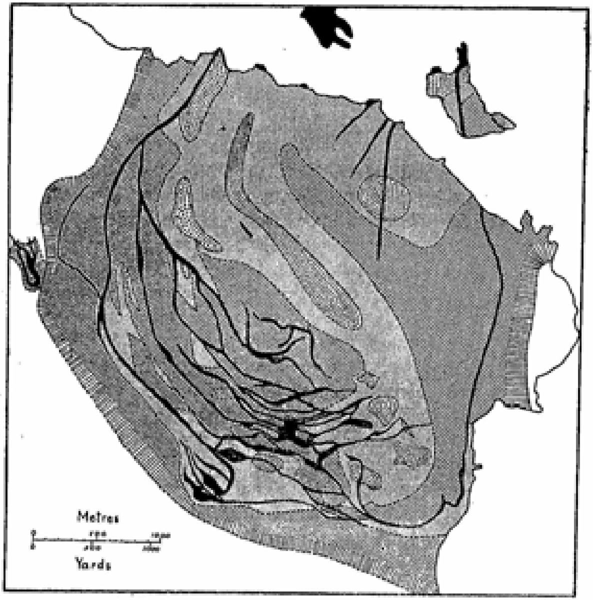

FIG. 81. Simplified geological plan of the Alnö complex, Sweden, after map by H. von Eckermann. The country rock is mica-schist and mica-gneiss. Within the complex these rocks have been enriched with alkali and converted into "fenite." The outermost zones (indicated by thin lines on the plan) are respectively quartz-fenite and syenitic fenite. Small dots indicate leucocratic nepheline-fenite; heavy dots indicate melanic nepheline rocks (foyaite, ijolite, etc.). Solid black represents carbonate rocks forming dikes and cone-sheets.

Sweden, first described by A. G. Högbom. In this case we have again to do with a stock of less than ten square miles, consisting of rocks rich in nepheline, intruded into a region of almost limeless granite gneiss. There is no single great enclosure of limestone, as in all the above instances, but the new map prepared by H. von Eckermann shows bodies of calcite- and silicate-calcite rock forming a series of cone-sheets which converge towards a common apex some 1,000 or 1,500 metres below the present land surface. Von Eckermann has demonstrated that the larger part of the alkaline mass is not of direct magmatic origin, but has been generated in consequence of the permeation of the gneiss by alkaline solutions derived from "a primary carbonatitic liquid, rich in potash, magnesia, lime, carbon dioxide, and fluorine." No suggestion is offered as to the source of this liquid, so the main problem remains unsolved.

Finally we have to mention the vents filled with brecciated rock and crystalline limestone which have lately been described by F. Dixey, W. C. Smith and C. B. Bisset in the Chilwa series of Nyassaland. The largest vent is four miles across, and several of the vents have emitted phonolitic lavas or else they contain small plugs of foyaite, ijolite, nephelinite in addition to carbonate rock. These limestones are undoubtedly intrusive, and since there are no limestones of similar quality or comparable mass in the Archæan of the region, the authors claim that the carbonates are of magmatic origin and comparable to the carbonatites of Fen and Alnö.

Other volcanic plugs filled with carbonatite, and associated with ijolite, foyaite, nephelinite and phonolite have lately been reported by W. Pulfrey in Kavirondo, Kenya; by K. A. Davies and B. C. King in Uganda; and by F. P. Mennell in Southern Rhodesia.

ORIGIN OF CARBONATITE

S. I. Tomkeieff has discussed the role of carbon dioxide in igneous magmas, with special reference to the Fen area. He contends that the greater part of the CO_2 normally present in eruptive magma escapes during consolidation, but under suitable conditions it may be so concentrated during the last magmatic stages that "not only did the carbonates crystallize out, but they formed an independent carbonatite magma-fraction."

Nobody can doubt the facts so clearly set forth by von Eckermann and Dixey. Some limestone is intrusive; that has long been known. Plastic deformation of limestone is no new thing to Alpine geologists, who have described it again and again. The conversion of normal sedimentary limestone into intrusive carbonatite is clearly demonstrated in Ontario. It is certainly astonishing to find it intruded on such a large scale as in Nyassaland, Kenya and Uganda, where it seems to have come up like great salt domes; but if one admits the existence of small intrusions of mobilized limestone it is not difficult to accept the evidence of larger ones. To admit that limestone may be mobilized and intruded in a plastic condition is a very different thing from believing in a carbonate magma, or a silicate-carbonate magma so charged with carbon dioxide that it could give rise, in crystallizing to the hollaite and the kasenite of the Fen area. Can a silicate magma dissolve a quarter of its own weight of carbon dioxide? We have no reason to suppose that it can. If it could, what sort of pressure would be necessary to keep the gas in solution? Something tremendous, we may be quite sure. Yet the carbonatites of East Africa and the limestone of the Kaiserstuhl, which are associated with lavas and breccias, must have been pushed up nearly or quite to the surface. And even if it were possible to concede the magma and the pressure, have we any ground to suppose that, in crystallizing, the silicate-carbonate magma would differentiate into separate bodies of pure silicate rock and pure carbonate rock as Tomkeieff suggests? Neither geological observation nor chemical theory justifies such an assumption.

The writer finds it easier to accept the evidence presented to his senses at Bancroft, Loch Borolan and Sekukuniland, and to regard all so-called "carbonatites" as mobilized or hydrothermally redistributed sedimentary limestones. To the argument that no limestone is known in the basement rocks of Fen and Alnö, and only a few thin dolomites in Nyassaland, he can only reply that between imagining a deep-seated mass of limestone in the Archæan, which would be no such strange thing, and imagining a carbonate magma of unknown origin, which would be a very strange thing indeed, he prefers the solution that puts the smaller strain on the imagination.

THE FOYAITE-GABBRO ASSOCIATION

In Daly's original account of the desilication hypothesis, basaltic rather than granitic magma was supposed to be the parent of the feldspathoidal rocks. Since the silica content of basaltic magma is much lower than that of granitic magmas, desilication should be more easily effected in this case. Now olivine-basalt does indeed give rise on occasion to nepheline-basalt and even to phonolite, but it is seldom that a direct connection can be established between gabbro and foyaite. Where these rocks are associated, as in the Bushveld Complex and in Essex County, Massachusetts, the foyaite is much younger than the gabbro, and there is no connecting link between them. Yet G. M. Saranchina has described an occurrence of nepheline-garnet-melilite rocks at a gabbro-limestone contact on Patin Mountain, in the lower Jenissei region of Western Siberia. The gabbro holds xenoliths of limestone, and B. M. Kupletsky (2) says "without doubt the rocks were formed during the assimilation of carbonate rocks by the basic magma." Foyaite, shonkinite, and leucite rocks are also associated with a gabbro-monzonite series which is intrusive in limestone on the Kaindy River, north-east of Tashkent, Turkestan (Kupletsky (2)).

The production on a relatively minute scale of melilite and nepheline in the contact zone of a dolerite dike which cuts chalk, at Scawt Hill, in northern Ireland, was conclusively demonstrated by C. E. Tilley and H. F. Harwood.

DALY'S HYPOTHESIS—CONCLUSION

In all the cases described above, limestone can be seen in direct contact with the intrusive rocks and there is evidence of reaction between them. The products are such characteristic minerals as diopside, sphene, apatite, melanite, scapolite, wollastonite, pectolite, melilite, perovskite; and such characteristic rocks as diopside-pyroxenite, jacupirangite, ijolite, malignite, borolanite—that is, mesotype and melanocratic rocks containing nepheline and some of the above minerals. *So far as these rocks are concerned, Daly's hypothesis is proved.* But what can we say of the leucocratic foyaites which are so much more abundant than the rocks named above, and which are characteristically poor in lime? In the entire Pilansberg complex, covering 200 square miles, the rocks are all ægirine-foyaites and tinguaites and none of them holds even five per cent of lime.

It is certainly difficult to understand why the greatest bodies of foyaite should present the least evidence of enrichment with lime, while smaller intrusions such as Magnet Cove, Iron Hill, Fen, Alnö, present it abundantly. Can the explanation be that these complexes present *different erosion-levels* to our observation? In the Transvaal, Pilansberg is a laccolite and still retains some of its roof of volcanic rocks; the Sekukuniland stock has no trace of a roof; and Palabora is in the deeply dissected Low Veld which has been cut far down into the Archæan basement. One might compare Pilansberg to the head of a mushroom, Sekukuniland to the stalk, and Palabora to the root. Similarly Lujaur-Urt is a laccolite and retains some of its cover; but Iivaara, Kuolajärvi, Alnö and Fen are stocks in the old gneiss and have been planed down nearly to their roots.

One does not wish to push a hypothesis too far, but it seems to the writer that Daly's hypothesis of the genesis of nepheline rocks has been proved in some of the cases that we have discussed, that it is probable in others, and that it is not impossible in any of the remaining cases.

The greatest difficulty which this hypothesis has to overcome is not that of proving desilication but rather that of finding a source of heat to neutralize the cooling effect of the endothermic reaction with limestone. In this connection it is essential to realize that the amount of foyaite present, in any complex for which approximate data are available, is a very small fraction of the amount of granite or gabbro involved. In the Bushveld complex the estimated volume of foyaite is less than one per cent of the volume of granite. The proportion is roughly the same in the Haliburton-Bancroft district; in Essex County, Massachusetts; and in the famous Christiania (Oslo) district of Norway. It is therefore permissible to suppose that the temperature of the small portion of magma which actually enters into reaction with limestone may be maintained by outflow of heat from the main body of magma which adjoins or underlies it.

OTHER HYPOTHESES

Other hypotheses of the genesis of nepheline-bearing rocks have been advanced from time to time, but none of them has a backing of field evidence even remotely comparable with that which lends probability to Daly's hypothesis. To explain the genesis of the Kola rocks, B. M. Kupletsky has supposed that a strong concentration of juvenile gases in a granitic magma causes the feldspar molecules to dissociate and yield feldspathoids. If this is so, where is the evidence of it? Where is the vast amount of silica that must have been set free by the reaction, and why do we never find quartz and nepheline together? In Goranson's study of the system albite-water, no nepheline was formed.

N. L. Bowen (2) has proposed an ingenious theory based upon the incongruent melting of orthoclase and the actual occurrence of "pseudo-leucite" in certain rocks. From a magma containing the components of orthoclase, with "not more than a moderate excess of free silica," leucite may crystallize at an early stage. This has actually happened in the "wyomingite" of the Leucite Hills and the "wolgidite" and other leucite rocks of Western Australia. With slow cooling, all this leucite would normally be resorbed at a lower temperature; but if there was in the early stage some relative movement of crystals and liquid, so that one portion of the liquid was impoverished with leucite and another portion enriched with it, then these two bodies of magma must behave differently in cooling. The first portion, in consequence of the removal of leucite, is relatively enriched with silica; no more leucite will be formed from it, and any crystals that remain will react with the liquid to form orthoclase. The second portion, having been enriched with leucite, no longer has enough free silica to convert all the leucite into orthoclase; instead, the crystals are replaced at the reaction temperature by an aggregate of orthoclase and nepheline, the so-called pseudo-leucite.

Bowen conceives that under favourable circumstances these two minerals might crystallize about independent nuclei instead of forming the customary pseudomorphs; in that case the final product would be an orthoclase-nepheline rock retaining no indication of its leucitic parentage. The possibility may be admitted that some rather limited bodies of nepheline rocks have originated in this way; but the special factors involved in such a genesis must place severe restrictions upon the general applicability of the hypothesis. In the first place the magma must have a special composition, neither too sodic nor too siliceous; crystallization must begin at a temperature above the leucite-orthoclase boundary surface; extensive differential movement of crystals and liquid must take place before the reaction temperature is reached; and finally the products of reaction, orthoclase and nepheline, must crystallize separately instead of forming the usual pseudoleucite aggregates. It is unlikely that these conditions will often be satisfied.

Bowen applied this hypothesis to the explanation of the Loch Borolan laccolith. Assuming that the density of the syenitic magma could not be very different from that of rhyolitic obsidian (2·37), he judged that leucite crystals, which have a density of 2·46 at ordinary temperature, would be slightly denser than the magma and would therefore sink and accumulate about the base of the laccolith. There are actually in this position certain rocks which contain little rounded aggregates of orthoclase and zeolites, long considered to be pseudo-leucite and so described by Horne and Teall. To this extent the facts seem to fit the hypothesis, but there are difficulties which must not be ignored. The most serious is that all the rocks which contain the supposed pseudo-leucite are also rich in melanite and have a density of about 3. It is impossible that leucite crystals could have sunk into a layer of magma of this density. Even if we make the assumption that the sinking of leucite took place before the magma acquired the components of melanite from its limestone walls, it is still impossible to believe that the leucite crystals could have remained suspended in a magma which was steadily increasing its density by assimilating limestone. Until these difficulties can be explained away, it will not be permissible to regard the Borolan laccolith as demonstrating the truth of Bowen's hypothesis.

A more favourable testing ground for Bowen's hypothesis has been found in the Highwood Mountains of Montana, where both leucite and pseudo-leucite are present and the nature of the latter is hardly in doubt, in spite of a complication introduced by the presence of analcime. In this region, as E. S. Larsen and colleagues have shown, differential movement of leucite (or pseudoleucite) and liquid undoubtedly took place, although it was a flotation, not a sinking of the crystals. Larsen says that "In rocks with a fine-grained groundmass, the leucite is in sharply bounded crystals. . . . As the groundmass becomes coarser, the pseudoleucite becomes coarser-grained and forms less sharply bounded crystals. In rocks with still coarser groundmass the leucite merges with the groundmass. . . . In the final stages, all evidence of pseudoleucite is lost, and the resulting rock is a shonkinite." In this case most of the requirements of Bowen's hypothesis seem to be satisfied. But the parent magma of the shonkinitic rocks in the Highwood Mountains was already undersaturated with silica at the earliest stage that is accessible to our observation. The hypothesis does not explain how this undersaturated, strongly potassic magma came into existence.

As regards the other soda-feldspathoids, sodalite, haüyne, nosean and cancrinite, it seems unnecessary to discuss their genesis apart from that of nepheline, with which they are almost invariably associated. They are most simply regarded as addition products of nepheline containing chloride, sulphate, and carbonate radicles respectively.

Rocks Containing Analcime

The case of analcime is peculiar, for this mineral contains twice as much silica as nepheline does and in addition it holds one molecule of loosely combined water. Although it is best known as a secondary mineral in vapour cavities, where it is associated with zeolites and calcite, analcime is becoming more and more widely recognized as an *apparently* primary constituent of certain sodic rocks. In these it may form euhedral crystals, sometimes 1-2 centimetres in diameter, set in a fine-grained trachytic, phonolitic, or basaltic groundmass. In some syenitic and foyaitic rocks analcime is entirely allotriomorphic, filling the sharply angular interstices between feldspar laths; in this case the feldspar may either be unaffected or it may be deeply corroded and replaced by analcime. Again, in some melanic dike-rocks and lavas, analcime forms rounded bodies (*ocelli*) in which it is occasionally associated with idiomorphic nepheline crystals. In the last two cases it would probably be correct to regard the analcime as a deuteric rather than a magmatic product.

Good trapezohedral crystals of analcime were discovered by C. W. Knight in a trachyte tuff in Alberta, Canada; Knight was led to predict the discovery of analcime-trachyte, to which he gave in advance the name blairmorite. Blocks of such a rock were collected later by J. D. MacKenzie, but no body of the rock seems to have been found in place. The rock consists of large flesh-red crystals of analcime in a fine-grained groundmass of analcime, sanidine, nepheline, pyroxene and melanite. Lavas of the same character as blairmorite have since been recognized in Rhodesia (Dixey and Smith), and a remarkable analcime-tinguaite was described by J. A. Bartrum in New Zealand. H. S. Washington demonstrated in 1914 that the supposed leucite crystals in certain basalts of Sardinia are actually analcime, thereby throwing suspicion upon other occurrences of supposed leucite in highly sodic rocks. Although analcime normally contains no more than a trace of potash, Larsen and Buie have recently described an occurrence of potassic analcime in rocks of the Highwood Mts., Montana.

The question arises whether a silicate containing 8 per cent of water, which it gives off about 500° C., could possibly be the first mineral to crystallize from a trachytic or basaltic magma, even if one admits that that magma was a very hydrous and strongly sodic one. The unlikelihood of such a thing has led many authorities to regard these analcime insets as pseudomorphs after leucite crystals. It is known that leucite can be transformed into analcime, and analcime into leucite, by the action of solutions of soda and potash salts respectively. Nevertheless, other authorities have considered the insets to be primary analcime. The truth is that there is no known way of deciding between the two possibilities. F. A. Bannister has compared the X-ray spectrum of one of these questionable analcime crystals with that of leucite, but finds it impossible to decide whether the analcime is primary or pseudomorphous. J. W. Gruner showed that the atomic structure of analcime is not destroyed by heating to 700° ; and J. B. Lyons holds that " the field facts do not demand that the analcime did not crystallize at magmatic temperatures."

In the second type of occurrence, analcime is confined to the angular spaces between the feldspars, which may be either anorthoclase or plagioclase. The mineral does not simply line these cavities with separate crystals, as happens when a vapour cavity is filled with zeolites, but each interspace is completely filled with analcime which shows the same cleavage directions throughout. Can it have been formed at the expense of some soda-rich mineral that formerly occupied the cavity? J. F. Williams has described the alteration of interstitial nepheline into analcime in Arkansas, but this explanation will not serve in every case. The writer has described a foyaite at Franspoort, Transvaal, which contains idiomorphic crystals of nepheline as well as interstitial analcime. The former has crystallized before feldspar and ægirine; the latter crystallized after feldspar and it surrounds idiomorphic needles of ægirine (S. J. S., 4).

One may suppose that when the temperature of the magma was comparatively high and the concentration of water had not reached its maximum, the available silica was distributed between nepheline and albite and crystallization began with these phases. As the temperature fell and the concentration of water increased, the stability region of analcime was reached and the last of the magma crystallized in that form.

But if this view is correct, it must compel us to reconsider the case of blairmorite and other rocks with porphyritic analcime. In some of these rocks there is nepheline in the groundmass which surrounds the analcime insets. A return from the hydrous condition of the magma that favours the production of analcime to the less hydrous condition that permits nepheline to be formed is surely inconceivable, since it would involve the destruction of the analcime crystals. Thus one seems compelled to accept the older view, that the insets of analcime in these rocks are pseudomorphous after leucite.

As regards the classification of these puzzling analcime rocks, we may agree with A. Lacroix, who says, "It is not necessary to make a special group parallel to that of the nepheline rocks; it is preferable to consider them as a special case of the latter, from which they differ only in this, that richness in magmatic water has permitted the crystallization, at low temperature, of analcime in place of nepheline."

EXTRUSIVE ROCKS

After this long discussion of the intrusive nepheline rocks, the *extrusive* rocks may be treated very briefly. Some phonolite is certainly connected with foyaite underground, as in Pilansberg and Leeuwkraal, South Africa. But small bodies of phonolite are sometimes developed, just as small bodies of trachyte are developed, in connection with olivine-basalt. The phonolitic trachytes of St. Helena and other oceanic islands sometimes carry small amounts of nepheline or sodalite. It is unlikely that assimilation of limestone played any part in these cases. A more likely solution may be found in the equation

$$\text{plagioclase} + \text{olivine} \Longleftrightarrow \text{nepheline} + \text{pyroxene}.$$

That this is a real reaction is indicated by the occurrence of plagioclase and olivine in some rocks, of nepheline and clino-pyroxene in others, and of all four minerals in a third group. Thus from an olivine-basalt magma a nepheline-bearing magma may be developed under certain conditions; and the separation of residual liquid from such a magma during its crystallization might yield a small body of phonolite.

The nepheline-plagioclase lavas may also originate in more than one way. The reaction that we have just discussed may produce a nepheline-tephrite or basanite in place of common olivine-basalt; but the quantity of nepheline that could be formed by this reaction is limited by the rather small content of olivine in the average olivine-basalt. The nepheline content might be increased by gravity concentration; but in the genesis of nepheline-rich tephrites and basanites it is likely that Daly's reaction has played its part, as it certainly has in the case of the leucite-tephrites of Mount Vesuvius.

The formation of nepheline-dolerite from an olivine-basalt magma, under conditions that were almost those of extrusion, has been fully and convincingly described by Tilley and Harwood. At Scawt Hill, in northern Ireland, a small plug of olivine-dolerite has broken through chalk, and between the two there is a zone of intense reaction. The basaltic magma has assimilated so much chalk that it gave rise to pyroxene-rich dolerite and pyroxenite, as well as a small amount of nepheline-dolerite. In this reaction, nepheline was formed at the expense of plagioclase, and at the same time the common augite of the dolerite, which contains a small excess of $MgSiO_3$ over $CaSiO_3$, was replaced by a purple titanaugite in which $CaSiO_3$ is in excess. With still further assimilation of lime, melilite was formed in addition. The total amount of nepheline-dolerite is small, because the whole intrusive body is small and chilling took place very rapidly; but the occurrence affords at least a qualitative proof of Daly's hypothesis.

THE OCCURRENCE AND GENESIS OF LEUCITE ROCKS

Rocks containing leucite have a very restricted distribution and they are almost entirely extrusive. Even when they make a big show on the surface, as in the Roman region and in central Africa, the actual volume of leucite-bearing rock is small. H. S. Washington made an approximate computation of the volume of leucitic lava emitted by the seven volcanoes in the Roman region. The distance from the most northerly of these volcanoes, Bolsena, to the most southerly, Mount Vesuvius, is about 170 miles, and the area covered by the ejectamenta is about 2,340 square miles. This is more than double the area of Umptek and Lujaur-Urt together; yet Washington's estimate of the volume of all the leucitic lava in this area is only 40 cubic kilometres or about 10 cubic miles. No doubt much of the tuff within the area represents disintegrated leucitic lava; but even if we doubled or trebled Washington's estimate the result would not be very impressive.

It is questionable whether any other region of leucite rocks equals the Roman region in size. The nearest is the volcanic field at the north end of Lake Kivu in central Africa, which covers an area of 1,450 to 1,500 square miles (Combe and Simmons). Some of the lavas were very fluid and spread far over the low land, but there is no reason to estimate the thickness of the flows at more than a few score feet, and a considerable proportion of the lava is not leucite-bearing. Probably 10 cubic miles is an ample estimate for all the leucitic lava in this field.

The Leucite Hills in Wyoming are a group of 22 little hills formed of cones and sheets of (mainly) leucite-bearing lava (Kemp and Knight). The hills are scattered through an area of 30 by 25 miles, but the lavas have not spread far from the vents, and Washington indicates that the total volume of leucite rock is no more than one-fiftieth of that in the Roman region.

In the West Kimberley district of Western Australia, a group of 20 small craters, plugs and dikes have yielded leucite rocks which resemble most closely those of the Leucite Hills. The volume of rock exposed is quite small (Wade and Prider).

Leucitic lavas occur in the East Indies, especially in Java and Celebes (Iddings and Morley). There are five volcanoes strung along a 300 mile line in north-east Java which have yielded chiefly leucitic lava. The minor occurrence on Celebes is 500 miles away, so it is hardly justifiable to consider it as part of the Java field. It is not possible to form any estimate of the volume of leucite-bearing lava in this region.

The absence of plutonic rocks containing leucite is remarkable. The "missourite" of the Highwood Mountains, Montana, which is generally named as the plutonic equivalent of leucite-basalt, is not really a plutonic rock at all; it occurs in very limited amount in a stock or volcanic plug among rocks of the most variable texture and composition. Among the ejected blocks on Monte Somma there are some that have been described as leucite-syenite, leucite-foyaite, and leucite-bearing pyroxenite, but these rocks are not found in place. Blocks of leucite-ijolite have been found on Niligongo, one of the volcanoes of the Lake Kivu field. Granular rocks containing pseudo-leucite together with nepheline, pyroxene, melanite, have been described on the Kaindy River in Turkestan (V. Nikolajew). The borolanite of N.W. Scotland is a doubtful member of the pseudo-leucite class.

As regards the genesis of this very rare group of rocks, it is to be observed that leucite is commonly associated with some proportion of nepheline, and that melilite and melanite are rather frequent accessory constituents of leucite rocks. This suggests that assimilation of limestone may have played a part in the genesis of these rocks. The question has been investigated, in the case of Mount Vesuvius, by A. Rittmann. This best known of all volcanoes has drilled its vent through *two miles* of Cretaceous limestone and Triassic dolomite. Rittmann shows that the degree of silication, or the ratio of the actual molecular proportion of silica to that necessary for complete saturation of all the silicate minerals, has declined steadily in the course of time from 0·89 in the oldest prehistoric pumice to 0·67 in the recent lavas. In terms of minerals this means that there has been a progressive reduction in the feldspar content of the lavas and a corresponding increase in the content of leucite and nepheline, combined with increasing colour-index. Rittmann concludes that the development of the Vesuvian magma from the ancient Somma magma has involved assimilation of the Triassic dolomite and gravitational differentiation of the contaminated magma.

Other well-known leucite regions are also underlain by limestone; this is true of the Leucite Hills; the Eifel; Bohemia; the West Kimberley district; Kaindy River, Turkestan; the Trans-Caucasus; and the Java and Celebes fields. In Java, H. A. Brouwer has found a remarkable demonstration, although on a minute scale, of the formation of a leucitic glass from an andesitic magma by reaction with an enclosure of limestone.

The Lake Kivu region is a critical one for the hypothesis that limestone plays a part in the generation of leucitic lavas. A. D. Combe and W. C. Simmons, of the Geological Survey of Uganda, mapped the Bufumbira area, the eastern portion of the Kivu field, and they do not record any limestone there or in western Uganda in general; consequently A. Holmes and H. F. Harwood (1), who have furnished an admirably detailed account of the petrology of the lavas, dismiss Daly's hypothesis from consideration on the ground that "there is little or no limestone or dolomite to be assimilated."

It cannot be too often repeated that the absence of outcrops of limestone is no proof of its absence in depth, especially among Archæan rocks.[1] The following points suggest that the proportion of limestone in the basement of western Uganda may be greater than the outcrops suggest.

Right on the northern margin of the Kivu volcanic field, near Rutschuru, Lacroix and Delhaye have described a considerable body of foyaite. In the composition of this rock both cancrinite and melanite play a part, and calcite is also mentioned as a constituent. To anybody who has studied the genesis of nepheline, melanite and cancrinite, this occurrence is very significant. An entirely similar occurrence is recorded by J. Thoreau in Urundi, south of Lake Kivu, with diopside-pyroxenites and scapolite-pyroxenites.

In the Toro-Ankole district, north and south of Lake George, there are many explosion craters, some with tuff cones and small bodies of lava. A number of ejected blocks from these craters, collected by the Geological Survey of Uganda, were described by Holmes and Harwood (2). Among them were fragments of various leucite, melilite and nepheline-bearing lavas, many of them highly carbonated; also melanite-pyroxenite, sphene-pyroxenite, tremolite-amphibolite; besides gneiss and granite. The association of melilite with diopside, melanite, sphene, perovskite and tremolite in these boulders should make one hesitate to assert the non-existence of limestone in the basement of western Uganda.

What is the alternative to reaction with limestone, as an explanation of the genesis of leucite-bearing and melilite-bearing lavas in western Uganda and the Lake Kivu field? Holmes offers us a series of purely hypothetical reactions involving augite-peridotite, "emanations," and feldspathized sediment. "Emanations" seem to play the same part in some modern petrology that "humours" played in medieval medicine; they may be used to explain anything. The only tangible evidence for Holmes' speculations is the occurrence

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[1] A distinguished reviewer in the *Journal of Geology* takes exception to this statement, claiming that it shows "blind faith in a theory." Surely the man who shows blind faith is the one who makes positive statements about the composition of a solid prism of rock after seeing only one of its end faces.

of occasional nodules of peridotite and pyroxenite in the Bufumbira lavas and tuffs. It is not clear that these fragments are any more numerous than those that contain melanite and perovskite, or any more valuable as evidence in the investigation. It seems to the writer that the evidence of the lime-silicate fragments, added to that of the Rutschuru foyaite and the melilite-bearing lavas, puts the case for deep-seated limestone in the basement rocks of western Uganda and the Lake Kivu region in a stronger position than any hypothesis that depends upon hypothetical emanations.

An account of the genesis of leucite rocks must explain not only the deficiency of silica, but also the excessive content of potash in relation to soda. Rittmann has considered this aspect of the problem at Mount Vesuvius and thinks the explanation is to be found in expulsion of soda from the magma into the country rocks. It is not easy to study the contact rocks beneath an active volcano, but the ejected blocks of metamorphosed limestone contain nepheline and sodalite as well as lime silicates.

There are to be found in the literature quite a number of descriptions of potash-rich rocks at limestone contacts. For instance, Dixey and Smith comment upon the high potash content of the feldspar that is associated with the carbonatites of Nyassaland. G. A. Joplin noted the appearance of orthoclase in the reaction zone between diorite and limestone at Ben Bullen, New South Wales. A simultaneous enrichment with potash and lime is recorded in rocks of shonkinitic or malignitic character at Predazzo, Loch Borolan, Sviatoy Noss, and Kaindy River (S. J. S., 5). Since limestone cannot furnish potash to a magma, it seems to follow that the enrichment with potash is relative and that it has been effected by expulsion of soda.

The granite-limestone contact at Palabora, Transvaal, was described in Chapter V, and it was shown (*Fig.* 8) that there is evidence of three distinct processes, namely desilication of granitic magma, assimilation of lime and magnesia, and elimination of soda (S. J. S., 6). From a soda-potassic granite there has thus been formed a microcline-diopside rock that is as nearly as possible free from soda. Where the soda has gone is by no means clear, nor the form in which it was expelled; but perhaps the likeliest supposition is that it escaped in the form of a very fluid foyaitic magma, to form independent intrusions at a higher level in the crust. It is not thought that the microcline-diopside rock existed as a magma, but it has acquired the very high content of potash which is a prerequisite for the formation of leucite.

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The part of wisdom is to know that we don't know, and not fool ourselves with long words.—UPTON SINCLAIR.

CHAPTER XIX

METEORITES

THE greatest weakness of petrography, as a method of investigating the composition of the earth, is that direct observations can only be made upon rocks that have crystallized within a few thousand feet from the surface, say a thousandth part of the earth's radius. There is good reason, no less on petrological than on geophysical and astronomical grounds, to suppose that the matter that forms the interior of the planet is different in many respects from surface rocks; but direct observation of this matter is naturally impossible. It is here that meteorites come to our assistance. It is quite immaterial what view one holds regarding the origin of meteorites. Astronomy, supported by the spectroscope, has shown that the universe is formed of the same elementary materials throughout; so whether one regards meteorites as shattered stars or shattered planets, nebular knots or condensed comets, as bolts from the sun or as bombs from terrestial volcanoes, the important fact for us is that *meteorites bring to our notice a type of rock existing in the solar system but different from anything that occurs in the outer shell of our planet;* and this material has just the sort of composition that petrology, geophysics and astronomy combine to indicate as likely to be found in the interior of the earth. For this reason, the study of meteorites has an extraordinary interest for the petrologist, and no discussion of eruptive rocks can be complete which ignores this fascinating aspect of the subject.

Meteorites show their relationship with the more familiar eruptive rocks in many ways. Most of the constituents of meteorites, though not all, are identical with common rock-forming minerals, and they show the same crystalline habit as these. The textures of meteorites are partly similar to those of eruptive rocks and partly different; but the differences between meteorites and terrestial lavas are not more striking than those between lavas and plutonic rocks, or lavas and tuffs. On the other hand, meteorites have scarcely any characters in common with sedimentary rocks or crystalline schists.

THE CHEMISTRY OF METEORITES.

The abundant chemical elements in eruptive rocks are oxygen, silicon, aluminium, iron, calcium, sodium, potassium, magnesium, and titanium, in that order, with hydrogen and sulphur occasionally rising to prominence. (See the table on page 31.)

The common elements in the iron meteorites are iron and nickel, with small quantities of cobalt, manganese, and copper. The elements of the stony meteorites (disregarding admixed grains of nickel-iron) occur in the following order of abundance, according to a computation by G. P. Merrill:—oxygen, silicon, magnesium, iron, aluminium, calcium, sodium, chromium, manganese, potassium.

There is a difficulty about combining these statements into a conclusion that could be applied to all meteorites. Stony meteorites fall much more numerously than iron ones, but owing to their brittle character they tend to be dissipated in the form of small fragments and dust and so to escape notice. The largest known meteoric stone reached the earth as a lump of about one ton in weight. Again, the similarity of meteoric stones to many terrestrial rocks makes their discovery largely a matter of accident. The iron meteorites, on the other hand, are comparatively big. The Cape York iron, from Greenland, is eleven feet long and weighs over 36 tons. The Hoba meteoric iron, in South-West Africa, measures ten feet by nine as it lies in the ground, and the apparent thickness is about four feet. L. J. Spencer (1) estimates the weight to be about sixty metric tons. Even a small lump of iron is more likely to attract notice than a lump of stone; thus it is certain that no comparison of irons and stones, either by number or by weight, can give a trustworthy base for computing the relative abundance of the chemical elements in the irons and stones combined. W. A. Wahl holds that the order of abundance is approximately as follows:—iron, oxygen, silicon, magnesium, aluminium, calcium, nickel, sodium, sulphur, potassium, cobalt, phosphorus, carbon, hydrogen.

By adding together all the reliable chemical analyses that have been made, without taking weights into account at all, O. C. Farrington formed the following estimate of the average composition of meteorites :—

iron	72·06	aluminium	·39
oxygen	10·10	sodium	·17
nickel	6·50	phosphorus	·14
silicon	5·20	chromium	·09
magnesium	3·80	potassium	·04
sulphur	·49	carbon	·04
calcium	·46	manganese	·03
cobalt	·44	other elements	·05

In addition to those named, a number of other elements have been detected in very small quantity; these are chlorine, copper, tin, titanium, vanadium, gold, the metals of the platinum group, and the inert gases. No element has been found in meteorites that was not already known on the earth; on the other hand, some of the comparatively common elements have not yet been detected in meteorites. Of these fluorine is the most conspicuous example; while the almost total absence of the strongly radioactive elements from meteorites is a point of some significance.

Elements which occur in the native, or uncombined, state are iron, nickel, cobalt and carbon, commonly; manganese, copper, hydrogen, nitrogen and other gases, gold and the platinum group of metals, in traces. Copper is a very scanty component of most meteoric irons, but the Richardton stone exhibits specks of native copper. The platinum metals have been found in Arispe, Cape York, Canon Diablo and other irons, and gold in Melrose. The purely stony meteorites, such as Juvinas and Stannern, contain none of these very heavy metals. The presence of native phosphorus in the Saline stone, Kansas, was announced by O. C. Farrington.

W. A. Wahl has pointed out that the elements which are found in the native state are those whose oxides have a low heat of formation, as the following table shows :—

$\mathrm{Mg} + \mathrm{O} = \mathrm{MgO} + 145{\cdot}8$ cal.
$\mathrm{Ca} + \mathrm{O} = \mathrm{CaO} + 131{\cdot}3$,,
$\mathrm{Al}_2 + \mathrm{O}_3 = \mathrm{Al}_2\mathrm{O}_3 + (3 \times 131{\cdot}2)$,,
$\mathrm{Na}_2 + \mathrm{O} = \mathrm{Na}_2\mathrm{O} + 100{\cdot}2$,,
$\mathrm{K}_2 + \mathrm{O} = \mathrm{K}_2\mathrm{O} + 97{\cdot}1$,,
$\mathrm{Si} + \mathrm{O}_2 = \mathrm{SiO}_2 + (2 \times 90{\cdot}9)$,,

$\mathrm{Mn} + \mathrm{O} = \mathrm{MnO} + 90{\cdot}8$ cal.
$\mathrm{C} + \mathrm{O} = \mathrm{CO} + 68{\cdot}2$,,
$\mathrm{Fe} + \mathrm{O} = \mathrm{FeO} + 64{\cdot}6$,,
$\mathrm{Co} + \mathrm{O} = \mathrm{CoO} + 63{\cdot}8$,,
$\mathrm{Ni} + \mathrm{O} = \mathrm{NiO} + 59{\cdot}7$,,
$\mathrm{Cu}_2 + \mathrm{O} = \mathrm{Cu}_2\mathrm{O} + 43{\cdot}8$,,

Each element in this list is able to reduce the oxides of those which follow it in the list; thus no iron can become oxidized until all the sodium is in the form of the oxide; no nickel, until all the iron has been oxidized, and so on. This leads directly to a conclusion that has been established by G. T. Prior, that in meteoric stones generally, the poorer they are in nickel-iron, the richer that iron is in nickel, and the richer in iron are the magnesium silicates.

The Mineralogy of Meteorites

The minerals of meteorites may be discussed in two groups, as follows:—

(1) Minerals which occur in terrestrial eruptive rocks;

(2) Minerals which have not been found in terrestrial rocks.

The common minerals of group (1) are olivine, ortho-pyroxenes, clino-pyroxenes (generally rather poor in lime and nearly free from alumina), plagioclase feldspars ranging from anorthite to oligoclase, chromite, and pyrrhotite (including the variety troilite). The less common minerals of this group are quartz, which has been found in the St. Marks stone as well as in several irons; tridymite, graphite, magnetite, chlorapatite, various hydrocarbons, and some soluble salts such as NaCl. Minute diamonds have been identified in Cañon Diablo iron, and a red spinel was reported by W. M. Foote in the Holbrook stones, but this has been disputed.

The chief members of group (2) are the alloys of iron with nickel, including a little cobalt, manganese and copper. When a polished surface of an iron meteorite is treated with a suitable etching reagent, a peculiar structure is often revealed which shows that the iron is built up of intersecting plates of at least two materials, differing in colour and in resistance to acids, while a third material fills the angular interstices. The broader plates, which are more readily attacked by acids, consist of an alloy called *kamacite*, having the approximate composition Fe 93 per cent, Ni(+Co) 7 per cent. Narrower plates, the edges of which stand out in relief on the etched surface, are composed of *taenite*, another alloy containing from 14 to nearly 50 per cent of nickel (and cobalt); this alloy is much less abundant than kamacite and is sometimes entirely absent. The material which fills the interspaces is called plessite; it is intermediate in composition between kamacite and taenite, and it is perhaps just a microcrystalline aggregate of these two alloys.

Alloys of iron and nickel are not entirely unknown in terrestrial rocks. Masses of iron containing one to two per cent of nickel were found in basalt on Disko Island, West Greenland, and grains of nickeliferous iron have been found in gravels derived from serpentine and peridotites in New Zealand and elsewhere; but these irons differ both in structure and in composition from the iron of known meteorites.

Other minerals which occur in meteorites, but not in eruptive rocks, are *lawrenceite* (ferrous chloride); *cohenite*, a carbide of iron and nickel corresponding to the formula $(Fe,Ni)_3C$; *schreibersite*, a phosphide of the same metals (with a little cobalt) having the formula $(Fe,Ni)_3P$; *oldhamite* (calcium sulphide); *daubréelite*, a double sulphide of iron and chromium, $FeCr_2S_4$, which is the sulphur analogue of chromite; *moissanite* or carborundum (silicon carbide, SiC); *merrillite*, a rare phosphate corresponding to the formula $3CaO,Na_2O,P_2O_5$; *osbornite*, titanium nitride, TiN; and *maskelynite*, once thought to be a distinct mineral of isometric crystallization, but now considered to be plagioclase feldspar which has been heated to a temperature at which the atomic space lattice broke down.

THE TEXTURE OF METEORITES

In respect of texture, meteorites fall into two main groups. Those of one group resemble products formed by the crystallization of a liquid; those of the other group are comparable rather to volcanic breccias and tuffs. To the former group belong most of the irons and only a few stones, the majority of meteoric stones being brecciated or tuffaceous.

The iron meteorites (*siderites*) show three varieties of texture, characterized respectively by (a) the arrangement of plates of kamacite parallel to the faces of a cube (hexahedral irons or hexahedrites); (b) the arrangement of plates of kamacite and taenite parallel to the faces of an octahedron (octahedral irons or octahedrites); (c) an apparent absence of any regular structure (ataxites). Accessory minerals such as schriebersite, cohenite, and pyrrhotite are generally present;

they form irregular nodules in the nickel-iron, and it is clear that they were the first substances to crystallize; they were followed by kamacite, taenite, and plessite, in that order.

In the *syssiderites* there is a continuous, sponge-like matrix of nickel-iron enclosing isolated grains of olivine or pyroxene. These are sometimes well-formed crystals, but more frequently rounded or broken grains; and the iron itself may show a brecciated structure. The "pallasites" of Krasnojarsk and Imilac, specimens of which are to be seen in most meteorite collections, are beautiful examples of this type.

The structure of the *aerolites* or meteoric stones naturally interests us more than that of the irons. In these stones which hold little nickel-iron, and which in that respect resemble terrestrial rocks most closely, the texture is often ophitic and quite comparable to that of a gabbro or basalt. The Juvinas stone, for example, consists of augite and anorthite in ophitic intergrowth, with a little chromite, magnetite and pyrrhotite; it exhibits drusy cavities, which are lined with well-formed crystals. The Jonzac stone is as coarse-grained as many gabbros, some of the anorthite crystals exceeding a centimetre in diameter. Such stones are generally given the name "eucrite," which was originally applied to an anorthite-gabbro. There are also granular stones such as Chassigny and Bishopville, which are composed of uniform grains of olivine or pyroxene, often held together very loosely.

It must be admitted that stones which resemble eruptive rocks as closely as these do are very rare. Among the brecciated stones, however, there are many, such as Petersburg and the "howardites," which are made up of fragments of ophitic stones. G. T. Prior says of the brecciated Simondium meteorite that it "resembles in many parts a terrestrial basalt perhaps more closely than any other meteorite." In thin sections under the microscope, this meteorite shows particles of metallic iron, in grains generally less than a millimetre in diameter, large rounded olivine crystals, and smaller irregular grains of enstatite, lath-shaped feldspars (labradorite to anorthite), and grains of magnetite. As for the Stannern stone, if it had not actually been observed to fall, it is difficult to see how it could be distinguished from a terrestrial basalt-tuff.

The great majority of stony meteorites have a characteristic texture to which no exact analogue can be found among eruptive rocks. They are made up largely of little rounded bodies (chondrules) of various kinds, embedded in a fragmental base. A common type of chondrule is a little spheroid composed of one or more well-formed, rounded, or skeleton crystals of olivine, intergrown with glass. In another type there is a fibrous structure, due to eccentrically arranged needles of enstatite, which seem to show, by this arrangement, that crystallization started from the outside of the chondrule. Other chondrules contain both olivine and enstatite, together with glass; others contain nickel-iron; some hold plagioclase; and some are just little drops of glass. Chondrules are typically spheroidal in form, but they are frequently flattened or indented, and entire chondrules and broken ones are often found together. Occasionally a large chondrule is seen to enclose a smaller one. Chondrules have been supposed to have originated through the rapid cooling of liquid silicates, thrown up by explosive action into a hot (solar? stellar?) atmosphere. This view implies a mode of formation comparable with that of a tuff.

Entire meteoric stones are surrounded by a thin crust of glass, generally of a very dark colour owing to the presence of minute grains of magnetite. The crust is believed to be formed during the passage of the stone through the atmosphere. Thin branching veins of a similar dark glass often traverse the interior of a stone; they resemble the veins of "pseudotachylyte" which are developed in certain ancient rocks in South Africa, India and Scotland. Pseudotachylyte is a compacted and partly vitrified silicate dust, generated within fissures in granite, gneiss and other hard rocks which have been subjected to violent mechanical shock. The resemblance between pseudotachylyte veins and the dark veins in meteoric stones may be held to support the view that meteorites have been produced by the sudden disintegration of some larger celestial body. Pebbles and rough lumps of a highly siliceous glass have been found in several parts of the world where there are no volcanoes and certainly no glass-factories; for instance, in the Arabian and Libyan deserts and on the Nullarbor plains of southern Australia, as well as in Tasmania, India, Indo-

China and Java. Many of them are black or dark green, just like obsidian, and they take the form of little rounded pebbles or buttons; they contain 70 to 80 per cent of silica. But the latest discoveries in the Arabian and Libyan deserts and in central Australia present a nearly colourless glass with 90 to 97½ per cent of silica, and L. J. Spencer (2) suggests that this material may have been formed "by the fusion of desert sand in the heat developed by the fall of gigantic meteorites." No such explanation is applicable to the widely distributed "australites" of southern Australia, which have been picked up over an area of some 2,000,000 square miles, especially on the Nullarbor limestone plains (C. Fenner). To account for these a shower of glass meteorites has been suggested. But until a glass meteorite is observed to fall—which has not yet happened, although the number of observed falls of meteorites is nearly six hundred—it is advisable to keep an open mind about the origin of these curious glasses.

Geographical names such as moldavite, billitonite, australite, by which these glasses have been known in the past, are now being replaced by the simple descriptive name *tektite*. In a recent study of the tektite problem, by V. E. Barnes, it is shown that many tektites contain minute particles of fused quartz (lechatelierite), just as fulgurites do. This observation, along with the very high silica content, distinguishes tektites from all known volcanic glasses. Barnes concludes that some, if not all, tektites have been formed by the melting of sandy sediments by lightning.

CLASSIFICATION AND DESCRIPTION OF METEORITES

Until quite recently, meteorites were classified primarily on the basis of texture, with little regard to composition except as concerns the proportion of nickel-iron. A quantitative chemical classification of meteorites, on the lines of the norm classification of eruptive rocks, has been introduced by O. C. Farrington. The classification of G. T. Prior is a genetic one, based on the hypothesis that meteorites have separated from a single magma which has passed through successive stages of progressive oxidation. Since the metallic nickel cannot become oxidized until all the iron is in the state of oxide, as has already been shown, it follows that the ratio of Fe : Ni in the metallic portion is related to the stage of oxidation of the magma.

Prior recognizes the usual four classes, irons, stony irons, chondritic stones and achondritic stones, and within each class he sets up four groups with the following distinctive characters:—

Group 1.—Nickel-iron; ratio Fe : Ni greater than 13.
Magnesian silicate; enstatite nearly free from iron.
Feldspar; oligoclase.

Group 2.—Nickel-iron; ratio Fe : Ni less than 13, greater than 8.
Magnesian silicates; bronzite and olivine containing more than four times as much MgO as FeO.
Feldspar; oligoclase.

Group 3.—Nickel-iron; ratio Fe : Ni less than 8, greater than 2.
Magnesian silicates; hypersthene, clino-pyroxene and olivine containing less than four times but more than twice as much MgO as FeO.
Feldspar; oligoclase.

Group 4.—Magnesian silicates; clino-pyroxene and olivine containing less than twice as much MgO as FeO.
Feldspar; anorthite.

In order to illustrate the variations in the composition of meteorites, for the purpose of comparing them with terrestrial rocks, we shall give the mineralogical composition (mode) of a number of examples, arranging them simply in the order of increasing content of nickel-iron. In view of the scarcity of quantitative descriptions of eruptive rocks, it is interesting to find that in most accounts of meteorites the mineralogical composition is stated quantitatively and very accurately.

METEORITES CONTAINING LESS THAN 10 PER CENT OF NON-SILICATES

It is among the stones containing less than 10 per cent of non-silicates that the closest resemblance to eruptive rocks is observed. The Juvinas stone, to which we have already referred, has, according to C. Rammelsberg, the following composition, which is just that of an anorthite-gabbro :—

pyroxene	62·6	chromite	1·3
anorthite	34·5	magnetite	1·2
apatite	0·6	pyrrhotite	0·2

The Petersburg stone is a brecciated eucrite, very similar to Juvinas in composition; Rammelsberg gives pyroxene 68·6, anorthite 30, pyrrhotite 0·6, in this case. In the Sherghotty eucrite, G. Tschermak found pyroxene 73·4, plagioclase (maskelynite) 22·5, magnetite 4·5. The Shalka stone is practically a pyroxenite, consisting almost entirely of a variety of hypersthene rich in ferrous oxide.

None of the above stones contains any olivine. In the Hendersonville chondritic stone olivine is an important constituent. The following mode is given by G. P. Merrill :—

pyroxenes	51·62	nickel-iron	2·59
olivine	40·48	troilite	4·43
chromite	·80	schreibersite	·08

The Goalpara stone (achondrite), as described by G. Tschermak, contains 62 per cent of olivine, 30 of pyroxene (clino-bronzite according to Prior), and 8½ per cent of nickel-iron. A still higher content of olivine appears in the Ornans stone, which F. Pisani has analyzed with the following result :—

olivine	75·10	nickel-iron	1·85
pyroxene	15·26	chromite	0·40
troilite	6·81		

Finally, in Chassigny, we have an example of a stone that is almost entirely composed of olivine (rich in ferrous oxide) with just a little chromite and pyroxene.

METEORITES CONTAINING FROM 10 TO 30 PER CENT OF NON-SILICATES

We find an olivine-free example in the Khairpur stone, which belongs to Prior's group of enstatite chondrites. The mode given by Prior is :—

enstatite	60·31	chromite	·39
oligoclase	8·25	daubréelite	·29
nickel-iron	18·29	oldhamite	·34
troilite	7·03	graphite	·41
		rust, etc.	4·39

Stones that contain olivine are much commoner than the above type. Prior has shown that most chondritic stones approximate to one chemical and mineralogical type, of which the average composition is as follows:—

olivine	44	nickel-iron	9
bronzite	30	troilite	6
feldspar	10	chromite, etc.	1

How closely some of these stones, found in widely separated regions, resemble one another, and how nearly they approach the above average, may be shown by quoting Prior's own analyses of the Launton stone, which fell in England in 1830, the Amana or Homestead stone (Iowa, 1875), and the Baroti stone (India, 1910).

	Launton	*Amana*	*Baroti*
feldspar	10·82	9·76	10·25
olivine	41·91	40·15	42·12
bronzite	29·50	30·75	30·38
nickel-iron	8·48	10·99	8·91
troilite	5·98	6·25	6·79
apatite	·36	·22	·56
chromite	·45	·44	·27
magnetite, ilmenite	·33	—	·94

METEORITES CONTAINING FROM 30 TO 60 PER CENT OF NON-SILICATES

Among these meteorites Daniel's Kuil is olivine-free and Saint-Sauveur holds very little olivine. The first of these modes is by G. T. Prior, the second by A. Lacroix.

	Daniel's Kuil	*St. Sauveur*
feldspar	8·47	13·38
enstatite	55·76	41·25
olivine	—	1·40
nickel-iron	25·45	27·49
troilite	7·96	14·26
oldhamite	·86	1·15
daubréelite	·40	·22
chromite	·34	—
apatite	—	·34
graphite	·32	·10

In Hainholtz and Vaca Muerta, which hold almost equal parts of silicates and non-silicates and fall in the group of *meso-*

siderites, olivine is again very scantily represented. Both modes are by G. T. Prior.

	Hainholz	*Vaca Muerta*
anorthite	14·5	17
pyroxene	27	32
olivine	1·5	1·5
nickel-iron	46·5	41·5
troilite	4	2
schreibersite, etc.	2	3
chromite	·5	1
rust	4	2

The Mount Vernon meteorite is a pallasite unusually rich in olivine. W. Tassin gives its composition as follows:—

olivine	63·15	chromite	1·00
nickel-iron	33·12	troilite	·69
schreibersite	1·95	carbon	·09

METEORITES CONTAINING MORE THAN 60 PER CENT OF NON-SILICATES

A siderolite described by A. Lacroix from the Adrar desert falls within this range of composition, but is nearly free from olivine. The mode is:—

feldspar	1·7	nickel-iron	71·2
hypersthene	17·3	troilite	8·3
olivine	0·9	schreibersite	0·5

The Tucson iron still contains a little olivine and is therefore intermediate between the pure irons and the pallasites. J. L. Smith has given the following mode:—

nickel-iron	93·81	schreibersite	0·94
olivine	5·06	chromite	0·41

We conclude with a few examples of siderites containing no silicates at all. The Wichita octahedrite, according to E. Cohen's analysis, contains 85 per cent of kamacite, 2·6 of taenite, 6·3 of schreibersite, and 6 per cent of cohenite. For the Zacatecas iron, which is also an octahedrite, Cohen gives the following data:—

nickel-iron	95·04	nickel phosphide	1·08
schreibersite	3·06	carbon	0·03
troilite	0·26	chromite	0·07
daubréelite	0·02	residue	0·44

The Mart iron, with a still greater proportion of metal, also belongs to the octahedrites. Its composition has been determined by G. P. Merrill and H. N. Stokes to be :—

nickel-iron	98·31	troilite	0·05
schreibersite	1·06	chromite	trace

For the Coahuila iron, a hexahedrite, E. Cohen gives the following mode :—

nickel-iron	98·3	carbon	·011
schreibersite (rhabdite)	1·6	chromite, etc.	·003
daubréelite	·027		

The Pre-Geologic History of the Earth

When so much remains unknown about the recent history of the earth it would be foolish to speak with assurance of its pre-geologic history. Yet there is one aspect of the matter upon which all natural philosophers are agreed : the earth and the associated planets are so completely dependent upon the sun that we cannot conceive of them except as its offspring. We believe that the history of the earth began with its separation from the sun. It is not necessary to know how the separation took place; nevertheless the Tidal Theory of Chamberlin and Moulton, with the modifications introduced by Jeans and Jeffreys, is considered to afford an adequate explanation.

This theory supposes the sun to have been deprived of a small part of its gaseous envelope by the attraction of a close-passing star. The extruded filament of gas, too tardy to catch the fast-receding star and too elongated to be wholly drawn back into the sun, began to condense in consequence of loss of heat by radiation and adiabatic expansion. J. H. Jeans has deduced by mathematical reasoning that condensed portions of the gaseous mass would separate in succession from the main body, thereafter to carry on an independent existence as planets. H. Jeffreys has computed the minimum radius of a planet formed in this way and finds it to be 1400 kilometres. This result is important, for it means that asteroids whose dimensions fall below this limit, and all such minute bodies as meteorites, can not have been formed by direct condensation from the gaseous state. It follows that meteorites must either be fragments of larger solid bodies such as former planets or else they must have been formed by the aggregation of small solid particles. This agrees with the observation that the texture of the stony meteorites resembles that of a volcanic tuff.

Jeffreys shows further that the cooling-history of the new-born planet must depend upon its size. A large body of gas composed of sufficiently dense elements will be held together by gravitation and will cool down by radiation from the outside. Drops of liquid will be formed which will fall like rain-drops towards the centre of gravity of the mass, where they will build up a liquid nucleus having its heaviest components at the centre. Smaller bodies of gas will behave differently, for gravity may be insufficient to prevent their expansion. The form produced by continued expansion, with rapid loss of heat, will be a hollow sphere composed of liquid drops, enclosing some uncondensed gas. At this stage two courses of development become possible; either expansion will continue until the drops are dissipated in space, or else expansion will cease and gravity will take charge and draw the drops together.

Jeffreys claims that by any mode of aggregation *a planet must pass through a liquid stage,* for the drops falling towards the centre of gravity will pass through a body of gas that is hotter than themselves, so the temperature of the drops must be raised by conduction as well as by friction and mutual impact. Once a liquid nucleus was formed there would be direct condensation of gas upon it. For these reasons it seems to be a valid contention that "any matter aggregated into the nucleus would be liquid."

The temperature at the base of the sun's photosphere is calculated to be about 6300° Abs., which is far above the boiling point of any of the chemical elements. Iron, nickel, and cobalt all boil below 3000° C. Only carbon and a few of the rare metals, such as tantalum, platinum, and tungsten have higher boiling points than this. The chemical composition of the envelope is known from the relative intensity of the spectral lines; it appears that the most abundant elements in the envelope are calcium, iron, hydrogen, sodium, nickel, magnesium, cobalt, silicon, aluminum, titanium, manganese, in that order. These are just the elements that are commonest in the earth-crust, with the exception of potassium which seems to play a minor part in the sun. At the enormous temperature of the sun's surface these elements are uncombined. Oxidation and other chemical reactions took place in the condensing body of gas as its temperature fell towards 3000°, and this must for a time have neutralized the loss of heat by radiation, thus giving time for the orderly condensation of the mass. Of the elements named above, magnesium has the highest energy of oxidation and it would be the first to combine with oxygen. It would be followed by calcium and aluminum, then by sodium, potassium, silicon, manganese, carbon, hydrogen, iron, cobalt, nickel, in that order. The temperatures at which these oxides would condense to the liquid state are unknown, but the melting-points and boiling points of some of the oxides in the pure state, under atmospheric pressure, are as follows:—

	m.p.	b.p.
MgO	2800°	—
CaO	2570°	2950°
$MgAl_2O_4$	2135°	—
Al_2O_3	2040°	2250°
SiO_2	1710°	2230°

Of the common silicates the most infusible are forsterite, 1890°, and mullite, 1810°.

It is an open question whether the nucleus of heavy metals which escaped oxidation would be covered by a zone of liquid oxides (magnesia, lime and alumina) or by a zone of liquid silicates. The only scrap of direct evidence bearing upon this question is given by the pallasite group of meteorites; these consist of a sponge of metallic iron enclosing grains of magnesium silicate (olivine), not of magnesium oxide. We may therefore assume that the liquid zone overlying the iron core of the earth consisted even in the beginning of liquid silicates, not of liquid oxides.

The radius of the iron core seems to be satisfactorily indicated by seismological data, for there is abundant evidence of the existence of a nucleus which strongly reflects and refracts earthquake waves. The diameter of this nucleus, according to J. B. Macelwane, is probably between 53 per cent and 58 per cent of that of the whole earth. Other surfaces of reflection and refraction have been suspected at various levels in the silicate shell, but the evidence for these is by no means so consistent and definite as for the surface just above the half-radius, which we interpret as the top of the iron core.

The composition of the silicate shell of the earth has the greatest interest for the petrologist, yet he has no means of satisfying his curiosity about it except by turning once more to the meteorites. If from the composite analysis of all meteorites, quoted on page 344 (after O. C. Farrington) we subtract all iron, nickel, cobalt and sulphur, the remainder has the following composition:—

SiO_2	58·2
MgO	33·1
Al_2O_3	3·81
CaO	3·34
Na_2O	1·20
K_2O	·26
	99·91

In terms of minerals this corresponds to enstatite 65, diopside 10, olivine 10, plagioclase 15 parts. A direct study of the composition of chondritic meteorites was made by G. T. Prior with the following result: enstatite 30, olivine 44, feldspar 10, ores 15 parts; if the ores are deducted this becomes enstatite 35, olivine 52, plagioclase 12. Of these two computations Farrington's is perhaps the more useful, since Prior has omitted the gabbro-like ophitic stones. But whether we consider Farrington's data or Prior's the composition indicated is that of *a slightly feldspathic peridotite or picrite in which magnesium greatly exceeds calcium and all other metals.*

One may question the propriety of basing conclusions about the composition of the earth upon the evidence of the stony meteorites, of which we know only a few tons in all; yet the conclusion just reached is not out of harmony with the data of petrology which make it probable that the magmas of deepest origin are rich in magnesia. Furthermore one cannot fail to be impressed by the resemblance between the structure of chondritic meteorites and the picture which we have formed of droplets of rapidly congealed liquid which escaped the control of gravity during the expansion of the gas-cloud, and were afterwards aggregated in the solid condition. Let us suppose, then, that the iron core of the growing earth was surrounded by a liquid shell composed largely of magnesia and silica and having approximately the composition of enstatite-peridotite. The freezing of the system $MgO\text{-}SiO_2$ is completely known from the work of Bowen and Andersen, and the slightly feldspathic peridotite described above comes near enough to this system to follow a similar course of crystallization. The earliest crystals formed would be of olivine (nearly pure forsterite), and the temperature of commencing crystallization somewhat below 1600°. Bowen has shown that in a liquid of the composition diopside 56, enstatite 44, the early crystals of olivine sink; it need not be doubted, then, that olivine crystals would sink in such a liquid as we are discussing. Having sunk a certain distance they would be remelted; thus in the course of time the surface portion of the earth's silicate shell would be largely deprived of magnesia and would therefore be relatively enriched with lime, alumina and alkalis. But the sinking of olivine crystals out of a magma which contains enough silica to have formed enstatite leaves that magma relatively enriched with silica; for instance, the 65 per cent of enstatite indicated by Farrington's data correspond to 45 per cent of olivine and 20 per cent of silica. Thus, as long as olivine continues to crystallize, the silicate shell must become ever more and more siliceous and feldspathic in composition towards the top, and more and more magnesian downwards, with no sharp discontinuity anywhere. In the case of the hypothetical magma that we are discussing, the sinking of 10 per cent of olivine would leave the magma exactly saturated with silica (using Farrington's data). If 20 per cent of olivine settled out, the magma would be left with 5·6 per cent of excess silica; if 30 per cent settled out, the magma would contain 13 per cent of excess silica. Even when the temperature of the liquid sank so far that enstatite began to crystallize instead of olivine, it is likely that crystal settling would still go on and that the upward concentration of feldspathic molecules and silica would be continued.

The crystallization of a calcic plagioclase at a still later stage would leave the surface liquid relatively enriched with soda and still more so with potash.

It seems to follow from these considerations that the granitic shell of the earth might have been formed in consequence of crystallization and crystal-settling in the supposed enstatite-peridotite or meteoritic earth-magma. On paper, a shell of this magma less than 100 miles thick could furnish the materials for a "granitic" shell 20 miles thick, within which there would be a gradual passage from true granite at the surface to tonalitic and gabbroic rocks below. The top of the underlying liquid zone would probably be more feldspathic than the original peridotite magma on account of some sinking and remelting of calcic plagioclase; thus a zone of olivine-basalt might intervene between the solid crust and the underlying enstatite-peridotite magma. There would be no solid peridotite (except perhaps locally and temporarily) because the sinking crystals of olivine and enstatite would of necessity be remelted as soon as they reached a certain limiting depth, determined by the temperature gradient.

An alternative line of speculation is opened by J. W. Greig's discovery of *liquid immiscibility* in melts of silica with magnesia, lime, and iron oxide, at temperatures near 1700°. No immiscibility was detected in alkalic or aluminous melts at that temperature, but it is not known how these melts would behave at still higher temperatures. If the newly condensed shell of liquid silicates about the growing earth were capable of developing two immiscible liquid phases, one richer in alkalis and silica and the other richer in lime and magnesia, then the problem of the earth's granitic shell would be solved.

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(For all other references and descriptions, see Farrington (2), Flight, or Wülfing.)

PART II

ROCK DESCRIPTIONS, ARRANGED IN ACCORDANCE WITH THE CLASSIFICATION DESCRIBED IN CHAPTER XIV

The strongest recommendation of an artificial system (besides its approaching to a natural method) is that it shall be capable of easy use.

—Wm. Whewell

PART II

CHAPTER XX

THE OVERSATURATED ROCKS (DESCRIPTION)

KEY TO THE CONTENTS OF THIS CHAPTER

A.—Mineralogy of the oversaturated rocks. *p.* 363
B.—Classification and Nomenclature. *p.* 368
C.—Granite and Granodiorite. $XO\alpha$, $XO\beta$. *p.* 371
D.—Rhyolite. $DO\alpha$, $DO\beta$. *p.* 377
E.—Tonalite. $XO\gamma$, $XO\delta$ *p.* 382
F.—Dacite. $DO\gamma$, $DO\delta$. *p.* 385

A.—THE MINERALOGY OF THE OVERSATURATED ROCKS

THE proportion of quartz in an oversaturated rock ranges from almost nothing to a maximum which approaches fifty per cent; and of what we may call average granite, quartz forms between twenty-five and thirty-five per cent.

The crystals of quartz in plutonic rocks are sometimes enclosed in the outer zones of the feldspar crystals, but the bulk of the quartz occupies the interstices between the feldspars and the other constituents, and it is certain that in all deep-seated rocks quartz is one of the last minerals to crystallize. It has been shown by Day, Sosman, and Hostetter that quartz undergoes a remarkably rapid contraction below 575° C.; from this point down to 500° C. the contraction amounts to 2·4 per cent of the volume of the quartz. This must be an important cause of the production of shrinkage cracks in granites, and it may explain why residual veins are especially characteristic of oversaturated rocks.

Tridymite, which is the stable form of silica between 870° and 1470° C., has not been recognized in deep-seated rocks, but it is found in some lavas and also in foreign rock fragments which have been enveloped in lava and baked by its heat. Minute hexagonal plates of tridymite are sometimes seen in cavities in porous lava, and it is likely that this mineral, as well as the still rarer cristobalite, has been formed by some special reaction in which hot gases took part, and that it is not a direct product of magma at all. But all observers are not agreed upon this matter, and a summary of divergent opinions has already been presented (see page 27).

Although the characteristic mineral of the oversaturated rocks is quartz, the most abundant mineral is certainly feldspar. This is usually a moderately potassic kind, either orthoclase (microcline), anorthoclase, or microperthite, but a plagioclase feldspar, ranging in composition from albite to labradorite, is sometimes present to the almost complete exclusion of orthoclase. Whether potash-feldspar or soda-feldspar predominates in granite as a whole, it is hard to say. The answer clearly depends on the definition of granite. The average composition of 236 analyses of granite, computed by R. A. Daly, indicates a slight excess of albite over orthoclase; but an average for the granitic rocks of Japan, computed by Suzuki and Nemoto, shows albite in considerable excess over orthoclase. On the other hand a collection of 137 analyses of the granites of Sweden, made by P. J. Holmquist, shows that varieties with dominant orthoclase and those with dominant albite are almost equal in number.

It is certain that potash feldspar, whether free or held in solid solution in the outer zones of the plagioclase crystals, is never wholly absent. There is a distinct relation between the amount of quartz and the amount of potash feldspar present in a rock. As the plagioclase becomes more and more calcic, quartz dwindles, and when anorthite exceeds albite in the plagioclase there is seldom more than a trace of quartz left. Thus in the labradorite- and bytownite-bearing rocks (gabbro, anorthosite, etc.) oversaturation is not a common condition.

The sympathetic variation of quartz and orthoclase, and their antipathetic relation towards anorthite, are illustrated by the following series of analyses of closely related rocks from Idaho (W. Lindgren).

	(1)	(2)	(3)	(4)
Quartz	29·20	28·04	25·00	8·45
Or	18·07	15·84	11·21	7·57
Ab	27·19	33·54	30·25	26·20
An	9·53	11·15	13·88	20·45

Duparc and Borloz go so far as to maintain that the association of calcic plagioclase (labradorite to anorthite, without mantles of more sodic feldspar) with much quartz demonstrates that the rock containing these minerals is not a simple magmatic product. They have described various rocks from Bolivar, Venezuela, which hold quartz, basic labradorite, diopside and hornblende, and which might pass for tonalites, although they have actually been formed by reaction between granite and limestone.

The great variation that may be found in the composition of the feldspar in different parts of one continuous body of "granite" is shown by the following table, compiled by W. J. Sollas, relating to the soda-granite of the Wicklow Mountains, Ireland.

Croghan Kinshela	1	part	orthoclase	to	25	parts	albite
Rockabill	1	,,	,,	to	1·2	,,	,,
Blackstairs	1	,,	,,	to	1·1	,,	,,
Poulmounty	1	,,	,,	to	1·2	,,	oligoclase
Three Rock Mountain	1	,,	,,	to	1·2	,,	,,
Carnsore	1	,,	,,	to	1·4	,,	,,
Ballinaclash	1	,,	,,	to	3·9	,,	,,
Aughrim	1	,,	,,	to	5·3	,,	,,
Ballyknockan	1	,,	,,	to	1·2	,,	andesine
Ballinamuddagh	1	,,	,,	to	1·6	,,	,,
Coolboy	1	,,	,,	to	3·9	,,	,,
Cushbawn	1	,,	,,	to	5·1	,,	,,

A table illustrating the variation within the Vermilion granite, Minnesota, was given on page 250.

Since feldspar makes up the greater part of all oversaturated rocks, the arrangement of the feldspar crystals mainly determines the texture of the rock. We have to recognize three principal varieties of texture, as follows:—

(1) The crystals are large, but often unequal in size, and are closely packed together without definite arrangement. Owing to mutual interference, only a minority of them have developed their ideal shape. This is ordinary "granitic texture," and is characteristic of the majority of plutonic rocks of all kinds.

(2) The feldspars are more tabular in shape, owing to the large development of the pinacoids, and they and the mica scales or hornblende needles are arranged in roughly parallel position. This is "parallel texture."

(3) Large crystals (insets) of orthoclase or microcline are embedded in a groundmass of smaller feldspars, quartz and mica. This is "porphyritic texture." The feldspar of the groundmass is sometimes different from that of the insets; thus in the well-known porphyritic granite of Shap Fell, West-

morland, the insets are pink orthoclase and the feldspar of the groundmass is white oligoclase.

The *heavy constituents* of oversaturated rocks include white and black micas, various kinds of hornblende and pyroxene, tourmaline, garnet and topaz. Sphene, zircon, apatite, fluorspar and magnetite are usually present in small quantity, and so, at times, are pyrite and other sulphides. Fayalite (iron olivine) makes an occasional appearance in oversaturated rocks, and eudialyte, astrophyllite, allanite, and other minerals which contain the oxides of the rare metals are often detected in small quantity. Andalusite, cordierite and sillimanite are not normal minerals of eruptive rocks; when they appear in granite there is reason to suspect that assimilation of aluminous sediments has taken place. Nevertheless, E. S. Hills claims that both andalusite and sillimanite are sometimes primary minerals of granite. J. J. Runner has recorded andalusite and sillimanite in pegmatites of the Black Hills, S. Dakota, and E. W. Heinrich has described sillimanite and kyanite-pegmatites in Montana.

The mica of oversaturated rocks includes ordinary muscovite, the lithia-micas (lepidolite and zinnwaldite), and several varieties of biotite. Muscovite is seldom found in large quantity except in rocks with an abundance of free quartz, that is, in granites of a very acid type, and it is developed in large crystals in pegmatite veins. Lithia-mica is almost confined to the aplitic and pegmatitic facies of the more acid granites, and to contact zones. Biotite is seldom entirely absent from quartz-bearing rocks; it is a deep brown variety, owing its colour to ferric oxide, but it becomes greenish with advancing alteration. The presence of minute enclosures of apatite, zircon, xenotime, allanite, and other scarce minerals is a peculiarity of the biotite in most granites. Round the enclosed grains, some of which are radioactive, the mica is generally more deeply coloured than elsewhere; these "pleochroic haloes" have afforded a means of measuring the age of the host mineral.

The quantity of mica in an eruptive rock seldom exceeds fifteen per cent; in average granite it may be said to range from about three to ten per cent. Muscovite is generally subordinate to biotite, and it is only in some very acid granites and pegmatites that the relation is reversed. In lavas, mica is even scarcer than in the plutonic rocks; many rhyolites and dacites carry no more than a trace of biotite.

Green hornblende very often accompanies biotite in the plagioclase-bearing granites, especially the more calcic ones, and it may be associated with pale green diopside. Orthopyroxene (enstatite or hypersthene) is occasionally present even in potassic granites and rhyolites, but only in small quantity. In the charnockite series of Uganda, A. W. Groves claims that hypersthene has been formed from amphibole and clinopyroxene by plutonic metamorphism. At the calcic end of the oversaturated rock series we find rocks consisting of plagioclase and augite or hypersthene, with only a trace of quartz and orthoclase left.

The soda-pyroxene, ægirine or acmite, makes its appearance only in the most alkaline varieties of granite, from which lime-feldspar is almost or quite absent. Ægirine often exhibits a partial transformation into one of the blue soda-hornblendes; thus a crystal may have a core of ægirine and a mantle of riebeckite or arfvedsonite. These soda-rich minerals as a rule build spongy crystals which surround and incorporate quartz and feldspar grains; they have clearly been introduced when the rock was essentially solid. In certain cases in Finland and Scandinavia an ordinary granite has been impregnated with ægirine by emanations from a younger soda-rich intrusion.

Tourmaline is only occasionally an important constituent of granitic rocks; like muscovite, it is practically confined to the more quartzose varieties, and it is rare in lavas. It crystallizes at a late stage, and much of it is formed at the expense of earlier feldspar and biotite. The usual variety is the black iron-tourmaline, which appears brown or blue in thin section; but pink, green, or colourless alkali-tourmaline of hydrothermal origin is found in some pegmatite veins, often in association with lithia-mica. Tourmaline is almost never associated with hornblende or pyroxene.

Garnet is not an abundant mineral in granitic rocks, although individual occurrences are fairly numerous. A content of 2 to 4 per cent of garnet is recorded by T. T. Quirke in granites of French River, Ontario. Garnet is sometimes prominent in aplite and pegmatite veins, and occasionally in acid lavas. The variety is almandine or spessartite.

Fayalite is a very rare constituent of these rocks. It has been found in a few cases only in drusy cavities in granite, in pegmatite veins, and in cavities in lava. In central Wisconsin, however, fayalite appears in quantity ranging from 1 to 5 per cent, as a normal constituent of certain hornblende granites and syenites (S. Weidmann). L. Hawkes has found fayalite in small quantity among the insets in an Icelandic lava, and V. Sobolev has described a "quartz-fayalite diabase" from the Tunguska River, Siberia, which contains 8 per cent of fayalite along with 26 per cent of micropegmatite.

Topaz, too, is largely confined to cavities and to pegmatite veins, where it is accompanied by muscovite, tourmaline, and sometimes cassiterite. In some rocks topaz replaces the feldspar, and there have been produced in this way rocks that consist almost wholly of quartz and topaz.

Epidote or clino-zoisite is generally thought to be a secondary constituent, but C. R. Keyes has described well-formed crystals of this mineral in the granites of Maryland, and considers them to be primary constituents of the rocks. Clino-zoisite, and also allanite (cerium-epidote), are very common in the granites of New England. A. Laitakari has described epidote-granites in Finland.

B.—The Classification and Nomenclature of the Oversaturated Rocks

The long-established custom of naming granitic rocks according to the particular species or mixture of feldspars they contain, to the neglect of characters possessing greater genetic significance, has given us a score of loosely defined rock names for most of which it is hard to find any justification. To granite and quartz-diorite there have been added tonalite, adamellite, granodiorite, quartz-monzonite, banatite, farsundite, trondhjemite, yosemitite, opdalite, besides others that have a special textural connotation. For the dyscrystalline rocks we have rhyolite, dacite, rhyodacite, liparite, pantellerite, comendite, all defined in terms of feldspar, besides a number of purely descriptive names. Since there is no clearly established physico-chemical boundary anywhere in the orthoclase-albite series of feldspars, and certainly none in the albite-anorthite series, it follows that the supposed "species" of granite and rhyolite are lacking not only in definition but also in physico-chemical significance. This shows that the feldspars do not furnish an ideal basis for the classification of rocks, but it is too late to do anything about that. In the absence of natural boundaries we must do the best we can with artificial ones; and what we chiefly require of an artificial boundary is that its position shall be clearly established. If we proceed on the lines laid down in Chapter XIV, we arrive at a very simple grouping of the oversaturated rocks which does away with the need for most of the names cited above.

Eucrystalline Division

$Or > An$ — GRANITE	$Or > Ab$ — Potash-granite	
	$Ab > Or$ — Soda-granite	
$An > Or$ — TONALITE	$Ab > An$ — Soda-tonalite	
	$An > Ab$ — Lime-tonalite	

Dyscrystalline Division

$Or > An$ — RHYOLITE	$Or > Ab$ — Potash-rhyolite
	$Ab > Or$ — Soda-rhyolite
$An > Or$ — DACITE	$Ab > An$ — Soda-dacite
	$An > Ab$ — Lime-dacite

In this arrangement the troublesome rocks which contain two kinds of feldspar, and which have given rise to so many new names, fall almost entirely into soda-granite. Most of these rocks contain a moderately calcic plagioclase (oligoclase to andesine) together with a potassic feldspar, but a few have a plagioclase that is nearly pure albite. To separate these two facies, the procedure adopted in the first edition of this book was to divide the soda-granite field by a boundary corresponding to the composition Or_3An (1st edition, *Fig.* 26). This gives two nearly equal fields which were called the albite-granite field and the granodiorite field respectively. The latter takes in most of the well-described examples of granodiorite, but the so-called albite-granite field is too broad to agree with the usual idea of an albite-granite. It may be better to confine the name albite-granite to rocks in which the plagioclase is literally albite (Ab to Ab_9An_1), that is, to a narrow strip along the Or-Ab boundary of the field (*Fig.* 35), and to apply the name granodiorite to all the remainder of the soda-potash field. That is the course that will be followed in this volume.

Within each subdivision of granite and rhyolite four reaction-types are possible, as explained in Chapter XIV. If "specific" names are to be used in the system, it is at this stage that they should be introduced. There would then be four species of potash-granite, four of soda-granite, four of potash-rhyolite and four of soda-rhyolite, each with its distinctive name. It is not proposed to take this step, however, since it is generally agreed that new names are an abomination and the same result can be achieved by means of symbols. Thus a peraluminous potash-granite is symbolically XOpα, a peralkaline soda-granite is XOkβ, and so on. Among the tonalites and dacites there will be only three species in each sub-group, since the characteristic minerals of the peralkaline type are not compatible with lime-rich feldspar.

It is hardly necessary, in naming oversaturated rocks, to take the colour index into consideration, for mesotype and melanocratic rocks are so uncommon among granites that no special recognition has ever been given to them. The use of a symbol which includes the colour index will generally be sufficient to characterize any such rocks.

There remains the question of the proportion of quartz and the boundary between granite and syenite. Although according to the accepted definition syenite contains "neither quartz nor feldspathoid," yet in practice the name has constantly been applied to rocks such as the Plauen syenite, which holds 5 to 10 per cent of quartz, and even (in the form quartz-syenite) to rocks with 15 to 20 per cent of quartz. To a chemist it seems more important to know that a solution is acid than to know the precise degree of acidity; so until somebody gives a sufficient reason why a rock with 15 per cent of quartz should have a different name from one with 25 per cent of quartz, the writer will call them both granite. But to avoid a complete break with the past he agrees that rocks with no more than *ten per cent* of free silica (as in the Plauen syenite) may be placed in syenite rather than in granite. The same boundary will apply to tonalite and diorite. The slightly oversaturated rocks will therefore be discussed along with the saturated ones in a later chapter. They will be distinguished by symbols in which a small letter o takes the place of a capital O.

C.—GRANITE AND GRANODIORITE. XOα, XOβ

Peraluminous Type.

This type is characterized by the presence of alumina in molecular excess of the sum of potash, soda, and lime. The excess goes typically into muscovite, but if enough magnesia is available, biotite may partly or wholly take the place of white mica, and in exceptional cases a lithia-mica is produced. Other peraluminous minerals are tourmaline, topaz, and certain kinds of garnet (almandine, spessartite).

The well known granites of Cornwall are muscovite-biotite granites with subordinate tourmaline and topaz. An admirable quantitative examination of the St. Austell granite mass, carried out by W. A. Richardson, shows that the average composition of all varieties in that area is as follows:—

quartz	32·6	tourmaline	1·8
orthoclase	34·5	topaz	1·4
plagioclase	19·2	fluorspar	0·4
muscovite	4·5	magnetite	0·3
biotite lithia-mica	4·7	apatite	0·3
		cordierite	0·3
		Colour index	(13)

Within this single granite mass, which has an area of 32 square miles, the chief constituents show the following fluctuations:—

quartz, from	20 to 48	white micas, from	0 to 15
orthoclase, from	17 to 52	tourmaline, from	0 to 7½
plagioclase, from	9 to 31	topaz, from	0 to 7
dark micas, from	0 to 15	Colour index	(4 to 22)

It follows from the figures given for orthoclase and plagioclase that some part, even of this most typical of granites, is soda-granite. The published analyses of Cornish granites are about equally divided between potash and soda varieties, but a first-class analysis of the Land's End granite, by W. Pollard, shows 34 per cent of normative orthoclase to 23 per cent of normative albite.

A. Brammall and H. F. Harwood have given a very detailed account of the granites of Dartmoor, England, and their contact phenomena. The quartz content of these granites ranges between 20 and 40. The principal feldspar is microperthite, but some sodic feldspar is always present too. In the calculated feldspar orthoclase typically exceeds albite, so the rock as a whole is a potash-granite. Biotite is the usual dark mineral, forming 5 to 10 per cent of the rock, with subordinate muscovite and occasionally a little tourmaline or garnet. In short this is a typical example of the peraluminous type of granite, and Brammall and Harwood have furnished the most complete petrological and chemical description that has ever been made of such a rock.

The composition of the Vermilion batholith, in Minnesota, has been studied by F. F. Grout. Nine-tenths of this mass, which covers an area of more than 2,000 square miles, is a simple biotite-granite with the following average composition :—

quartz	25	oligoclase	20
orthoclase	30	biotite	2
microcline	20	accessories	3
		Colour index	(5)

Towards the margin this granite develops plagioclase and hornblende and passes over into syenite, diorite and hornblendite.

A soda-granite that has been very completely investigated is the beautiful porphyritic granodiorite of Shap Fell, Westmorland, which has insets of pink orthoclase in a light grey groundmass. The composition has been worked out by A. Holmes, who finds that the insets make up one-fifth of the rock and that the total composition is quartz 24, orthoclase 36·2, oligoclase 33·6, biotite 5·8. Making allowance for the albite that is held in solid solution in the orthoclase, the actual composition of the feldspar is Or 24, Ab 34, An 12.

Peraluminous varieties preponderate among the granites of New England, which range from moderately potassic to strongly sodic. M. Billings has made a quantitative study of the granites of New Hampshire, from which we take the following examples :—

	Owl's Head granite.	*Bethlehem gneiss.*	*Kinsman quartz-monzonite.*
quartz	30	30	29
potash-feldspar	49	19	19
oligoclase-andesine	10	31	25
muscovite	2	3	8
biotite	7	17	19
accessories	2	—	—
Colour index	(11)	(20)	(27)

F. Chayes has given the following modal analysis of 21 thin sections of the Barre, Vermont, granite:

quartz	27·2	muscovite	8·3
microcline	19·4	biotite	8·1
plagioclase	35·2	rest	1·8
		Colour index	(18)

J. A. Shimer has given 18 modal analyses of granites from Maine, New Hampshire, Massachusetts, Vermont, and Connecticut. The average composition of the feldspar in these granites is orthoclase 27·4 per cent, plagioclase (oligoclase-andesine) 27·8 per cent.

Metaluminous Type.

Granites and rhyolites of this type contain one or other of the following mineral associations: biotite together with hornblende, augite (or primary epidote); hornblende together with augite (or epidote); hornblende alone (or epidote alone). Fayalite is occasionally present in small quantity, and in a soda-granite at Wausau, Wisconsin, it amounts to about 5 per cent (S. Weidmann). The metaluminous type is much rarer among the potash- than among the soda-granites. An important example is the red granite of the Transvaal. P. A. Wagner gives the following mineralogical analysis of this rock:—

quartz	33·3	biotite	0·5
orthoclase	30·1	hornblende	0·75
albite	25·6	ore	0·3
anorthite	4·0	apatite, etc.	0·4
		Colour index	(2)

The central granites of the Vosges Mts. is also a metaluminous potash-granite; it has been studied by Friedlaender and Niggli, who give the following mode:—

quartz	24·3	biotite	17·0
orthoclase	31·5	hornblende	6·2
albite-oligoclase	19·1	accessory	1·9
		Colour index	(25)

The gray granite of Ontario contains orthoclase, albite, and oligoclase in roughly equal amounts, with biotite, green hornblende, and frequently ortho- or clino-pyroxene as well (Adams and Barlow). The red granite of the same area carries a smaller proportion of dark minerals. An analysis of the red granite of Methuen township affords the following approximate estimation of the mineralogical composition: quartz 28, orthoclase 18, albite 42, anorthite 5, biotite, hornblende and ore (together) 7 per cent.

The most characteristic examples of *granodiorite* are metaluminous. The rock of the Boulder batholith, at Butte, Montana, is typical; it has been described by (1) W. H. Weed and (2) J. P. Iddings, and the two descriptions clearly refer to slightly different facies.

	(1)	(2)
quartz	23·7	22·5
orthoclase	19·9	17·6
albite	23·0	25·5
anorthite	11·5	17·0
biotite	4·2	9·8
hornblende	15·3	4·4
diopside	—	3·3
ore	1·2	0·7
sphene and apatite	1·3	0·1
Colour index	(22)	(18)

Subaluminous Type.

Subaluminous granite is very uncommon, for although enstatite or diopside is often present, as in the gray granite of Ontario, it is nearly always accompanied by hornblende and biotite.

The pre-Cambrian batholiths of Blue Ridge, Virginia, are composed of hypersthene-granite (which has been called syenite in spite of the presence of 15 to 30 per cent of quartz). The feldspar is orthoclase and andesine. Hypersthene is the principal melane, but there is some clino-pyroxene and sometimes hornblende too (Watson and Cline).

Hypersthene-granite is the acid member of the "charnockite series" of Madras. This rock is mainly composed of quartz and microcline, with a little plagioclase and only 3 per cent of hypersthene (H. S. Washington, 1). F. D. Adams has studied the same series of rocks in Ceylon and gives the following modes :—

quartz	23·7	29·5
orthoclase	30·0	20·0
albite	20·4	23·6
anorthite	11·1	11·7
diopside	3·7	7·5
hypersthene	8·8	
ore, apatite	1·8	8·4
Colour index...	(14)	(16)

The subaluminous type is not uncommon among dyke-granites. The description, by A. Harker, of the granophyre of Carrock Fell, Cumberland, is applicable to many rocks of this kind. The approximate composition of the rock is as follows:—quartz 21, orthoclase 21, albite 43, anorthite 3·7, pyroxene 8·6 per cent. The groundmass of the rock shows every stage of gradation "from a coarse and irregular pegmatite to spherulitic intergrowths of a delicacy surpassing the resolving power of the microscope." Harker observes that granophyres (dyke-granites with graphic texture) generally carry pyroxene, and that in Carnarvonshire the typical granophyres have augite without biotite; the coarser granophyres and porphyries hold both augite and biotite; and the granites have biotite alone.

Peralkaline Type.

Granites of this type hold soda and potash in molecular excess over the available alumina; whenever this happens the potash seems to exercise the stronger claim to combine with alumina, and it is always soda that is left in excess and compelled to enter soda-iron pyroxene, soda-iron amphibole, or uncommon titanosilicates and zirconsilicates such as astrophyllite, eudialyte and ænigmatite. The commonest of these soda minerals is ægirine. Riebeckite and arfvedsonite are sometimes primary minerals, but in other instances they have been formed by the transformation of the original ægirine. In certain granites of Finland and Scandinavia, ægirine is said to have been introduced by solutions escaping from adjacent soda-rich magma.

Among the most characteristic rocks of this type are the riebeckite-ægirine granites of Quincy, Massachusetts, which have been described by C. H. Warren. A typical example of the Quincy granite gave quartz 33, sodipotassic feldspar 55, and ægirine and riebeckite 10 per cent, with small quantities of apatite, zircon and astrophyllite. A darker variety from Copper Mine Hill yielded 7 per cent of ægirine and 8·5 per cent of riebeckite. The Quincy granite is cut by a younger mesotype granite (the Squam River granite) which contains 40 per cent of soda-hornblende (catophorite) and brown biotite, along with 35 per cent of sodipotassic feldspar and 25 per cent of quartz (C. H. Clapp). So many beautiful varieties of peralkaline granite occur in Massachusetts that this region, rather than the Christiania district, deserves to be the "type locality" for such rocks.

The "ekerite" of W. C. Brögger is an arfvedsonite-ægirine granite, and his "grorudite" is an ægirine-microgranite which forms many dykes in the Christiania district of Norway. The composition of a typical example is as follows:—quartz 23, soda-potash feldspar 53, ægirine and arfvedsonite 22 per cent. Similar rocks have now been recognized in many parts of the world. When instead of ægirine, riebeckite is developed, the names "paisanite" and "dahamite" have been used instead of grorudite; but so many names are not required for rocks which are all of one chemical type.

The name "rockallite" was given by J. W. Judd to a mesotype soda-granite with 43 per cent of ægirine, 26 per cent of soda-potash feldspar, and 30 per cent of quartz. This curious rock was thought to form the entire islet of Rockall, in the North Atlantic, but Lacroix has shown that rockallite only constitutes occasional dark patches (segregations) within a more leucrocratic ægirine granite of which the islet is entirely composed.

The riebeckite-granites of Nigeria have been described by A. D. N. Bain. The main variety has the simple composition:—quartz 37·8, soda-rich microperthite 57·9, riebeckite 4·2. Other varieties contain in addition to riebeckite some proportion of iron-rich biotite, thought to have been formed at the expense of riebeckite; and still others contain fayalite.

D.—RHYOLITE, DOα, DOβ

Most oversaturated lavas, and many of the dyke rocks, are very ill-crystallized. A few insets of quartz and feldspar may be present and these are easily identified, but the bulk of these rocks is cryptocrystalline to glassy. Purely qualitative methods of petrography can do little to resolve such rocks as these, and so far as mineralogical names have been given to them they have been based on the insets alone, with the assumption that the glassy or cryptocrystalline base of the rock, which possibly constitutes ninety per cent of the whole, has the same composition as the insets. This assumption is quite unjustified and may be seriously wrong.

Many petrographers believe that no systematic classification of cryptocrystalline and glassy rocks is possible without chemical analysis. Nevertheless, it is quite possible to determine the "potential mineralogical character" of a glassy rock without analyzing it, if the specific gravity and the refractive index of the glass can be ascertained. It has been shown by C. E. Tilley and W. O. George that the refractive index of the rhyolite glasses lies between 1·48 and 1·51, and the density of the commoner varieties between 2·1 and 2·4. The peralkaline type, which holds heavy soda-iron silicates, may be 2·5 or even over. Simple qualitative tests will quickly show whether soda or potash predominates in the glass, and with this information it will often be possible to classify the rock. It is useful to know that the great majority of glassy rocks are oversaturated, about three-fifths of them being soda-rhyolites, one-fifth potash rhyolites, and one-fifth dacites.

Peraluminous Type

Few rhyolites have been adequately described from the viewpoint of quantitative mineralogy. Hungary is the classic region for such rocks in Europe, and A. Vendl has made a study of the rhyolites of that region. They are hypocrystalline to glassy rocks, and they range in composition from highly potassic to moderately sodic. Many of them show little insets of biotite and the majority are at least potentially peraluminous as one may judge from the following norms.

	Potash-rhyolite.	*Soda-rhyolite.*	
quartz	22·5	44·4	33·9
orthoclase	57·8	18·9	22·2
albite	5·2	26·2	24·1
anorthite	—	5·8	12·0
corundum	6·1	2·5	1·8
hypersthene	1·1	·6	1·1
iron ores	5·4	2·3	2·4

Both potash- and soda-rhyolites have been described in the Creede district of Colorado, by Emmons and Larsen. They are porphyritic rocks with insets of orthoclase, soda-rich plagioclase, and biotite in a cryptocrystalline groundmass composed of quartz and orthoclase. The calculated composition of the Willow Creek rhyolite is quartz 35, orthoclase 53, albite 5, biotite 5 to 7. Some of these lavas contain tridymite in place of quartz.

The composition of a rhyolite from the Marquesas Islands has been furnished by T. Barth; this has a special interest on account of the rarity of oversaturated rocks in the Pacific basin. The rock is largely made up of spherulites of feldspar and quartz, with 14 per cent of siliceous glass in which there are occasional scales of biotite and tiny grains of a mineral which is thought to be corundum. The norm shows quartz 24, orthoclase 25, albite 37·5, corundum 5·6, so the rock is a peraluminous soda-rhyolite.

Garnet-bearing rhyolites have been described at Chalk Mt., Colorado (W. Cross); at Bay View, Maryland (F. Bascom); and at Coburn Mt., Maine (E. S. C. Smith). The latter rock holds pink garnets up to 4 mm. in diameter in an exceedingly fine-grained groundmass of quartz, orthoclase and accessory minerals. An occurrence of garnet within vesicles in rhyolite at Ely, Nevada, has been described by A. Pabst.

Metaluminous Type

Beautiful examples of metaluminous soda-rhyolite (rhyodacite) are described by H. H. Robinson in the San Francisco district of Arizona. All degrees of crystallinity are represented among these lavas, but in most cases a cryptocrystalline to glassy base makes up about three-quarters of the rock. The insets are generally of plagioclase, with some of augite or hornblende and rarely a few of quartz. It is interesting to compare the actual mineral composition of each rock with its "potential" composition as calculated from chemical analysis; the comparison shows how misleading it is to name a hypocrystalline rock by the crystalline phases only.

		Kendrick Peak	*San Francisco Mtn.*	*Mormon Mtn.*
Actual development	plagioclase ...	18·3	13·7	18·6
	biotite	2·5	4·9	—
	hornblende			8·1
	pyroxene	1·3	1·1	—
	ore	0·6	1·7	1·9
	groundmass ...	77·3	78·6	71·3
Potential composition if fully crystallized	quartz	24·6	20·0	22·0
	orthoclase	20·6	18·9	14·5
	albite	37·2	37·3	39·0
	anorthite	7·8	11·4	10·0
	biotite	2·0	2·5	—
	hornblende ...	1·0	2·5	8·0
	hypersthene ...	1·3	1·7	0·9
	ore	3·7	3·7	1·9
	apatite	0·3	0·3	0·3
	Colour index	(8)	(11)	(11)

Among the rhyolites of Hungary there are few that contain actual crystals of hornblende or augite, but some of the cryptocrystalline varieties are potentially metaluminous, corundum being subordinate to hypersthene in the norm, as in the following instance (A. Vendl):—

quartz	29·3	corundum	0·3
orthoclase	29·5	hypersthene	3·3
albite	28·8	magnetite	0·7
anorthite	7·5	ilmenite	0·1

An obsidian from Iceland, which was minutely studied by F. E. Wright and analysed by J. B. Ferguson, has the composition of a metaluminous rhyodacite. No minerals at all have developed in the rock except a little fayalite and tridymite in shrinkage cavities, but the norm is as follows:—quartz 40, orthoclase 17, albite 27, anorthite 9, corundum 0·6, hypersthene 4·3 per cent. There is too little corundum to convert all the magnesia into biotite, so if the rock had crystallized completely it is likely that an aluminous hornblende would have been formed.

Subaluminous Type

Subaluminous rhyolites are well represented in the Lipari Islands, and have been described by H. S. Washington (2). An example from Monte Guardia is a very fine-grained, ash-gray rock with small insets of pyroxene. The groundmass is microcrystalline, and is composed of minute laths of orthoclase, prisms of pyroxene, grains of magnetite, and interstitial quartz. According to Washington's analysis the proportions of these minerals must be nearly as follows:—quartz 30, orthoclase 29, albite 34, anorthite 4, pyroxene 10, accessories 3; colour-index 13. Other rhyolites of Lipari are completely glassy, but they may be recognized as potentially subaluminous types by the appearance of diopside and hypersthene in the norm, corundum being absent.

The "toscanites" of Tuscany, Italy, are mostly subaluminous rhyolites. These lavas have insets of orthoclase and labradorite, with augite, hypersthene, and some biotite, set in a glassy groundmass. The calculated composition of a rock from Monte Amiata is:—quartz 15, orthoclase 36, albite 25, anorthite 10, pyroxene 12, magnetite 1 per cent; colour-index 13 (H. S. Washington, 3).

The rhyolites of the Yellowstone Park are mostly subaluminous. They have been minutely described by J. P. Iddings. The insets are of quartz, orthoclase, plagioclase, and pale green augite or occasionally a little ortho-pyroxene. The groundmass varies from an almost structureless glass to globulitic and spherulitic types, and to cryptocrystalline and microcrystalline. It is mainly composed of orthoclase and quartz or tridymite.

Peralkaline Type

As a general name for the peralkaline rhyolites, "pantellerite" may be used. The name was given by Förstner to lavas of Pantelleria, near Sicily, which have ægirine and ænigmatite as their characteristic dark minerals. The "comendite" of Sardinia is a similar rock in which a soda-amphibole takes the place of ægirine. These lavas show all the usual varieties of texture, but either ægirine or soda-amphibole may appear among the insets. The proportions of the essential minerals in some of the lavas of Pantelleria are stated as follows, by H. S. Washington (4):—

	(1)	(2)	(3)	(4)	(5)
quartz	13	15	20	28	30
soda-microcline	70	63	63	45	51
soda-pyroxene	9	12	14	17	7
ænigmatite	6	10	3	10	12
magnetite	2	—	—	—	—
Colour index	(17)	(22)	(17)	(27)	(19)

In nos. 4 and 5, the feldspar holds more orthoclase than albite.

Completely glassy lavas of this type may be recognized by the appearance of acmite or sodium metasilicate in the norm. For example, an obsidian from Yucatan, which was described and analyzed by H. S. Washington (5), gives the following norm:—

quartz	35	sodium metasilicate	2·7
orthoclase	27	diopside	3·1
albite	24	hypersthene	0·2
anorthite	6·5	ilmenite	0·1

Had this magma crystallized, it must have given as its chief dark mineral a pyroxene intermediate between acmite and diopside, or more likely an amphibole of the corresponding composition. (Washington has observed that rocks which contain soda-amphiboles commonly have sodium metasilicate in the norm, while those with soda-pyroxene rarely show sodium metasilicate, and then only in small quantity.)

R. A. Daly describes "quartz-trachytes" of Ascension Island which are actually soda-rhyolites, holding 23-25 per cent of normative quartz. A dark green obsidian which shows only minute needles of soda-diopside and some small insets of soda-orthoclase proves to have the following composition:—quartz 21, orthoclase 23, albite 44, soda-diopside 12 per cent.

E.—Tonalite. XOγ, XOδ

This division includes all oversaturated rocks in which the anorthite molecule exceeds the orthoclase molecule. The name tonalite was originally given to a hornblende-biotite rock occurring at Tonale, in the Southern Alps, but its significance has gradually become broadened, and we shall use it to embrace all eucrystalline rocks which satisfy the definition given above, regardless of the nature of the dark minerals. The corresponding lavas are the dacites, which take their name from the province of Dacia in Transylvania, in the loop of the Carpathian Mountains.

As a whole, the tonalites and dacites are appreciably less acid than the granites and rhyolites, although some tonalites have been described which hold over forty per cent of quartz. They are also rather richer in dark minerals; and extremely leucocratic facies, such as frequently occur among granites, are unusual in this group. Mesotype and melanocratic facies are known, but like the corresponding facies of granite, they are probably due to contamination of the magma by reaction with sediment. The dividing line between tonalite (with dacite) and diorite (with andesine) we shall draw, as before, at 10 per cent of quartz, for the reason that the majority of rocks that have been described as diorite actually hold a little quartz.

By the application of the albite : anorthite ratio, we separate the soda-tonalites and soda-dacites from the lime-tonalites and lime-dacites. The latter type, in which anorthite exceeds albite, is scantily represented, the majority of tonalites and dacites being rich in soda. Peraluminous, metaluminous and subaluminous types all occur among tonalites and dacites, but the peralkaline type is unrepresented, because soda-pyroxenes and soda-amphiboles are incompatible with much anorthite.

Peraluminous Type

Tonalite is a characteristic rock of the Alpine chain. The rocks of the Adamello Alps range from soda-granite to tonalite; so do those of the Rieserferner stock in the eastern Tyrol. The Brünner massif in Austria and the "granites" of the Carpathians include typical tonalites. The composition of the tonalite of the Brünner massif, according to F. Sellner, and of two Carpathian tonalites described by P. Radziszewsky, are as follows:—

	Brünner.	*Carpathians.*	*Carpathians.*
quartz	35	24	28
orthoclase	2·7	4·6	8·7
plagioclase	55	45	48
biotite	7.4	24	18
ore	—	0·2	—
zircon, apatite, etc.	0·3	2·4	0·4
Colour index	(8)	(26)	(18)

The peraluminous type of tonalite is also represented among the "granite-diorites" of Becker and Lindgren in Idaho; and F. L. Ransome's account of the Madera quartz-diorite, in the Globe district of Arizona, shows that this, too, is a true tonalite.

	Idaho.	*Globe.*
quartz	25·0	26·9
orthoclase	11·2	6·4
albite	30·2	23·2
anorthite	13·9	20·4
biotite	16·0	17·7
muscovite	—	1·0
ore	0·6	2·8
sphene, apatite, etc.	2·1	1·4
Colour index	(19)	(23)

Metaluminous Type

This is the commonest type of tonalite, just as it is the commonest type of soda-granite. It is well illustrated by the Marysville batholith, Montana, described by J. Barrell. Beside the latter we place two analyses of tonalites from the Haystack stock, in the same state; these are due to W. H. Emmons.

	Marysville.	*Haystack.*	*Haystack.*
quartz	22·2	21·8	12·1
orthoclase	15·6	11·7	9·7
albite	26·3	30·1	31·2
anorthite	21·2	15·1	20·8
biotite	7·2	4·8	6·7
hornblende	5·5	—	—
pyroxene	—	11·2	13·5
ore	1·7	4·9	4·9
apatite, etc.	0·3	0·4	1·1
Colour index	(15)	(21)	(26)

R. A. Daly describes great intrusions of tonalite in British Columbia. The examples quoted below refer to the Osoyoos batholith (1), the Remmel batholith (2), the Chilliwack batholith (3), and the Trail batholith (4).

	(1)	(2)	(3)	(4)
quartz	37·0	27·0	19·1	25·9
orthoclase	7·5	7·2	2·0	19·2
plagioclase (Ab>An)	33·1	50·7	55·7	28·1
biotite	10·8	5·7	10·3	13·4
hornblende	3·1	4·3	11·1	12·3
epidote	8·0	—	—	—
sphene	0·3	0·6	—	0·2
ore	—	3·8	1·3	0·6
apatite, zircon	0·2	0·7	0·5	0·3
Colour index	(22)	(15)	(23)	(27)

Melanocratic types containing 40 to 50 per cent of hornblende occur in the Purcell Mts. (Daly).

A quantitative study of the average mineral composition of the rocks of the Coast Range in South-Eastern Alaska has been made by F. E. and C. W. Wright, with the following result :—

quartz	19·4	apatite	0·6
orthoclase	6·6	magnetite	0·9
andesine ($Ab_{56}An_{44}$)	47·4	pyrite	0·1
hornblende	7·6	sphene	1·3
biotite	11·6	decomposition products	4·5

It is clear from these figures that the average rock of the Coast Range is considerably more calcic than granite, and the bulk of it probably belongs to tonalite.

E. S. Larsen has described the rocks of the batholith of Southern and Lower California, which is composed to the extent of 50 per cent. of tonalite. The modes of two of the most abundant varieties of this tonalite are given below (weight per cent).

	Green Valley	*Bonsall*
quartz	12 to 30	18
potash feldspar	1 — 20	—
plagioclase	40 — 60	50
biotite	3 — 10	18
hornblende	3 — 15	13
rest	1	1
Colour index	(6)—(26)	(32)

Subaluminous Type

Few examples of subaluminous tonalite have been traced in petrographical literature; no doubt they are rare, just as subaluminous granites are. On the other hand, subaluminous dacites may be expected to occur very frequently, but it is hard to detect them in the literature, since they have generally been called "andesite" and very few andesites have been described in quantitative mineralogical terms.

The "hypersthene-quartz-diorite" of the Shevaroy Hills in Madras, one of the intermediate members of T. H. Holland's "charnockite series" is a subaluminous tonalite. The mode of this rock has been determined by H. S. Washington (1): it is quartz 15·5, feldspar (mainly andesine) 55·5, hypersthene 23·6, ore 5·4 per cent. (Colour index 29.)

The rock named "enderbite" by C. E. Tilley (2) from Enderby Land, Antarctica, falls in this group with the following mode:—quartz 42·5, antiperthite 53, hypersthene 3, magnetite 1.

F.—DACITE. DOγ, DOδ

No description of a *peraluminous* dacite has been traced in the literature. Many rocks have been described under the name of mica-dacite, but they all fall into rhyolite in our system.

Metaluminous Type

Dacites of the metaluminous type have been described in considerable number in eastern Australia. H. I. Jensen and H. C. Richards have described such rocks in the Glass House Mountains and elsewhere in Queensland. The Glass House dacite is microcrystalline, with insets of plagioclase and quartz in a groundmass which contains augite, a little hornblende, and magnetite. Jensen's analysis indicates the following composition:—quartz 17, orthoclase 17, albite 30, anorthite 20, melanes 15. Examples of hypersthene-biotite dacite have been studied by A. B. Edwards in Victoria. Quartz, plagioclase, biotite and hypersthene all appear among the insets, and the groundmass is chiefly quartz and orthoclase.

A hornblende-dacite was described by J. S. Diller, at Mt. Shasta, California. It is a light gray rock with a few insets of hornblende. The groundmass is sometimes holocrystalline, but more often it holds some brown glass; it is composed of plagioclase with some hornblende and some little grains of hypersthene and magnetite. The mode is not stated quantitatively, but the norm indicates the following composition:—quartz 20, orthoclase 7, albite 37, anorthite 25, melanes 10.

Subaluminous Type

The "andesite" of Hakone volcano, Japan, is siliceous enough to be called dacite. The following mode is one of those supplied by H. Kuno.

Insets	plagioclase (An_{81})	13·5	weight %
	augite and hypersthene	2·4	
	magnetite	1·4	
Groundmass	quartz	11·6	
	tridymite	6·3	
	anorthoclase	16·5	
	plagioclase (An_{58})	31·2	
	pigeonite	12·3	
	accessory	3·8	
	Colour index	(20)	

Dacites accompanied the eruption of Souffrière, St. Vincent, in 1902. According to J. S. Flett, these lavas have insets of zoned plagioclase, augite and hypersthene, in a groundmass of feldspar, quartz, tridymite, hypersthene and glass. Analyses of the lavas and their ashes show from 10 to 15 per cent of quartz, with anorthite in large excess over orthoclase.

J. S. Diller has described as "hypersthene andesite" a lava on the west slope of Mount Shasta, California, which appears to be a lime-dacite. It is a light gray, compact rock, with a few insets of hypersthene. The mineralogical composition is not stated quantitatively, but the norm shows 16 per cent of quartz, 8 of orthoclase, 26 of albite, 39 of anorthite, and 11 of dark minerals.

Some remarkable rocks of basaltic appearance, from the Firth of Clyde, Scotland, have been described by G. W.

Tyrrell under the name of "cumbraite." One of these rocks exhibited 58 per cent of crystals in a glassy base, this proportion being made up of anorthite 7, labradorite 31, enstatite and augite 20 per cent. The composition of the glass that forms the remaining 42 per cent of the rock is such that if it had crystallized it must have formed quartz (about 35 per cent), orthoclase (about 35 per cent), and albite (about 30 per cent). Combining these figures with those given first, it appears that if the cumbraite had crystallized completely it would have had nearly the following composition:—quartz 14·7, orthoclase 14·7, albite 28·8, anorthite 21·8, pyroxene 20, with a trace of magnetite. The rock consequently falls into soda-dacite.

Another rock whose appearance belies its composition is the "weiselbergite" of Oberkirchen, Germany. This is a more than half glassy rock with a few insets of plagioclase and some microscopic prisms of augite and bronzite. Rosenbusch described it as augite-porphyrite; Zirkel considered it a glassy type of diabase. The potential composition has been worked out by E. Tröger, who finds it to be quartz 20, orthoclase 6, oligoclase 54, augite and bronzite 16, accessory 4; thus the rock is a soda-dacite.

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CHAPTER XXI

THE SATURATED (AND SLIGHTLY OVER-SATURATED) ROCKS.
(DESCRIPTION)

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J.—Perknite. XSH p. 417

A.—The Mineralogy of the Saturated Rocks

The minerals of the saturated rocks are the same that occur in the oversaturated class, with the exception that quartz and tridymite are typically absent. Along with the excess of silica, muscovite, lithia-mica, tourmaline and topaz virtually disappear too. Alkali-feldspars are no longer dominant, and pure potash-feldspar is quite scarce, anorthoclase and microperthite taking its place. But the plagioclase series occurs in its entire range from practically pure albite to practically pure anorthite. Some analyses of rocks containing plagioclase show a deficiency of silica, or an excess of sodium aluminate, which suggests the existence of the molecule $Na_2Al_2Si_2O_8$ or soda-anorthite as a constituent of plagioclase; but this assumption still awaits proof (see Chapter II, p. 17).

The dark minerals of the saturated rocks include biotite and all the chief varieties of amphibole and pyroxene. Among the amphiboles the green ferrous hornblende is commonest in the more siliceous rocks; the brown ferric variety, into which the green hornblende transforms at high temperature, is found especially in rocks of low silica content. The soda-amphiboles are confined to the syenites and trachytes and are remarkable for their very late crystallization. Although the hornblende in the more siliceous rocks is apparently primary, that in many of the more melanic rocks has been formed by the transformation of earlier pyroxene crystals. The so-called "uralite" is a pseudomorphic amphibole which retains the external form of the pyroxene crystals it has replaced. Sometimes the pyroxene crystals have mantles of hornblende around them; in other cases relics of pyroxene form the cores of large hornblende crystals; and it sometimes happens that the ophitic texture or the diallagic structure of augite has been preserved, although the material has gone over entirely into amphibole.

Pyroxene occurs in many varieties. Ortho-pyroxene (enstatite to hypersthene) has a great development in the norites and pyroxenites, and clino-pyroxene (augite or diallage) in the gabbros, diabases and basalts. The intermediate type, enstatite-augite or pigeonite, has perhaps a wider distribution than any of them; it appears to be the commonest pyroxene in andesites, basalts and diabases. The lime-iron pyroxenes, diopside and hedenbergite, occur in syenites and trachytes; so do soda-diopside and ægirine.

Ænigmatite and fayalite are scarce constituents of the more siliceous lavas, and a large number of uncommon species containing rare earths have been found in syenites and syenite-pegmatites. Sphene is particularly common in rocks containing hornblende. Of minerals other than silicates, magnetite and ilmenite are the most important, and apatite is a common accessory constituent. Sulphides of iron, copper and nickel, such as bornite, chalcopyrite, pentlandite and pyrrhotite, are sometimes present in the more melanic rocks, and in some cases they must be regarded as original constituents.

B.—CLASSIFICATION AND NOMENCLATURE OF THE SATURATED ROCKS.

The subdivision of this class is complicated by the abundance of melanic and even hypermelanic rocks, many of which have received distinctive names. The colour index must therefore find a place in the system.

We begin, as before, by separating the rocks in which Or>An from those with An>Or. There is no difficulty about naming the orthoclase division; it contains the syenites and trachytes, and we subdivide them according to the Or : Ab ratio into potash and soda varieties. The shonkinites or melanic syenites form a parallel series with colour index greater than 30. There are no lavas of this composition.

It is customary to recognize an intermediate group of rocks between syenite and diorite, to which the name *monzonite* is given. F. Becke states that the ratio of Or : Ab : An in typical monzonite is about 35 : 45 : 20, and that the dark and light constituents are approximately equal in quantity. The typical monzonite is therefore a mesotype rock, but W. C. Brögger (1) described leucrocratic rocks of similar mineralogical character in Norway, giving them the name *akerite*. Lastly, there is a group of syenites containing feldspar that is rich in soda and virtually free from anorthite. These may be called albite-syenite, but the specific name *umptekite* is commonly used for such rocks.

Saturated rocks in which An>Or have many names, especially diorite, anorthosite, gabbro, norite, gabbrodiorite and diabase, with andesite and basalt for the extrusive forms. All anorthosite and most diorite and andesite are leucocratic, but the vast majority of gabbros, norites, diabases and basalts have a colour index between 35 and 60. Here is a suitable basis for subdivision, and the Ab : An ratio provides another. The four groups of rocks so defined may be called soda-diorite and lime-diorite, soda-gabbro and lime-gabbro. These are group names, not specific names.

Finally, all rocks in which the proportion of feldspar falls below 10 per cent are brought together under the group name of perknite.

The complete scheme appears below.

EUCRYSTALLINE ROCKS

		Leucocratic	*Mesotype-Melanocratic*	*Hypermelanic*
Or>An	Or>Ab	Potash-syenite	Shonkinite	
	Ab>Or	Soda-syenite (a) umptekite (b) akerite	Monzonite	
An>Or	Ab>An	Soda-diorite	Soda-gabbro	
	An>Ab	Lime-diorite	Lime-gabbro	
				Perknite

DYSCRYSTALLINE ROCKS

		Leucocratic	*Mesotype-Melanocratic*	*Hypermelanic*
Or>An	Or>Ab	Potash-trachyte	(None known)	
	Ab>Or	Soda-trachyte (a) albite-trachyte (b) trachyandesite	(None known)	
An>Or	Ab>An	Soda-andesite	Soda-basalt	(None known)
	An>Ab	Lime-andesite	Lime-basalt	

In the absence of a mineralogical analysis of a saturated rock, the norm calculated from the chemical analysis is not always a reliable guide to the Or-Ab-An ratio, for a certain quantity of lime and alumina that belongs to other minerals is diverted to anorthite in the calculation; hence the normative anorthite may largely exceed the amount actually present in the rock. Thus a shonkinite from Celebes, which was described by Iddings and Morley, is composed of augite and biotite with subordinate orthoclase, yet the norm shows 13 per cent of anorthite.

All four reaction-types are represented among the syenites and trachytes, but the peralkaline type does not appear in the plagioclase division. With increasing lime content the peraluminous type becomes uncommon.

C.—SYENITE. X̄SaL, XSβI.

Peraluminous Type

The syenites of Durbach and Meissen, Germany, are always quoted as examples of mica-syenite. The Durbach syenite is a marginal facies of the Black Forest granite; in part it is a pure mica-syenite, but the biotite is very often accompanied by hornblende. Quartz is generally present, and no doubt a good deal of rock that would be better described as subacid granite has been included under the name of syenite. The biotite-syenite of Meissen, Saxony, is also quartz-bearing, and it is doubtful whether there is any true mica syenite in the area at all.

Syenitic facies occur among the granites of Ontario.

Adams and Barlow describe a soda-syenite that consists of red microperthite with three or four per cent. of biotite, and a little muscovite, calcite, and apatite. Another soda-syenite, from Craigmont, has about 85 per cent of feldspar, 13 of titanomagnetite, and a little biotite. A red syenite at Monmouth has the following composition :—quartz 4·2, microperthite 84·5, muscovite 4·8, biotite 3·9, rest 2·3; colour index 9.

Potash-syenites are much less common than soda-syenites. A. Lacroix (2) describes a highly potassic syenite-aplite in Madagascar, consisting of microcline and albite, with muscovite, biotite, and a very little quartz. The calculated norm shows 63 per cent. of Or, 18 of Ab, and 6 of An.

Metaluminous Type

The hornblende syenite of the Plauenscher Grund. Dresden, has long been regarded as the typical example of this kind of rock; yet a good deal of the rock that passes for syenite, and is found in rock collections all over the world, holds 10 per cent or more of quartz. H. Ebert has assembled 17 determinations, made by nine different investigators, of the composition of this rock. Quartz is present in every case, ranging from 5 to 12 per cent. Potash-feldspar predominates in general, but plagioclase sometimes exceeds it in amount. Hornblende is the only dark silicate and varies between 10 and 20 per cent.

The "nordmarkite" of the Christiania district of Norway is a very leucocratic syenite passing into subacid granite. The dark minerals are generally brown biotite and green hornblende, and a little acid plagioclase is present; but peralkaline rocks have been described under the same name.

Metaluminous orthosyenites have been described at various points in Quebec and Ontario. The syenite of Brome Mountain, Quebec, described by J. A. Dresser (1) holds 90 per cent of anorthoclase along with 10 per cent of biotite, pyroxene, hornblende, sphene and apatite. The same geologist (2) has described the syenite of Shefford Mountain, Quebec, which consists of microperthite with very little pale green augite, green hornblende and brown biotite, as well as traces of sphene and occasionally a few grains of quartz.

Beautiful examples of soda-syenite have been described by L. V. Pirsson (1) in Montana. The syenite of Highwood Peak has pale green diopside and brown biotite as its chief dark minerals, but some green hornblende has been formed by transformation of diopside, and there is a good deal of white mica present which is a bleached form of biotite. The syenite of Yogo Peak is very similar to the last, and both are similar to the syenites of Tripyramid Mountain and Belknap Mountain in New Hampshire, which have been described by Pirsson and Washington.

	Highwood Peak.	*Yogo Peak.*	*Tripyramid.*	*Belknap.*
quartz	8·2	6·1	1·0	4·2
soda-potash feldspar	71·2	72·5	86·5	70·5
biotite	12·0	—	1·0	7·6
hornblende	—	12·9	6·0	13·4
diopside	7·7	5·4	1·0	—
accessories	1·1	3·1	4·5	4·3
Colour index	(21)	(21)	(12)	(25)

All the above rocks belong to the Umptek type. The more calcic soda-syenites of the Aker type take their name from a locality in Norway, but the published descriptions of Norwegian akerites are inadequate. A typical akerite (leucocratic monzonite) occurs in the Leeuwfontein stock, Transvaal. The feldspar which makes up about 80 per cent of the rock has cores of oligoclase-andesine surrounded by mantles of anorthoclase: and the melanes are brown hornblende, diopside, biotite and accessories. (S. J. S., 1.)

Very similar rocks occur in the Monteregian Hills of Quebec. The akerite of Shefford Mountain, described by J. A. Dresser (2), carries plagioclase, soda-orthoclase, pyroxene, biotite, and brown hornblende. The akerite of Mount Yamaska is described as follows by G. A. Young :—

quartz	7·9	diopside	0·8
orthoclase	16·7	ore	6·0
albite	43·0	apatite	2·2
anorthite	9·2	accessories	1·4
biotite	11·8	Colour index	(21)

The same type occurs in Tripyramid Mountain, New Hampshire (L. V. Pirsson, 2) and at Cuttingsville, Vermont

(J. W. Eggleston). Each of these rocks carries about 60 per cent of feldspar, consisting partly of soda-orthoclase and partly of oligoclase or andesine, together with brown hornblende and either diopside or biotite, besides the usual accessory minerals.

Subaluminous Type

Subaluminous syenites seem to be distinctly uncommon. for although diopside is one of the commonest constituents of syenites, it is either accompanied by hornblende or mica, which take the rock into the metaluminous class, or else it holds acmite in solid solution, and the rock is then peralkaline.

H. von Eckermann gives the mode of a Swedish syenite-porphyry as follows:—quartz 7, orthoclase and albite 82, diopside 5, sphene 2, ore 2·5; colour index (9). J. Phemister has given the name of "perthosite" to certain syenites in the north-west of Scotland, which consists almost entirely of perthitic feldspar, with only 1 to 3 per cent of dark minerals. In a typical case the dark minerals were diopside 1·75, magnetite 0·5, sphene 0·25 per cent, leaving 97·5 per cent of feldspar.

Peralkaline Type

This type is widely represented, for the pyroxene in most examples of syenite tends to hold a little of the acmite molecole or to be mantled by soda-diopside or soda-hornblende. The feldspar is always a purely alkaline variety, either soda-orthoclase, soda-microcline, albite, or some variety of microperthite.

The syenite of Umptek, Kola, which has given us the name umptekite, is a very leucocratic rock composed of microperthite, arfvedsonite and ægirine; it constitutes the border facies of the great foyaite intrusion at Umptek. A very similar rock, which is also associated with foyaite, has been described by P. D. Quensel at Almunge, Sweden; this rock contains two kinds of soda-amphibole, a brown arfvedsonite and a blue-green hastingsite. The Norwegian rocks which W. C. Brögger has called "lestiwarite" and "sölvsbergite" are peralkaline dyke-syenites.

The peralkaline granites of Massachusetts pass locally into syenite. The Beverley syenite is virtually free from quartz; the dark minerals are sometimes ægirine and sometimes soda-hornblende, with accessory biotite, magnetite and sphene (C. H. Clapp). The hornblende crystals sometimes exhibit cores of diopside. An analysis of the Beverley syenite by F. E. Wright shows that the rock must consist of about 38 per cent of orthoclase and the same quantity of albite, with some 20 per cent of metasilicates and a little quartz and ore. The rock of Wigwam Quarry, according to Clapp, has 70 per cent of alkali-feldspar, 8 per cent of quartz, and 22 per cent of riebeckite and ægirine.

Beautiful examples of peralkaline syenite are developed between granite and an enclosed mass of limestone at Palabora, Transvaal, as described on page 81. The characteristic melanes are a jade-green, more or less sodic pyroxene and a lavender-blue amphibole of the arfvedsonite type, or sometimes a deep blue riebeckite. With increasing colour index these syenites pass into shonkinite and pyroxene (S. J. S., 2).

Syenites with blue amphiboles and biotite have been described by A. Lacroix (2) in Madagascar. Among them there is a potash syenite or pegmatite consisting solely of microcline and an unusual variety of soda-amphibole (torendrikite); this rock is exceptionally rich in potash, containing 11·2 per cent of K_2O to 2·8 per cent of Na_2O.

D.—TRACHYTE. DSaL, DSβL

Peraluminous Type

Many rocks have been described under the name of mica-trachyte, but most of them are either so quartzose that they are properly rhyolites, or else they carry metasilicates in addition to biotite and therefore belong to the metaluminous type. The trachyte of Game Ridge, in the Rosita Hills of Colorado, seems to be a true peraluminous trachyte. According to W. Cross (2), it shows insets of sanidine, plagioclase and biotite in a groundmass of orthoclase and oligoclase, with a little quartz.

Metaluminous Type

The majority of trachytes are metaluminous. The widely known rocks of the Drachenfels and other hills in the Siebengebirge of the Rhine are porphyritic rocks with large insets of soda-orthoclase in a pale gray, microcrystalline to hypocrystalline base. Some plagioclase is always present, and among the scanty dark minerals biotite and hornblende are commonest, though diopside is sometimes present too. The hornblende crystals often have a corroded appearance, and are changed externally into grains of pyroxene and ore; this is just the reverse of the transformation usually observed in syenites. The glass base is highly siliceous, and a little quartz is sometimes recognizable in the groundmass. The norm of the Scheerkopf trachyte is quartz 6·4, orthoclase 47·8, albite 28·8, anorthite 7·8, leaving about 10 per cent of metasilicates, mica and accessories.

The trachytes of Monte Gibele, Pantelleria, are porphyritic rocks holding soda-microcline, colourless pyroxene (but sometimes sodic pyroxene, and then passing over to the peralkaline type), and a little brown hornblende. A trace of quartz may be present in the interstices. H. S. Washington (1) gives the composition of one of these lavas as follows :—

quartz	3	hornblende	5
soda-microcline	83	magnetite	2
pyroxene	7	Colour index	(14)

Metaluminous trachytes from Bolsena and Vetralla, Italy, are also described by Washington (2), who states that in these rocks "both olivine and quartz are wanting or only present sporadically." The following modes are given :—

	(1)	(2)	(3)
orthoclase	49·5	44·3	48
albite	21·7	27·0	27.
anorthite	14·4	14·1	12
diopside	6·4	3·2	7·4
biotite	3·4	4·8	1·0
accessory	4·6	6·6	4·1
Colour index	(14)	(14)	(12)

A perfect example of an almost precisely saturated soda-trachyte of the plagioclase-bearing variety (trachyandesite) is found in the Leeuwfontein complex, Pretoria; it is the effusive form of the magma which furnished the akerite described above (page 394). It is a dark gray rock of andesitic appearance, with some insets of plagioclase. Under the microscope each plagioclase crystal is seen to be zonally built, with a core of oligoclase-andesine and a broad mantle of anorthoclase in which tiny grains of dark minerals are embedded. The groundmass is composed of unstriated feldspar and greenish hornblende, gray diopside, and green to brown biotite, with titanomagnetite and apatite. The dark minerals are often clustered together, a grain of diopside being surrounded by hornblende, biotite and ore grains which seem to have resulted from transformation of the pyroxene. The composition of the rock is approximately as follows (S. J. S., 1):—

anorthoclase (about $Ab_7An\,Or_2$)	65 to 70% by volume.
plagioclase (about Ab_2An)	12 to 15
hornblende	12
diopside	2·5
biotite	2
titanomagnetite	2·5
apatite	1
Colour index	(20)

H. S. Washington (3) describes as "latite, Monfina type" a trachyandesite from Italy. This is a porphyritic rock with insets of biotite and augite, and some of feldspar, in a microcrystalline to cryptocrystalline groundmass. The mode is

quartz	6·8	biotite	7·3
alkali-feldspar (Or_2Ab)	33·3	ores	6·0
plagioclase (Ab_2An_2)	34·1	apatite	0·4
augite	12·1	Colour index	(26)

The rock of Mount Pennell, in the Henry Mountains of Utah, is a trachyandesite according to the data furnished by Whitman Cross (1, 2), and so are some of the "andesites" of the Rosita Hills, Colorado. The dark minerals are augite, biotite and hornblende, and the feldspars have cores of plagioclase with mantles of soda-orthoclase. In the mantles, tiny grains of dark minerals are enclosed, just as in the rock of Leeuwfontein, Transvaal.

Subaluminous Type

Subaluminous trachytes ought not to be too uncommon, but well-described examples of this type are not numerous.

Good examples have been found by H. S. Washington (5) on the island of Sardinia. The Cuglieri type is porphyritic, with large feldspar crystals and a few black augites in a light gray, microcrystalline groundmass. The feldspar is chiefly soda-orthoclase, but some acid plagioclase appears too. The mode is as follows:—quartz 2, soda-orthoclase (Or_7Ab_3) 62, andesine (Ab_3An_1) 26, augite 5, ores 5; colour index 10.

The Sennariolo type is dense and very fine grained, almost without insets of any kind, and is composed of very small laths of alkali-feldspar and a few grains of pale gray augite and magnetite, in a colourless, glassy base. The mode in this case can only be calculated from the chemical analysis: it is approximately quartz 2·5, soda-orthoclase 73, andesine 16, augite 3·5, ores 5; colour index 8.

The trachyte of Monte Arci has insets of soda-microcline in a dense gray base which is composed of feldspar and quartz with a few grains of augite and magnetite. The mode is quartz 8, soda-microcline 83, pyroxene 6, ores 3; colour index 9.

The trachytes of the oceanic islands possess a special interest because they seem to have been generated directly from an olivine-basalt magma. T. F. W. Barth (1) gives the following mode for a trachyte of the Marquesas islands: quartz 2, alkali-feldspar 65, plagioclase 25, pyroxene 2, ore 4; colour index 6.

Peralkaline Type

Peralkaline trachytes are well developed in eastern Australia and New Zealand, where they have been described by H. I. Jensen, H. C. Richards and P. Marshall. The trachytes of the Nandewar Mountains carry both ægirine and a deep blue hornblende, with about 90 per cent of feldspar (sanidine and anorthoclase) and sometimes a little interstitial quartz. Ægirine trachytes occur on Mount Flinders; Jensen gives the composition of one of them thus: anorthoclase 75 per cent, ægirine 20, magnetite 5. A similar rock in the Canobolas Mountains has 80 per cent of feldspar, 15 of soda-diopside, and 5 of accessories. Arfvedsonite and brown hornblende are the principal dark minerals in trachytes of the Warrumbungle Mountains; and arfvedsonite and riebeckite in those of the Glass House

Mountains. One of the latter holds 85 per cent of feldspar, 10 to 14 of riebeckite and other soda-amphiboles, a little ægirine, and accessory magnetite and zircon.

Similar rocks are found in New Zealand. P. Marshall describes trachytes with sanidine or anorthoclase and a little soda-diopside and magnetite, as well as one or two per cent of quartz, near Dunedin.

The widespread trachytes of Kenya have been described by J. W. Gregory and A. Neilson, and in greater petrographic detail by W. Campbell Smith (1). The earlier writers named these rocks "phonolitic trachyte" and even "phonolitic quartz-trachyte," using the adjective phonolitic in the sense of peralkaline. But the rocks are actually nearer to rhyolite than to phonolite, for all the specimens that have been studied quantitatively contain from 3 to 11 per cent of quartz. The melanes include ægirine, katophorite, cossyrite and fayalite.

Numerous occurrences of peralkaline trachyte have been recognized on oceanic islands. The trachytes of Ascension have been described by several writers, the latest and most complete account being that by R. A. Daly (2). The rocks contain about 80 per cent of soda-orthoclase or anorthoclase, and ægirine is the commonest melane, although riebeckite, cossyrite, and brown soda-amphibole may appear in addition. The norms show from 3 to 11 per cent of quartz. Entirely similar lavas have been described on Gough Island, in the South Atlantic, by W. Campbell Smith (2), who notes that many of them carry an iron-rich olivine in some abundance; and by T. F. W. Barth (1) on Maui and Hualalai, in the Hawaiian group.

The trachyte of Chamarel, Mauritius, contains about 90 per cent of anorthoclase, with ægirine and cossyrite as the characteristic melanes (S. J. S., 3). A chemical analysis of this rock shows silica to be so deficient that some nepheline appears in the norm although none could be detected in the rock. It is supposed that there is some of the hypothetical "soda-anorthite" in the feldspar. Such a rock may properly be described as a "phonolitic trachyte," for it is almost a phonolite.

E.—SHONKINITE AND MONZONITE. $XS\alpha M$, $XS\beta M$

The name *shonkinite* was given by Weed and Pirsson to a melanic syenite that occurs at Shonkin Sag in the Highwood Mts. of Montana. It was defined as "a granular plutonic rock consisting of essential augite and orthoclase," but the accompanying description shows that the rock is seldom so simple as that. The kernel of the definition is just that shonkinite is a mesotype to melanocratic syenite, and we shall use the name in that sense. The feldspar is always a potassic one. *Monzonite* is a mesotype to melanocratic soda-syenite containing a plagioclase with an appreciable content (more than 1/10) of anorthite. The lavas that are associated with monzonite in Montana, British Columbia, Sierra Nevada and elsewhere are all leucocratic (trachyandesite). No example of a mesotype lava of monzonitic composition is known.

No *peraluminous* shonkinite or monzonite seems to have been described. Many of the dyke rocks which pass under the name "minette" are mica-shonkinites.

Metaluminous Type

The shonkinites of Square Butte and Shonkin Sag are for the most part metaluminous, biotite being present in considerable amount. From a number of modes furnished by J. D. Barksdale it appears that pyroxene ranges from 27 to 52 per cent and biotite from 11 to 21. Much of the rock is olivine-bearing and it is not clear how much of it is strictly saturated. Here we have an illustration of the difficulty that arises when a name is attached to a particular rock-mass rather than to a particular association of minerals. The rock of Shonkin Sag includes at least three species, namely, shonkinite, olivine-shonkinite, and nepheline-shonkinite.

The shonkinite of Celebes, which has a special interest because of its association with leucitic lavas, contains nearly equal amounts of feldspar and melanes; the feldspar is almost entirely orthoclase and the melanes are augite and biotite, with accessory ore and apatite (Iddings and Morley).

The rocks that F. F. Grout (1, 2) described under the name shonkinite, from the Vermilion and Saganaga granite batholiths in Minnesota, would be better described as monzonites since they contain a large proportion of oligoclase. The modes of some of these rocks are given below.

	(1)	(2)	(3)
quartz	5	5	3
orthoclase	20	15	10
oligoclase	40	25	20
biotite	3	—	—
hornblende	30	55	65
accessory	2	—	2
Colour index	(35)	(55)	(67)

W. C. Brögger (2) gives the following statement of the mineralogical composition of the typical monzonite of Tyrol:—

quartz	2·5	pyroxene	14·5
soda-orthoclase	30·0	magnetite	3·0
plagioclase (Ab_3An_2)	32·0	apatite	2·0
biotite	10·0	sphene, zircon	1·0
hornblende	5·0	Colour index	(35)

L. V. Pirsson has described monzonites in the Belknap Mountains of New Hampshire (1); and at Yogo Peak (2), in association with the syenites mentioned on page 394.

	(1)	(2)	
soda-orthoclase	20·0	30·4	(Or_1Ab_1)
plagioclase (andesine)	27·2	27·2	(Ab_3An_2)
biotite	1·0	12·1	
hornblende	35·3	4·5	
diopside	—	20·7	
accessories, ore	16·5	5·1	
Colour index	(53)	(42)	

D. L. Reynolds (1) has given a very precise and detailed description of an augite-monzonite that forms part of the Newry igneous complex in northern Ireland. Augite and biotite in nearly equal parts make up almost half of this rock, the remainder being soda-orthoclase, andesine, and a little quartz. There are several mineralogical and textural features which indicate that this rock is a hybrid, not a simple magmatic product.

Subaluminous Type

Few examples of subaluminous shonkinite or monzonite have been noted. In the following instance, described by

C. W. Drysdale, the proportion of hornblende and biotite is so small that the rock may be considered to be subaluminous. An augite syenite at Franklin, British Columbia, has a coarse-grained marginal facies composed of large augite crystals, rising to an inch in diameter, set in a groundmass of smaller augite crystals with interstitial orthoclase. The mode is

orthoclase	17·1	ore	6·0
biotite	0·8	apatite	1·0
hornblende	1·5	sphene	0·4
augite	73·1	Colour index	(83)

The shonkinite of Palabora, Transvaal (see pp. 81-83) is a rock of very simple composition; it corresponds more closely to Weed and Pirsson's definition than the rock of Shonkin Sag does. Idiomorphic prisms of green diopside make up sixty per cent of the rock and are enclosed in large poikilitic plates of microcline. Two or three per cent of apatite, and mere traces of sphene and biotite, are the only other constituents. It is not considered that this rock crystallized from a magma of its own composition; it is a product of reaction between granitic magma and limestone. With increasing diopside content this rock passes into pyroxenite and pyroxene-apatite rocks (S. J. S., 2).

Peralkaline Type

This is distinguished from the last type by the entrance of more or less of the acmite molecule into the pyroxene. The shonkinite of the Highwood Mountains sometimes shows a transition towards this type, mantles of bright green acmite-diopside forming round the ordinary pale green pyroxene. Melanocratic syenites containing 40 to 70 per cent of sodic pyroxene has been described at Tveitasen and Melteig, in Norway (W. C. Brögger, 4), and at Ahvenvaara, Finland (V. Hackmann).

F.—Diorite and Anorthosite. xSγL, xSδL

Diorite is used here as a group-name for all saturated eucrystalline rocks that are rich in plagioclase. The group is divided by the Ab : An ratio into *soda-diorite*, which is diorite in the sense of Rosenbusch and includes the more sodic anorthosites, and *lime-diorite,* which takes in the rest of the anorthosites and other leucocratic rocks with a calcic plagioclase. One might use anorthosite as a group name in place of lime-diorite, but this would destroy the usefulness of the name anorthosite for "rocks composed almost entirely of feldspar." On the other hand it may be pointed out that this name has not infrequently been applied to rocks containing more than 20 per cent of dark minerals. But it is the group that matters, not the name we give it, and the group is concisely defined by the symbol xsδ L.

Peraluminous Type

Diorites of peraluminous type are little known. Many rocks are described loosely as "mica-diorite," but it generally appears on investigation that such rocks either hold far too much quartz to be diorites at all, or else they carry hornblende as well as biotite. Among dyke-rocks, the "kersantite" group is characterized by the association of plagioclase with biotite, but when quantitative data are available these rocks generally show more than 30 per cent of dark minerals.

Metaluminous Type

Diorite of this type is not uncommon, but well described examples are few. The Pinto diorite, in the Little Belt Mts. of Montana, has the following composition according to L. V. Pirsson (3):—orthoclase 17·5, andesine 63·5, biotite 3·5, hornblende 13, ore 2·5; colour index 19.

J. E. Wolff gives the following modes for the diorite of the Big Timber stock, in the Crazy Mts., Montana:—

	(1)	(2)	(3)	(4)
quartz	3	—	4	6
orthoclase	2	—	1	10
labradorite	70	68	68	55
biotite	14	4	12	10
hornblende	—	15	5	—
augite	5	6	7	13
hypersthene	—	—	1	4
accessory	5	2	1	1
Colour index	(24)	(27)	(26)	(28)

A soda-diorite described by E. H. Watson in the San Carlos Mts., Mexico, is interesting because it is suggested that it represents a residuum squeezed out of a body of gabbro in process of crystallization, and that the majority of the constituents were drained from the gabbro largely in the form of a crystal mush. The mode is quartz 5, orthoclase 23, andesine 60, hornblende 8, augite 2, biotite 1, magnetite 1; colour index 12.

Some facies of the " norite " of Sudbury, Ontario, are so leucocratic that they fall into the present division. The following modes are given by W. A. Zisman.

	(1)	(2)	(3)
quartz	3	8	6
potash-feldspar	1	1	4
andesine-labradorite	74	67	70
biotite	1·5	6	3
hornblende	2	5	3
augite	3	5	2·5
hypersthene	15	8	10
accessory	·5	—	1·5
Colour index	(22)	(24)	(20)

A. F. Buddington gives the following examples of diorite from New York State:—

	(1)	(2)	(3)
quartz	3	3	9
orthoclase	—	2	3
andesine	66·5	74	64
biotite	7	4	12
hornblende	21	10	11
pyroxene	—	4	—
accessory	2·5	3	1
Colour index	(30)	(21)	(24)

A soda-diorite is described by Friedlaender and Niggli in the Vosges Mts., containing quartz 5, orthoclase 9, andesine 69, hornblende, biotite and chlorite 14, accessories 3 per cent; colour index 17. In the same region there are numerous dykes of vogesite, which may be called dyke-diorite, of similar mineralogical character.

The ornöite of Stockholm, Sweden, is a perfect example of this type. E. Tröger gives the following mode: microcline 5, oligoclase-andesine 80, hornblende 12, accessory 3; colour index 15.

Many anorthosites are metaluminous. The Marcy anorthosite of the Adirondack Mts. is composed of large blue-gray labradorite crystals, averaging about an inch in diameter, together with 5 to 10 per cent of hornblende, pyroxene, biotite, garnet, ilmenite, and sometimes iron and copper sulphides. More sodic varieties occur too, in which both orthoclase and quartz are present (W. J. Miller).

Anorthosite of San Carlos, California, containing 95 per cent of plagioclase of the composition An_{88} with 4 per cent of hornblende and 1 per cent of augite, has been described by F. S. Miller.

Subaluminous Type

The "andesine anorthosite" of Amherst County, Virginia, described by Watson and Cline, is a subaluminous soda-diorite. It is composed of andesine and some microcline, with subordinate blue quartz and a little hypersthene, rutile, ilmenite and apatite. The norm is quartz 5·7, orthoclase 20, albite 40, anorthite 28, with only 2·8 per cent of femic constituents.

Both andesine-anorthosite and labradorite-anorthosite occur in St. Urbain, Quebec. The following modes are given by J. B. Mawdsley.

		andesine-anorthosite,		*labradorite-anorthosite*
orthoclase		8	5	—
plagioclase	An_{38-45}	68·2	—	—
	An_{48-55}	—	87	—
	An_{55-61}	—	—	90
hypersthene		14·8	7·4	7·2
ilmenite		·9	·5	1·6
hornblende, biotite		—	—	1·2

The anorthosites of the Bushveld complex contain almost nothing but plagioclase (bytownite) and bronzite, the latter forming 2 to 6 per cent of the rock (A. L. Hall). Varieties with 15 per cent and more of bronzite are described as anorthositic norite.

G.—ANDESITE. DSγL, DSδL

Probably no rock-name has been so abused as andesite. If one studies the norms of the "andesites" in Washington's

Tables, one finds that quartz is frequently present in quantities greater than 20 per cent, rising to 35 per cent. In other cases there is such a deficiency of silica that 15 per cent of olivine, or 11 per cent of nepheline, appear in the norm. The ratio of orthoclase to anorthite reaches such values as 30 to 1 and 15 to 1, while in some cases anorthite is entirely absent and the norm contains acmite. The proportion of femic minerals in the norm varies from 5 to 50. In our classification, these rocks would be spread over rhyolite, rhyodacite, dacite, trachyte, trachyandesite, basalt, and certain undersaturated groups, and certainly less than half of them would be left in andesite. But most descriptions of rocks that are called "andesite" are so imperfect that it is impossible to place the rocks properly in a quantitative classification.

Peraluminous andesites possibly exist, but it is very difficult to recognize them among the "mica-andesites" that have been described. Nearly all so-called mica-andesites carry hornblende or augite as well as mica, and, in addition, many of them are far too siliceous to deserve the name of andesite at all.

Metaluminous Type

Metaluminous andesites, containing hornblende and biotite, have been described, for example, in the Siebengebirge of the Rhine, in Hungary, Spain, the Andes, Central America, and the Western United States, but quantitative detail is mostly lacking. Many of those rocks are certainly rhyodacites and trachyandesites.

Hornblende-hypersthene andesites have been described on Lassen Peak and Mount Shasta in California, Mount Rainier in Washington, and Mount Hood in Oregon, but as far as one can judge from the norms, these rocks should be classed as dacite.

The lavas of Electric Peak and Sepulchre Mountain, in the Yellowstone National Park, are mostly too acid to be admitted to the group of andesites as defined here, and some of them have orthoclase in excess of anorthite; but about four of the analyses given by J. P. Iddings represent slightly oversaturated rocks with anorthite greater than orthoclase and a colour index apparently about or rather less than 30. Some of the rocks are microcrystalline, but many have a residue of dark glass. Augite, hornblende and biotite all appear among the dark minerals. Hornblende is present in almost every case, and the more hornblende, the less pyroxene does the rock carry. Biotite accompanies the hornblende in those rocks which hold least pyroxene.

Subaluminous Type

Subaluminous andesites are probably commoner than any other type, although a large proportion of the rocks that have been described as augite-andesite and hypersthene-andesite have a colour index greater than 30, and are consequently to be transferred to basalt.

The andesite of Pasto Volcano in Colombia, described by R. Küch, has the following mode: —

quartz	8·9	augite	14·0
orthoclase	9·4	magnetite	8·5
albite	14·9	apatite	0·3
plagioclase (Ab_7An_8)	43·9	Colour index	(23)

The andesites of the San Francisco volcanic district in Arizona, described by H. H. Robinson, are interesting as showing olivine crystals in a glassy base which contains free silica. In the following table the actual mineral composition of these rocks is given, together with the calculated composition :—

		San Francisco Mtn.	*Kendrick Peak*
Actual development of rock	groundmass	27·7	32·2
	feldspar	52·9	53·2
	pyroxene	10·4	7·4
	olivine	2·4	4·2
	magnetite	6·6	3·0
Calculated composition of entire rock	quartz	10·1	8·6
	orthoclase	13·1	12·8
	albite	35·6	33·0
	anorthite	19·2	18·5
	augite	8·0	9·3
	hypersthene	3·6	5·5
	olivine	2·4	4·3
	magnetite	6·9	6·5
	apatite	1·0	0·7
	Colour index	(23)	(27)

The first of these is a dark gray, compact or vesicular rock, with little insets of plagioclase which seldom exceed 1 millimetre in diameter. The groundmass shows little feldspar laths, grains of pyroxene, tiny olivine grains, magnetite and glass. In each of these rocks there is enough free silica to have converted all the olivine into hypersthene, if cooling had taken place with sufficient slowness to allow the reaction to take place. If we perform this operation on paper, the first rock will be enriched by 3·4 per cent of hypersthene, leaving 9 per cent of quartz; and the second will have altogether 11·6 per cent of hypersthene and 6·7 per cent of quartz. Both rocks are therefore abnormally developed andesites.

H.—Gabbro and Soda-Gabbro. XSγM, XSδM

The name gabbro, as used here, stands for the group of plagioclase-bearing rocks of mesotype to melanocratic character. It is divided by the Ab : An ratio into *soda-gabbro* and *lime-gabbro*. The name soda-gabbro covers a group of rocks about which there has been little agreement in the past, some writers putting them in diorite, some in gabbro and some making use of the non-committal name gabbrodiorite. Most of these rocks contain a large proportion of hornblende, so it may be preferable to use soda-gabbro as the group name and leave gabbrodiorite for the varieties rich in hornblende. The lime-gabbro group includes all normal gabbro, norite and diabase (dolerite).

Peraluminous Type

Mica-gabbro containing no other dark silicate than biotite is almost unknown. Among the dyke-rocks, however, the kersantites have the appropriate composition. These rocks consist typically of plagioclase and biotite, but quantitative descriptions are seldom available. J. H. L. Vogt has given the composition of kersantite at Bygdö, Sweden, in the following terms:—labradorite 60 to 70, biotite 20 to 25, magnetite 10 to 12, with some pyrite and apatite.

Metaluminous Type

The simple association of oligoclase with an abundance of hornblende has been observed by S. W. Wooldridge on the island of Sark, Channel Islands. In one case, measurements indicated 48 per cent of oligoclase and 50 of hornblende. This rock bears the same relation to diorite that shonkinite bears to syenite.

An extremely high content of brown hornblende is reported by T. F. W. Barth in a pegmatite at Brandbukampen, Norway. The mode is:—

quartz	0·8	pyroxene	4·4
andesine	10·5	sphene, ore	2·1
biotite	7·1	colour index	(89)
hornblende	75·0		

A simple lime-gabbro which is almost free from pyroxene was described by I. S. Allison in the Giant's Range of Minnesota. It consists roughly of 70 parts of labradorite and 30 parts of hornblende, the crystals sometimes enclosing cores of pyroxene.

F. Sellner describes mica-rich gabbrodiorite in the Brünner massif, Austria, and gives the following modes:—

	(1)	(2)
quartz	4·5	5·0
plagioclase	47·5	48·8
biotite	23·9	19·9
hornblende	22·7	24·6
accessories	1·4	1·7
Colour index	(48)	(46)

The hornblende-gabbro of Frankenstein, Germany, is on the whole a soda-gabbro, the composition of the plagioclase ranging from An_{35} to An_{58}. P. Tschirwinsky has given the following mode:—

plagioclase (An_{35} to An_{58})	62·2
augite	29·9
hornblende	6·2
titanomagnetite	1·6
pyrites	0·1
Colour index	(38)

The hypersthene-hornblende association is illustrated by the norite of St. Thomas' Mount, Madras, which H. S.

Washington (6) has described as follows: plagioclase (Ab_1An_1) 40·8, hornblende 19·6, hypersthene 31, ores 8·6. The hypersthene-biotite association is found in many norites, as at Tripyramid Mtn., New Hampshire. L. V. Pirsson (2) gives the following mode:—

quartz	3·0	augite	5·6
orthoclase	3·0	hypersthene	10·0
plagioclase	56·5	ore	11·0
biotite	6·0	apatite	4·4
		Colour index	(37)

The association of hornblende, augite and biotite is seen in the gabbros of Slesse Mountain and Ashnola, British Columbia (R. A. Daly, 1) and in the "diorite" of Belknap Mountain, New Hampshire (L. V. Pirsson).

	Slesse Mtn.	*Ashnola.*	*Belknap Mtn.*
quartz	9·5	0·5	—
labradorite	58·0	57·5	43·5
biotite	12·5	3·0	3·3
hornblende	12·8	21·8	26·3
augite	4·3	12·0	15·8
ores	2·3	3·6	7·7
accessories	0·6	1·6	3·4
Colour index	(32)	(42)	(56)

The same association is seen in the hybrid "augite-biotite-diorite" of the Newry complex, Ireland. This rock contains 24 per cent of augite (including a little hornblende), 25 per cent of biotite, and 45 of andesine (D. L. Reynolds, 1). With a soda-lime feldspar and a colour index of 55 the rock is a soda-gabbro in terms of the classification followed in this book.

Biotite, hornblende, augite, hypersthene, and even a trace of olivine are all present in a gabbro that J. E. Wolff has described in the Crazy Mts., Montana. Here we have the entire reaction-series of the ferro-magnesian minerals in a single rock. The mode is quartz 1, labradorite 25, biotite 8, hornblende 33, augite 21, hypersthene 4, olivine 1, accessories 7; colour index 74.

A colour index greater than 60 is shown by gabbros that have been described by W. G. Foye at Glamorgan, Ontario, and by G. A. Young at Mount Yamaska, Quebec. The modes are:—

	(1)	(2)
plagioclase	22·6	15·0
hornblende	44·3	21·2
augite	19·5	43·4
ilmenite	12·2	16·7
pyrite, apatite	1·2	3·7
Colour index	(75)	(85)

Norites that have been greatly contaminated by solution of aluminous sediments, leading to the appearance of cordierite, garnet and other highly aluminous minerals, have been described by H. H. Read and W. R. Watt in north-eastern Scotland. The mode of one such rock, occurring near Huntly, has been given as follows by Watt:—

quartz	1·0	hypersthene	3·0
orthoclase	2·8	garnet	16·6
albite	30·9	cordierite	14·4
anorthite	12·2	sillimanite	0·2
biotite	13·5	ilmenite, etc.	4·8
		Colour index	(36)

Subaluminous Type

This type includes common gabbro, norite, and diabase or dolerite. The rocks are common enough, but accurate statements of their mineralogical composition are less common.

B. V. Lombaard has given a number of modes for the different facies of the Bushveld norite. If we exclude the anorthosite facies in which the colour index falls below 30, then the range of composition is labradorite 46 to 70, ortho-pyroxene 3 to 46, clino-pyroxene 1 to 29, accessories 0 to 9; colour index 32 to 54. The average of 13 modes is labradorite 64, ortho-pyroxene 20, clino-pyroxene 15, accessory 1; colour index 36. A very little quartz is present in the more feldspathic facies.

For the Sudbury gabbro or norite, T. C. Phemister has furnished diagrams to indicate the variation in mineralogical composition along certain lines of traverse. It appears from these data that quartz is always present and often in greater amount than 10 per cent; also that a large proportion of the rock is leucocratic. We have already included some examples of the latter type in the lime-diorite group. Coleman, Moore and Walker present an average composition calculated from five analyses of norite from the "basic border," from which it appears that the average feldspar is sodic ($Or_{12}Ab_{57}An_{16}$), which would throw the corresponding rocks into the soda-gabbro group. In short, the Sudbury gabbro is a very variable rock which ranges in composition from tonalite to diorite, soda-gabbro, and leucocratic gabbro or norite. In addition, the original pyroxene has commonly been transformed into hornblende. Compared to the Bushveld norite the Sudbury rock is richer in silica and alkalis and much poorer in lime and magnesia.

W. Larsson has made a detailed study of the Nygard norite stock in Sweden. This little stock of about one square mile shows a wide range of composition, the colour index varying from 25 to 92. The mean of 46 modal analyses is as follows:—labradorite 50, ortho-pyroxene 32, clino-pyroxene 7, biotite 3·8, hornblende 2, ore 3; colour index 50. A little biotite is present in every case, and usually a little hornblende as well. This raises the question whether such rocks should be referred to the subaluminous or to the metaluminous type. So long as the proportion of biotite or hornblende is trivial—say less than one-tenth of the pyroxene—it may be ignored; but in the present instance the average proportion of biotite and hornblende is one-seventh of the pyroxene. One might describe such a rock as a transitional one, and indicate this by the symbol XSs-m8 (50).

The gabbro of the Insizwa lopolith in South Africa is mostly olivine-bearing, but it has an olivine-free facies for which D. L. Scholtz has given the following mode:—quartz 4·2, cryptoperthite 4·4, plagioclase 54, ortho-pyroxene 24, clino-pyroxene 9·6, hornblende 2·2, biotite 1, ore 0·3; colour index 37. In this case the proportion of hornblende and biotite is less than one-tenth of the pyroxene, so for the purpose of classification it may be ignored.

A. Holmes and H. F. Harwood (1) have made a very complete chemical and mineralogical study of the Whin Sill and related dykes in the north of England. The calculated mineral composition of the Whin Sill is given in the first column and that of five of the related dykes in the other columns.

	(1)	(2)	(3)	(4)	(5)	(6)
quartz	6·7	4·6	8·6	9·7	10·5	6·8
alkali-feldspar ...	4·4	6·3	3·0	6·1	3·0	—
plagioclase	40·4	40·5	42·4	43·4	42·5	45·5
pyroxenes	36·7	35·4	29·4	27·6	27·1	32·2
ores	8·9	9·4	12·6	11·9	9·8	8·9
accessories	1·4	·7	1 4	1·5	5·8	1·1
Colour index ...	(47)	(45)	(43)	(41)	(43)	(42)

The plagioclase in these six rocks holds in each case just under 50 per cent of anorthite, and the pyroxene includes both rhombic and monoclinic species.

Quartz-bearing gabbros and diabases of the above character have been widely recognized in Spitzbergen, Canada, eastern United States, British Guiana, Brazil, India, South Africa and Tasmania. F. Walker's data for the Palisade sill in New Jersey were quoted in Chapter IX.

I.—Basalt and Soda-Basalt. DSγM, DSδM

The *peraluminous* type is unknown.

Metaluminous Type

Metaluminous basalts are much less common than the corresponding gabbros. A hornblende basalt was described by H. H. Robinson at Bill Williams Mountain, Arizona. The visible constituents of this rock are hornblende (48 per cent), labradorite (12 per cent) and magnetite (2 per cent), with a cryptocrystalline groundmass which makes up 37 per cent of the rock. If the rock had crystallized completely, it would have had approximately the following composition :—

quartz	7·1	hornblende	47·6
orthoclase	4·4	magnetite	2·0
albite	29·8	apatite	·3
anorthite	8·2	Colour index	(50)

Among dyke-basalts, however, it is likely that many examples of this type could be found. Such are, for instance, many of the so-called "camptonite" dykes. Pirsson and

Washington have described a camptonite of Mount Belknap, New Hampshire, which consists of

labradorite	53·2	apatite	1·9
hornblende	31·7	calcite	6·7
iron ore	6·4	Colour index about 40	

Camptonites from San Carlos Mts., Mexico, have been described as follows by E. H. Watson :—

	(1)	(2)	(3)
andesine	46	—	—
labradorite	—	60	59
brown hornblende	30	13	30
augite	20	19	11
magnetite, etc.	4	8	—
Colour index	(54)	(40)	(41)

Subaluminous Type

Among the basalts and dyke-basalts the Tholey type (tholeiite) corresponds very closely to the diabase of the Whin Sill and Palisade type. Holmes and Harwood (1) have studied the tholeiites of the north of England and give a large number of chemical analyses, together with the calculated composition of each rock analyzed. A single example may suffice. The calculated composition of the Tynemouth dyke is as follows :—

quartz	5·2	pyroxene	37·4
alkali-feldspar	·8	ores	5·8
plagioclase	47·7	accessories	1·5
		Colour index	(45)

The basalts of the Deccan, India, have been studied by H. S. Washington (7). In a region of over 200,000 square miles some variation must be expected, but these basalts are remarkably similar all through the region. They are mainly dense, non-porphyritic rocks with not more than ten per cent of glass. In the majority of cases olivine is entirely absent and there is a small excess of silica. Augite and labradorite make up about 90 per cent of the rocks, and some magnetite is always present.

The following norms, calculated from two of Washington's analyses, give an idea of the mineralogical composition of the rocks :—

	(1)	(2)
quartz	5·5	2·4
orthoclase	5·6	6·7
albite	18·3	25·7
anorthite	21·1	24·5
pyroxene	35·1	35·1
magnetite and ilmenite	12·4	2·8
apatite	0·7	1·7

The Snake River basalts of Oregon are very similar to the Deccan "traps." The norm calculated from six analyses shows

quartz	2·2	pyroxene	32·7
orthoclase	7·8	magnetite and ilmenite	8·9
albite	24·6	apatite	2·0
anorthite	20·6		

H. S. Washington (4) describes subaluminous andesite-basalts at Monte Arci, Sardinia. These are dense, dark gray or black rocks, showing only rare and very small insets of feldspar. Both ortho- and clino-pyroxenes are present, and there is very small residuum of siliceous glass. The following mode is calculated from a chemical analysis :—

quartz (occult)	4	ores	8
Ab_1An_2	62	apatite	1
diopside	5	glass	3
hypersthene	17	Colour index about 31	

Among the basalts of Monte Ferru, in the same island, there are some which show the association of quartz and olivine, of which we have already discussed several instances. The olivine crystals form insets only, while the quartz is confined to the interstitial base of the groundmass. In each case described by Washington (5) there is enough free silica to have converted the whole of the olivine into enstatite, as the following figures show :—

	(1)	(2)
quartz	5	4
Ab_1An_2	57	50
pyroxene	21	32
olivine	10	8
ores	6	5
apatite	1	1
Colour index	(37)	(46)

A few examples of soda-basalt and lime-basalt, containing either no olivine or else a small proportion of olivine together with the corresponding amount of free silica (Bowen-Andersen reaction), have been described from the island groups of the Pacific, by T. F. W. Barth.

J.—PERKNITE. XSH

Under this name we place all saturated rocks which are composed, to the extent of ninety per cent or more, of pyroxene, hornblende or biotite, with which magnetite, ilmenite, chromite, spinel or a sulphide may be associated. An iron-rich olivine or an iron-rich garnet may also be present. Rocks that consist almost entirely of iron-ore (magnetite-rock, ilmenite-rock, chromite-rock) are not considered here. We may subdivide the perknites into ortho-pyroxenite or bronzitite, clino-pyroxenite or diopsidite, garnet-pyroxenite, hornblende-pyroxenite, and hornblendite.

Peraluminous Type

Rocks composed wholly of mica can hardly be said to have an independent existence. Larsen and Pardee have described lenses of biotite-rock, now altered to vermiculite, in pyroxenite at Libby, Montana, and have suggested for them the name glimmerite. Elsewhere this material is only known in nodules.

Metaluminous Type

Biotite-pyroxenite with a large proportion of mica occurs in the Newry complex, Ireland; it has the following composition:—biotite 45, augite 27, hornblende and actinolite 21, ore and apatite 6 (D. L. Reynolds, 1). Volcanic bombs consisting of augite and biotite occur on Monte Somma (A. Rittmann). At Libby, Montana, Larsen and Pardee describe a body of diopside-pyroxenite which contains a variable proportion of biotite and grades into the glimmerite mentioned above. This rock sometimes holds a little garnet, and apatite is always abundant. The mode of a single specimen is stated as follows: diopside 74, biotite 5, garnet 1, ore 10, apatite 10. A diopside-pyroxenite at Loch Ailsh, Scotland, has been described by J. Phemister; it holds diopside 60, biotite 19, hornblende 11, iron ore 9 per cent. Other occurrences of biotite-pyroxenite have been summarized by D. L. Reynolds (2).

In the "yamaskite" of Mount Yamaska, in the Monteregian Hills of Quebec, the pyroxene takes the uncommon form of titanaugite. The mode is titanaugite 57, hornblende 29, ilmenite and pyrite 7, anorthite 2 per cent (G. A. Young).

The hypersthene-hornblende association is exhibited by a rock from Madras, India, which was described by H. S. Washington (6). The mode is hypersthene 46, clino-pyroxene 5, hornblende 41, magnetite 8 per cent.

Biotite-hornblende-bronzite pyroxenite is recorded by R. J. Bridges, in Tanganyika Territory, East Africa. Bronzite forms 10 to 20 per cent of the rock, the remainder being composed of hornblende and biotite in nearly equal proportions.

Hornblendites in which the dark brown hornblende crystals contain cores or ghosts of diallage, thus demonstrating the origin of the rock by transformation of pyroxenite, were described by Wyllie and Scott at Garabal Hill, Scotland. A hornblendite containing 93 per cent of hornblende with 7 per cent of feldspar was described by F. F. Grout (2) at Saganaga Falls, Minnesota. This rock is a marginal facies of the Saganaga granite.

Many pyroxenites hold a little garnet. A. Lacroix (1) has described rocks consisting of diopside and garnet, diallage and garnet, and diallage with hornblende and garnet at Lherz, in the Pyrenees. The eruptive origin of these rocks is perhaps open to question.

Subaluminous Type

Diopside-pyroxenite with only a trifling proportion of biotite or hornblende has been described by W. C. Brögger (2) at Monzoni, Tyrol. Besides three or four per cent of biotite and hornblende the rocks contain a little plagioclase. Brögger (3) has also described pyroxenite of similar composition at Gran, Norway, where it forms a marginal facies of essexite. The pyroxenite of Palabora, Transvaal, consists of pale green diopside with only a little microcline and apatite.

The "websterite" of North Carolina contains both rhombic and monoclinic pyroxene. F. Bascom has described a rock of this character at Oakwood, Maryland, which consists simply of hypersthene (43 per cent) and diallage (56 per cent).

Ortho-pyroxenite, consisting of bronzite with no more than 5 or 6 per cent of plagioclase, diopside and chromite occurs largely in the deeper zones of the Bushveld norite lopolith (P. A. Wagner). Entirely similar rocks are found in the Great Dyke of Rhodesia and in the Stillwater complex of Montana.

Many pyroxenites carry a moderate or even large proportion of oxidic or suphidic ore. P. Geijer has described the passage of a syenite, at Rackberget, Sweden, into a rock composed of pyroxene, fayalite and magnetite in the proportion (to judge from a single text-figure) of about 50 : 20 : 30. The "Pusey iron ore" of Glamorgan, Ontario, is a heavy, black rock made up essentially of augite and titanomagnetite, the latter constituting from one quarter to over three quarters of the rock (Adams and Barlow). The pyroxene grains are completely enclosed in magnetite, but about each grain there is generally a narrow mantle of brown hornblende which has been formed from the pyroxene.

Peralkaline Type

Perknites consisting mainly of soda-pyroxene, generally a more or less sodic diopside, have been found in association with peralkaline syenite at Ice River, B.C., and elsewhere. Volcanic bombs of such material appear at Monte Vulture, Italy. One of these has the following composition: ægirine-diopside 39, hornblende 42, biotite 6, ore 6, apatite, etc., 7 per cent (A. Rittmann).

The original *eclogite* of Bavaria contains a green soda-pyroxene (omphacite or jadeite-diopside) with green hornblende and pink garnet. The eclogites are regarded by most people as metamorphosed basalts and tuffs.

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CHAPTER XXII

THE UNDERSATURATED ROCKS (NON-FELDSPATHOIDAL DIVISION)

(DESCRIPTION)

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A.—MINERALOGY

THE mineralogy of these rocks is the same as that of the saturated rocks, except for the appearance of the unsaturated minerals olivine, melilite, melanite, perovskite and corundum. Quartz and siliceous glass are entirely absent except in rare instances where the Bowen-Andersen reaction has been inhibited by rapid cooling of the magma.

Olivine is seldom associated with a potassic feldspar, for the reason that orthoclase can react with olivine to give hypersthene and biotite. Even the soda-rich feldspars are rarely found in association with olivine, and in general the feldspar of the olivine-bearing rocks ranges from labradorite to anorthite. Olivine crystallizes as a rule before all other silicates except anorthite. A. Harker has shown that in rocks that consist chiefly of these two minerals, like the allivalite and harrisite of the island of Rum, the mineral which is present in larger quantity crystallizes first. But in consequence of the great difference in melting-point between forsterite and fayalite, an iron-rich olivine crystallizes at a much lower temperature than the common iron-poor variety. S. I. Tomkeieff (1) has shown that the olivine in many dolerites and diabases is zoned and the outer zones are richer in iron than the cores.

In British Tertiary dolerites the proportion of fayalite in the cores of the olivine crystals is 18 per cent, but in the outer zones it is as much as 40 per cent.

Melilite is only found in rocks with an excessively high content of lime. It is seldom found apart from olivine, and it is generally the last of all the heavy silicates to crystallize. In the alnöite of Montreal, J. Stansfield found that the order of crystallization was olivine, iron ore, perovskite, biotite, melilite; and in the Oka Hills it was apatite, ore, mica, perovskite, melilite, followed by nepheline and haüyne. In the melilite-basalts of Namaqualand, South Africa, M. S. Taljaard observed that an early generation of pyroxene was resorbed and transformed into melilite, after which there was a later generation of titanaugite. Perovskite is an almost invariable associate of melilite and is seldom found in any other association.

Melanite is only found in a few syenites and trachytes. Its commonest associate is a more or less sodic diopside, though in borolanite it is associated with biotite.

The only significant occurrences of corundum are in pegmatite dykes, as already related. In these the mineral is associated with feldspar of low potash content, usually a soda-rich microperthite, albite or oligoclase; but in the kyschtymite of the Ural Mts. the feldspar is anorthite. The only constant associate of corundum is biotite.

B.—CLASSIFICATION AND NOMENCLATURE

The method of subdivision is strictly parallel to that used for the saturated rocks. The difficulty of nomenclature is greater than usual, because for some reason that is hard to understand the presence of olivine in a rock has seldom been indicated by a distinctive name. It is true that in Zirkel's time the name basalt implied the presence of olivine, and if one wished to indicate that a basaltic rock was olivine-free one had to say so in these words, or else call it augite-andesite; but this rule has been abandoned in English-speaking countries. Basanite remains one of the very few names that definitely imply the presence of olivine; peridotite is another and troctolite is a third, but the latter rock seems to have received its name more on account of its curious appearance than for any mineralogical reason. This neglect to mark the appearance of olivine in rocks by a change of name is astonishing when one considers how gabbro has been split up on the slenderest of mineralogical or textural grounds into norite, ossipite, hyperite, anorthosite, eucrite, gabbrodiorite, essexite, diabase, picrite, belugite, tilaite, and so on. Not one of these names indicates anything half so important as the crossing of a phase boundary.

To make the position still more remarkable, the extremely rare melilite rocks have received no less than 17 specific names (not counting compound names); and the decidedly uncommon leucite rocks have at least 34 such names (including four new names given quite recently); yet the supremely important and chemically distinct olivine-bearing rocks are treated as mere "varieties" of gabbro, norite, diabase and basalt, unworthy of distinctive names, and eminent petrographers have subscribed to the opinion that the appearance of olivine in a rock is not especially deserving of notice.

There is need of a complete series of names to distinguish the olivine-bearing rocks from those that are free from olivine, just as basanite has always been distinguished from tephrite. One would like to see the world-wide olivine-basalt magma distinguished from the olivine-free basalt of the continental shields by a name that would recall the dominance of olivine-basalt in the great ocean basins; unfortunately *pacificite* and *oceanite* have already been given to mere varieties, and *atlantite* to a nepheline-bearing lava. Alternatively, one might restore the original meaning of basalt, and use the already fairly familiar *tholeiite* for the olivine-free type. But even if petrologists could be induced to make this small change it would not be enough; we should still be in need of names for the olivine-bearing equivalents of gabbro, diorite, shonkinite, andesite and trachyte at least. There would be the fullest justification for introducing such names—but nobody would use them. The writer has therefore followed the simple plan of adding the prefix *sub-* to the name of any saturated rock, in order to indicate the presence of an unsaturated mineral of the group we are now discussing. The broad classification of these rocks proceeds then as follows:—

EUCRYSTALLINE ROCKS

		Leucocratic	*Mesotype-Melanocratic*
Or>An	Or>Ab	Potash sub-syenite	Sub-shonkinite
	Ab>Or	Soda sub-syenite	Sub-monzonite
An>Or	Ab>An	Soda sub-diorite	Soda sub-gabbro
	An>Ab	Lime sub-diorite	Lime sub-gabbro
			Hypermelanic
			Peridotite

DYSCRYSTALLINE ROCKS

		Leucocratic	*Mesotype-Melanocratic*
Or>An	Or>Ab	Potash sub-trachyte	(No distinctive name)
	Ab>Or	Soda sub-trachyte	(No distinctive name)
An>Or	Ab>An	Soda sub-andesite	Soda sub-basalt
	An>Ab	Lime sub-andesite	Lime sub-basalt
			Hypermelanic
			Melilite-basalt
			Kimberlite

It is to be remembered that these awkward hyphenated names are *group-names;* they are not intended for application to individual rocks.

C.—SUB-SYENITE. XVαL, XVβL

Peraluminous Type

The corundum-syenites and pegmatites of Ontario have as their feldspar a microperthite in which albite is in excess of orthoclase. The only constant dark minerals, besides corundum, are large plates of biotite or muscovite. Corundum is variable in amount, but locally it makes up as much as 75 per cent of the rock, and individual corundum crystals sometimes weigh more than thirty pounds (A. E. Barlow). The corundum pegmatites are intersected by quartz feldspar-pegmatites which hold no corundum.

The corundum rocks of the Ilmen Mountains, Russia, are very similar to those of Ontario. The syenite of Nikolskaia Ssopka contains a microperthite in which albite slightly exceeds orthoclase. The corundum is in short, blue prisms; in one example analyzed by Morozewicz, corundum made up 18·5 per cent of the rock, and the only other heavy minerals present are a little biotite and some secondary white mica. At Sseljankina the rock holds a good deal of biotite; and in the corundum pegmatites of the Borsowka River there are traces of rutile, apatite and zircon.

In the corundum syenite of Coimbatore, Madras, the same mineralogical characters are repeated. T. H. Holland reports biotite, muscovite, apatite, zircon, spinel, chrysoberyl and pyrope as accessory constituents of a rock otherwise composed of microperthite and corundum.

Metaluminous and Subaluminous Types

The appearance of olivine in a syenitic rock is generally accompanied by such an increase in the colour index that the rock becomes either shonkinite or monzonite. Examples will be found under these heads.

J. J. O'Neill has described a local facies of the essexite of St. Hilaire Mountain, Quebec, which holds over 70 per cent of soda-orthoclase, together with olivine, brown hornblende, biotite, apatite and iron ore, in that order of abundance.

A. Rittmann has recorded a hornblende-melanite syenite-porphyry, and a similar rock with augite instead of hornblende, among the ejected blocks of Mte. Somma. The modes are as follows:—

	(1)	(2)
sanidine	71	67
plagioclase	9	12
biotite	3	—
hornblende	13	1
augite	—	16
melanite	3	2
rest	1	3
Colour index	(20)	(22)

Peralkaline Type

This type is represented by a very few examples of melanite-syenite. Most rocks of this mineralogical character are mesotype or melanocratic and fall under shonkinite rather than syenite.

P. Eskola has described syenitic rocks with a considerable content of deep brown garnet (andradite or melanite) at Sviatoy Noss in Transbaikalia. The calculated composition of two of these rocks is given below. Eskola states that "no quartz was detected under the microscope in spite of much search with convergent light."

	(1)	(2)
orthoclase	24·8	31·1
albite	41·5	37·7
anorthite	3·2	10·8
soda-diopside	17·1	5·7
andradite	9·0	10·2
rest	3·4	2·4
Colour index	(30)	(18)

Melanite syenite has been recorded at several points in the limestone districts of Assynt, Scotland. At Loch Borolan it forms a transition rock between quartz-bearing syenites and nepheline-syenite, and the melanite is associated with biotite (S. J. S., 1); near Loch Ailsh, J. Phemister has found another variety consisting of soda-rich microperthite, ægirine and melanite.

D.—Sub-Trachyte. DVα, DVβ

The few rocks that fall into this group are more commonly mesotype than leucocratic, and nearly all are subaluminous.

The iron-olivine, fayalite, is not uncommon in rhyolite and trachyte; it occurs pretty constantly in the quartz-trachytes of Kenya, for instance; but magnesian olivine also occurs in some quartz-free trachytes. The presence of occasional grains of olivine in the trachytes of the Auvergne was recorded by Zirkel and by Rosenbusch, and the following proportions were found by the writer (S. J. S., 2) in a trachyte of Mont Dore:—

	Insets	Groundmass
soda-potash feldspar	16	49
diopside	2·8	20
olivine	2	—
iron-ore	—	9
apatite, biotite	·5	·7
Colour index	(35)	

The olivine in this case is not fayalite but a normal magnesian olivine.

The "ciminite" of Monte Cimino, Italy, as described by H. S. Washington, shows the association of a not very sodic orthoclase ($Or_{18}Ab_3$) with a decidedly calcic plagioclase (Ab_2An_3), but the bulk composition of the feldspar is such as to place the rock in the trachyte division. The mode is:—

soda-orthoclase	50·7	olivine	11·2
plagioclase	13·1	magnetite, apatite	1·8
augite	23·2	Colour index	(36)

An "oligoclase andesite" of Kohala, Hawaii, contains so much alkali-feldspar that it is better described as an olivine soda-trachyte (soda sub-trachyte). T. F. W. Barth has given the following mode:—

alkali-feldspar	39·0
plagioclase	30·0
pyroxene	14·3
olivine	7·4
ore	7·2
apatite	3·0
Colour index	(32)

The lavas described by H. H. Thomas as "marloesite" and "skomerite" are soda-rich olivine-trachytes, sometimes with and sometimes without hornblende. The "kaiwekite" of New Zealand (P. Marshall) and the "macedonite" of Victoria (Skeats and Summers) are rocks of similar character.

The common "minette" or "mica-trap" may be described as a melanic dyke-trachyte. L. V. Pirsson has described such rocks in Montana, and the following modes refer to minettes of Yogo Peak (1) and the Highwood Mountains (2):—

	(1)	(2)
alkali-feldspar	25	39·2
andesine	10	—
biotite	18	16·8
pyroxene	35	29·3
olivine	7	7·6
rest	5	8·1
Colour index	(65)	(61)

R. A. Daly shows that the shonkinite of Christina Lake, B.C. (page 429), has almost exactly the same chemical composition as the average minette of that region, and suspects that both were formed from the same magma.

An example of melanite trachyte, containing a soda-bearing pyroxene and therefore peralkaline, was recorded by J. D. Mackenzie among the Crowsnest volcanics in Alberta, Canada.

E.—SUB-SHONKINITE, SUB-MONZONITE. XVαM, XVβM

The original shonkinite of the Highwood Mts., Montana, is largely olivine-shonkinite. J. D. Barksdale has given a number of modes which show the following range of composition :—

feldspar, zeolites	29 to 53
biotite	11 to 22
augite	20 to 50
olivine	1 to 10
rest	2 to 5
Colour index	(47 to 71)

The feldspar in this rock contains much more potash than soda.

A typical mode given by L. V. Pirsson for the shonkinite of Middle Peak is given in column 1, below. Beside it is the mode of a shonkinite from Spanish Peaks, Colorado, by A. Knopf. In both rocks potash predominates largely over soda.

	(1)	(2)
soda-orthoclase	53·2	53·7
biotite	4·0	11·5
augite	32·3	24·8
olivine	1·7	5·4
ore and apatite	8·8	4·6
Colour index	(47)	(46)

The olivine-shonkinite recorded by Anderson and Kirkham in south-eastern Idaho is more melanic, as the following modes show, but the feldspar is still dominantly potassic :—

	(1)	(2)	(3)	(4)
orthoclase	35	30	40	32
andesine	5	10	8	2
biotite	5	12	12	15
diopside	42	30	20	42
olivine	10	15	20	8
magnetite, etc.	4	3	2	2
Colour index	(60)	(60)	(54)	(66)

A mesotype olivine-shonkinite was described by R. A. Daly at Christina Lake and other points in British Columbia. It is a coarse grained, dark grey rock composed of soda-orthoclase, labradorite, diopside, olivine, and large foils of dark green biotite. The mode is:—

soda-orthoclase	45·1	olivine	10·7
labradorite	10·6	magnetite	3·5
biotite	5·2	apatite	0.4
diopside	24·5	Colour index	(44)

In spite of the presence of plagioclase in appreciable quantity, orthoclase still exceeds albite in the feldspar. With further increase in the content of plagioclase at the expense of orthoclase, olivine-shonkinite becomes olivine-monzonite. W. B. Brögger describes an olivine-monzonite at Smalingen, Sweden, which has the following composition:—

orthoclase	20	biotite	6
andesine	33	magnetite	3
pyroxene	25	apatite, etc.	1
olivine	12	Colour index	(47)

A similar type of rock occurs at Kentallen and a few other places in Scotland, and has received the name of "kentallenite." According to J. J. H. Teall, the Kentallen rock is composed of olivine and augite, embedded in a groundmass of biotite, plagioclase and orthoclase, the last two approximately equal in quantity.

W. E. Tröger gives the following mode for the original kentallenite:—orthoclase 22, plagioclase (An_{22}) 20, biotite 10, augite 24, olivine 22, rest 2; colour index 58. Chemical analysis shows that the proportions of Or and Ab are almost equal.

Both olivine-shonkinite and olivine-monzonite are described by I. A. Brown, at Mount Dromedary, New South Wales. The normative composition of the monzonite is orthoclase 19·5, albite 16·2, anorthite 27·8, pyroxene 24, olivine 5·7, rest 4·9. Some of the potash which enters normative orthoclase is actually present in the form of biotite, so orthoclase and albite must be nearly equal as in kentallenite.

Shonkinitic rocks containing a lime-garnet occur at Poohbah Lake, Ontario, in association with nepheline-bearing rocks. A. C. Lawson gave the name "malignite" to the entire series, but the writer thinks it preferable to use the older name shonkinite for the varieties that lack nepheline. One of these garnet-bearing shonkinites has the following very unusual composition (S. J. S., 5) :—

potash-feldspar	39
diopside (including a little biotite and apatite)	54
garnet (andradite)	4
sphene	3
Colour index	(61)

F.- –SUB-DIORITE, SUB-ANORTHOSITE. XVγL, XVδL

Peraluminous Type

The "plumasite" of Plumas Co., California, as described by A. C. Lawson, is a corundum-oligoclase rock, the proportions of these minerals being, in a typical instance, oligoclase 83, corundum 16. Chemical analysis of the rock shows that no orthoclase is present at all.

A rock described by Adams and Barlow under the name of "dungannonite" is virtually the same as Lawson's plumasite. It consists of plagioclase ranging from albite to andesine, which makes up 75 to 95 per cent of the rock, together with corundum, biotite, muscovite, calcite, magnetite and scapolite. Nepheline is sometimes present in addition, taking the rock over to the feldspathoidal division of the undersaturated rocks.

The corundum pegmatites of the Transvaal are also similar to the plumasite of California. As described by A. L. Hall, they consist of plagioclase which varies from oligoclase to andesine (and is apparently free from potash, according to the analyses available), together with gray corundum and some biotite. Margarite and gibbsite are secondary products. The proportion of corundum is variable, but in an average specimen of reef Hall found it to be 49 per cent by weight, which is equal to 30 per cent. by volume.

Metaluminous and Subaluminous Types

The "essexite" of Brome Mountain, Quebec, was described by J. A. Dresser; it consists, to the extent of about 90 per cent, of labradorite or bytownite, with a little microperthite. The dark constituents are hornblende, augite, biotite, olivine, sphene, apatite and magnetite.

The subaluminous type is very common. We saw in Chapter XV that a certain proportion of the norite of Sudbury is leucocratic and falls therefore into the systematic pigeon-hole called lime-diorite. Similarly a good deal of certain well-known bodies of olivine-gabbro and olivine-norite falls into lime sub-diorite in this system. This does not mean that the writer wishes to hear—let us say—the Duluth gabbro described in future as the "Duluth lime sub-diorite." It only means that for the purpose of exact classification the more feldspathic gabbros are joined to the anorthosites rather than to the melanic gabbros. There has always been a boundary between "gabbro" and "anorthosite," and all that the writer seeks to do is to define that hitherto indefinite boundary by the colour index 30. In absence of any name that would command general acceptance, he has labelled this particular pigeon-hole "lime sub-diorite."

F. F. Grout has given a large number of measured modes of the Duluth gabbro, from which it appears that a good deal of this rock is leucocratic and falls into lime sub-diorite. Out of 58 modes given by Grout, 17 are of this character; that is to say, they show plagioclase in excess of 70 per cent, with 5 to 15 per cent of olivine, about the same of augite, and a little ore.

The olivine-norite of Sierra Leone also has such leucocratic facies. A chemical analysis furnished by F. Dixey shows only 26 per cent of femic minerals in the norm, and a micrometric analysis by the writer gave the following result: labradorite 74, olivine 14, pyroxene 7, magnetite 2·3; colour index 25.

A. Harker has given the modes of some gabbros and dolerites of the island of Skye, and it appears that these, too, fall within the sub-diorite group of our system. A gabbro of the Cuillin Hills gave 80 per cent (by weight) of labradorite; 16 of augite and olivine; 2 of ortho-pyroxene, and 2 of magnetite. A dolerite near Portree was found to hold roughly 77 per cent of labradorite, 11 of augite, and 12 of olivine and accessory minerals.

G.—SUB-ANDESITE. DVγL, DVδL

We have seen that olivine-gabbro is frequently leucocratic; the Sudbury gabbro, for instance, has facies which are olivine bearing and yet contain over 90 per cent of feldspar; but a leucocratic olivine-basalt is a very uncommon thing. T. F. W. Barth reports that such rocks are non-existent in the Pacific basin, the nearest approach to this type of rock being an "olivine-oligoclase-andesite" described by H. S. Washington from Kohala, Hawaii. Judged by the norm this rock must have a colour index of about 33, so it is just excluded from the group. One or two "andesine-andesites" from Etna, which have been described by Washington, Aurousseau and Keyes, must have a colour index close to 30.

The writer has described a feldspar-rich basalt at Kijabe, Kenya, which has a colour index as low as 25. The mode is approximately :—

plagioclase	insets	35	(volume)
	groundmass	35	
zeolites		5	
augite		12½	
olivine		2½	
iron-ore		6½	
apatite		3½	

This rock has undoubtedly been enriched with plagioclase either by gas-flotation or by gravity-flotation (S. J. S., 3). In this respect it resembles the anorthosites, so it is right that it should fall into the same pigeon-hole in the classification.

Basaltic lavas with an exceptionally large content of large feldspar insets have been recorded elsewhere; for instance, the "big feldspar basalts" of Mull, Scotland; but quantitative data are lacking.

The "Markle" type of basalt, well known in the midlands of Scotland, is a highly feldspathic olivine-basalt. Judged by the only available chemical analysis the original rock from Markle Quarry, East Lothian, must have about 80 per cent of plagioclase containing slightly more albite than anorthite.

Some of the "mugearite" or oligoclase-basalt of Scotland is also highly feldspathic. An example described by A. Harker has the following composition :—orthoclase 12·5, oligoclase 57·5, olivine with subordinate augite and ore 26·5; apatite 3·5; colour index 30 (by weight). A. G. MacGregor issues a warning that rocks similar to mugearite may be produced by the albitization of normal basalts.

The lavas of Pembrokeshire to which H. H. Thomas gave the name "marloesite" fall partly into this group. The feldspar is albite-oligoclase and the colour index may be as low as 15.

H.—SUB-GABBRO. XVγM, XVδM

Peraluminous Type

This type is represented by the "kyschtymite" or corundum-anorthosite of the Ilmen Mountains in Russia. J. Morozewicz gives the following data for an example of this rock:—anorthite 38, biotite 10, corundum 47·5, spinel 3 to 4; colour index 62. The order of crystallization was zircon, spinel, corundum, anorthite, biotite.

Metaluminous Type

The essexites of Norway, widely known through the work of W. C. Brögger, are examples of metaluminous sub-gabbro. T. F. W. Barth (4) has furnished a number of modes of these rocks (which he calls Oslo-essexite, to avoid confusion with other rocks to which the name essexite has been given). The range of composition is as follows:—

quartz	0 to 2	hornblende	0 to 3
alkali-feldspar	5 to 22	olivine	3 to 11
plagioclase	25 to 57	ore, apatite	4 to 10
biotite	3 to 14	Colour index...	(30) to (60)

A rock that A. Knopf has called "alkali-gabbro," occurring at Huerfano Butte, Colorado, is very similar to these Norwegian essexites:—

Potash feldspar	6	Augite	27·8
Labradorite	41·2	Olivine	6·5
Biotite	12·3	Ore, etc.	6·2
		Colour index	(52)

T. F. W. Barth (2) has described a number of hornblende-olivine gabbros in Seiland, Norway. The following modes are typical :—

	(1)	(2)	(3)
Plagioclase (An 45-60)	60	50	20
Hornblende	5	20	25
Augite	5	15	20
Ortho-pyroxene	5	5	15
Olivine	15	5	15
Spinel	5	—	—
Colour index	(35)	(50)	(75)

The same type of rock has been described by C. O. Hutton at Lake Wakatipu, New Zealand.

Subaluminous Type

The feldspar in these rocks is generally a calcic variety. A rock in which albite exceeds anorthite is the dolerite of the Prospect intrusion, near Sydney, New South Wales. This rock holds plagioclase of variable composition, but the mean is Ab_4An_3. The mode is given as follows by Jevons and colleagues :—

plagioclase	36	ilmenite and magnetite	13
titanaugite	36	accessory and secondary	5
olivine	10	Colour index	(64)

In the majority of cases the feldspar is labradorite to bytownite. The olivine gabbro of Pigeon Point, Minnesota, has the following composition, according to W. S. Bayley :—

labradorite (Ab_3An_4)	60	augite	27
olivine	10	apatite, etc.	3
		Colour index	(40)

In the olivine-gabbro of Duluth, Minnesota, the composition of the feldspar does not vary much from Ab_1An_2 according to F. F. Grout. The normal proportion of olivine is about 20 per cent; augite about the same amount; and ore about 10 per cent The plagioclase in the olivine-diabase of the Palisades, New Jersey, is also about Ab_1An_2 (F. Walker), and the composition of this rock is on the average plagioclase 33, pyroxene 40, olivine 22, rest 5; colour index 67.

The plagioclase in the gabbro of Tutuila, Samoa, has the composition Ab_3An_7, according to R. A. Daly, and the rock contains plagioclase 45, augite 24, olivine 25, ore and apatite 6·7; colour index 55.

A still more calcic feldspar (Ab_1An_3) was recorded by J. E. Pogue in an olivine-diabase in North Carolina. This rock holds plagioclase 46, augite 35, olivine 17, ore 2; colour index 54.

The plagioclase crystals in these rocks so commonly have a zonal structure that it is difficult to establish their average composition without chemical analysis. D. L. Scholtz observes that in the Insizwa norite of South Africa the composition of the plagioclase changes according to the abundance of olivine, labradorite in the normal norite becoming bytownite in the olivine-rich troctolite phase. The mean composition of the Insizwa norite, according to Scholtz' data, is about plagioclase 46, ortho-pyroxene 17, clino-pyroxene 15, olivine 19, rest 3; colour index 54.

The name "eucrite" is sometimes given to gabbros in which the feldspar approaches pure anorthite in composition. The "allivalite" of A. Harker is an anorthite-olivine rock of which anorthite constitutes half or more; and "harrisite" is the complementary type in which olivine preponderates. Such rocks are well developed in the islands of Rum and Mull, Scotland, and they differ from eucrite only by the virtual absence of pyroxene. Picotite or chromite is a common accessory constituent.

Melanocratic varieties of olivine-gabbro receive the name of picrite. F. Walker has shown that the accumulation of olivine crystals towards the base of a thick dolerite sill in the Shiant Isles, west of Scotland, has generated a picrite with the following composition:—plagioclase 26, augite 10, olivine 59, ore and zeolites 5; colour index 71. D. L. Scholtz has described a similar phenomenon at the base of the Insizwa sheet and gives the following representative modes:—

	(1)	(2)	(3)
plagioclase	26·8	18·0	10·2
ortho-pyroxene	14·0	11·0	4·6
clino-pyroxene	7·8	5·2	3·1
olivine	49·4	63·9	80·0
Colour index	(73)	(82)	(90)

An extremely melanocratic type, which is almost a peridotite, is the "cumberlandite" of Rhode Island, described by Warren and Powers. This rock contains 30 per cent of iron, and was once mined for use as an iron ore. The mode is:—

labradorite	10	magnetite	18
olivine (rich in iron)	40	ilmenite	20
serpentine, etc.	9	spinel	3
		Colour index	(90)

The name "ferrogabbro" was applied by Wager and Deer to some remarkable rocks of the Skaergaard intrusion, E. Greenland. These consist of an intermediate plagioclase, together with a very iron-rich olivine (from 56 to 100 per cent fayalite) and an iron-rich clinopyroxene (from 45 to 70 per cent ferrosilite).

I.—SUB-BASALT. DVγM, DVδM

Metaluminous Type

Lavas of this type seem to be scarce. V. M. Goldschmidt has described lavas of N.W. Spitzbergen which hold soda-plagioclase (Ab_3An) together with titanaugite, olivine, enstatite, brown hornblende of late crystallization, ore and glass. These lavas are remarkable on account of the abundant nodules of enstatite, olivine, diopside and spinel which they carry.

The original "camptonite" of Campton, New Hampshire, falls in this type. W. E. Tröger gives the mode as follows:—zoned plagioclase 47, brown hornblende 10, titanaugite 31, olivine 9, rest 3; colour index 53. For another camptonite at Lausitz, Saxony, Tröger gives this mode:—zoned plagioclase (average Ab_{58}) 40, hornblende 19, titanaugite 24, olivine 9, ore and apatite 8; colour index 60.

Subaluminous Type

Olivine soda-basalt is a fairly common type. It includes rocks that have been called andesine-basalt, kauaiite, kohalaite. T. F. W. Barth has given the modes of a number of these rocks from the Hawaiian islands. They have the appearance of normal basalts, with insets of olivine and sometimes of plagioclase, but most of them contain some interstitial alkali-feldspar and there may be a little residual glass. The following modes are typical :—

	(1)	(2)
alkali-feldspar	17	14
plagioclase	41	42
pyroxene	22	29
olivine	10	2
ore	8	11
apatite	2	1
Colour index	(42)	(43)

In the first of these rocks the normative feldspar has the composition Or 8, Ab 34, An 19; in the second, Or 6, Ab 36, An 17.

Normal olivine-basalt with An>Ab is such a common rock that there is no lack of modes. Most of these rocks, when fresh, contain only the four minerals plagioclase, augite, olivine, and iron ore, with a little apatite. An average computed from 60 published modes gives the following result :—plagioclase 48, augite 29, olivine 15, ore 8; colour index 52. Excellent descriptions of oceanic basalts have been given especially by Cross and Washington for Hawaii, by A. Lacroix for southern Polynesia, and a most useful summary of all that is known of the composition of Pacific lavas by T. F. W. Barth.

The proportions of the constituent minerals vary considerably. Examples of ultra-feldspathic basalts have already been given (page 433). Basalts with an exceptionally large content of olivine have been recorded in Hawaii by Daly; in Antrim, Ireland, by S. I. Tomkeieff; and in Juan Fernandez by P. D. Quensel, the modes being given in that order :—

	(1)	(2)	(3)
plagioclase	35·7	33·2	23
pyroxene	27·0	17·6	27
olivine	32·0	35·3	45
ore, apatite	5·3	4·1	5
chlorite, zeolites	—	9·6	—
Colour index	(64)	(66)	(77)

Quensel has given the name picrite-basalt to these olivine-rich rocks; others have called them chrysophyric basalt.

The pyroxene content is also highly variable. The Abyssinian basalt named "tokéite" by Duparc and Molly holds 60 per cent of augite, and the so-called "augitites" of Kenya hold some 50 per cent (S. J. S., 3). The "oceanite" of Réunion, named and described by A. Lacroix, contains according to Tröger's measurements 40 per cent of titanaugite and 34 per cent of olivine, with a colour index of 80. All these melanocratic rocks are abnormal varieties produced by the accumulation of early-formed crystals under the influence of gravity or convection currents.

When the norm of a basalt or olivine-basalt is computed from a chemical analysis, it sometimes happens that a little nepheline appears in the norm although no trace of any feldspathoid can be detected in the rock. If this was a rare occurrence one might put it down to analytical errors, soda and alumina being more subject to error than any other rock constituents; but the same curious result has been got by many different analysts. The assumption has been made that the molecule of "soda-anorthite," $Na_2Al_2Si_2O_8$, is present in the plagioclase, or in other words that a certain number of atoms of Ca have been replaced by Na_2 in the plagioclase structure. For lavas showing this peculiarity T. F. W. Barth has proposed the name "pacificite" although they are not restricted to the Pacific or even to the ocean basins.

J.—PERIDOTITE AND OTHER HYPERMELANIC ROCKS. XVII

The most general name for a rock composed almost wholly of dark minerals, including olivine, is *peridotite;* when the olivine is unaccompanied by other silicates (except in trivial quantity) the name *dunite* is used. There is no general name for melilite rocks, but the only known example of a melilite-pyroxenite has received the name *uncompahgrite*. Pyroxenites which carry an unsaturated garnet (usually melanite) are very uncommon rocks; the name *cromaltite* has been given to a rock of this type in the north of Scotland, which is a product of reaction between alkaline magma and limestone.

Metaluminous Type

H. S. Washington has described a hornblende-rich peridotite at Bahia, Brazil, with the following simple composition :—

hornblende	91·3
olivine	3·6
magnetite	5·1

In the hornblende-peridotite of Plumas Co., California, described by H. W. Turner, hornblende is quite subordinate to olivine. This rock holds:—

olivine	45	hornblende	19.6
serpentine	33	magnetite	1·4

A mica-rich peridotite, which is comparable with the micaceous type of kimberlite, is described in Arkansas by Miser and Ross. The mode is:—

phlogopite	56	augite	6
olivine serpentine	35	perovskite	1
		magnetite	2

Mica is also a large constituent of the "scyelite" which was described by J. W. Judd in Caithness, Scotland. It contains:—

hornblende	58·5	serpentinized olivine	22
mica	18·5	magnetite, etc.	1

In the next four examples, pyroxene is an important constituent in addition to hornblende or mica. The first is a peridotite of Tanganyika, East Africa, described by R. J. Bridges:—

biotite	17	olivine	16
hornblende	37	magnetite	2
augite	28		

G. W. Tyrrell describes a metaluminous peridotite near Ayr, Scotland, which has the following mode:—

olivine	65·2	biotite	2·0
titanaugite	20·5	ore	2·3
hornblende	10·0		

A hypersthene-bearing example is described by H. S. Washington at Bahia, Brazil; it forms the outer facies of an intrusive body, the core of which is the hornblende peridotite described above. The mode is:—

hornblende	29·6	olivine	7·5
hypersthene	59·4	spinel	3·5

The "yamaskite" of Rougemont, Quebec, is a pyroxene-rich peridotite with a small content of anorthite. J. J. O'Neill gives the following statement of the composition of yamaskite :—

anorthite	6·1	hornblende	8·8
augite	70·7	biotite	0·5
olivine	13·4	ore	0·5

In a pyroxenite described by Cross and Larsen at Iron Hill, Colorado, the only unsaturated mineral is perovskite. The proportions are pyroxene 75, biotite 10, perovskite 5, apatite 5, magnetite 5.

Rocks which contain melilite, with or without olivine (and of course without feldspar or feldspathoid) have their place here. The most remarkable of these rocks is the "uncompahgrite" of E. S. Larsen and J. F. Hunter, from Gunnison Co., Colorado. This is a coarse-grained rock, about two-thirds of which is melilite. Crystals and grains of pyroxene, magnetite, perovskite, apatite, biotite, and calcite are enclosed in large crystals of melilite which are sometimes as much as a foot in diameter. The second mineral, as regards abundance, is generally perovskite. The melilite crystals alter into a fine-grained aggregate of garnet, diopside, and vesuvianite. This is one of the very few eucrystalline rocks which contain melilite.

Subaluminous Type

These rocks consist largely of olivine and various pyroxenes, mica and hornblende being either absent or present only in trivial amount. The typical "dunite" of Dun Mountain, New Zealand, is almost a pure olivine rock, grains of chrome spinel and rare crystals of diopside being the only other constituents (Bell, Clarke and Marshall). It is associated with various pyroxene-bearing peridotites and with pyroxenites.

A remarkable porphyritic dunite has been described by A. Harker, in the island of Skye. It has large crystals of olivine set in a groundmass of smaller olivine grains and octahedrons of picotite (chrome-spinel). Another dunite of very simple composition is recorded by T. F. W. Barth at Seiland, Norway; it contains 88 parts of olivine and 11 of spinel, with a mere trace of pyroxene.

Olivine rocks are often rich in oxidic or sulphidic ores. M. K. Palmunen describes an "ilmenite-magnetite olivinite" at Susimäki, Finland, which carries a third of its weight of iron oxides. The olivine is itself richer than usual in iron, being about $Fo_{55}Fa_{45}$, and a little brown hornblende is present. The mode is:—

olivine	60·3 (by weight)	ilmenite magnetite	36·3
hornblende	3·0		
spinel	0·2	sulphides	0·2

A pyrrhotite-peridotite has been described by E. S. Bastin in Knox Co., Maine. It holds 60 per cent of olivine, 22·5 per cent of pyrrhotite, 10 per cent of magnetite, and small quantities of labradorite, hornblende, biotite, spinel and copper pyrites.

An unusual variety of peridotite or dunite is found in the Transvaal, in addition to the ordinary magnesian kind. This rock is dark brown in colour and remarkably dense, owing to the high iron-content of the olivine which has approximately the composition Fo_1Fa_1. P. A. Wagner gives the following mode:—

olivine (Fo_1Fa_1)	93·2	hornblende	1·5
diopside	5·0	ore	0·3

Perhaps the commonest type of peridotite is that which contains ortho-pyroxene along with olivine; the name "harzburgite" is given to such rocks. Harzburgite forms a facies of nearly every important body of olivine rocks, such as Dun Mountain, New Zealand; the Great Dyke of Rhodesia; the Serpentine Belt of New South Wales; and the peridotites of the Ural Mountains. A harzburgite from the Bushveld complex, Transvaal, has the following composition (P. A. Wagner):—plagioclase 5, diopside 4, bronzite 32, olivine 47, chromite 5, iron ore 7.

The Lherz type, or lherzolite, contains both ortho-pyroxene and diallage. W. E. Tröger gives the following data for the original lherzolite of the Pyrenees:—olivine 50, bronzite 32, diallage 15, ore 3.

Peralkaline Type

Two examples of melanite-pyroxenite may be quoted here. The first occurs in Assynt, Scotland, and was described under the name of "cromaltite" (S. J. S., 2); the second is a marginal facies of the Ice River alkaline complex in British Columbia (J. A. Allan). The modes are :—

	(1)	(2)
soda-diopside	51·9	65
melanite	15·6	5
biotite	16·5	—
sphene	—	15
ore	5·2	5
apatite	4·2	10
rest	6·6	

K.—Melilite-Basalt, Kimberlite. DVH

Dyscrystalline rocks corresponding to peridotite are little known; the nearest are the melilite-basalts and the dyke-serpentines. The melilite-basalts are dark, heavy rocks like fine-grained basalt, usually with a shiny surface due to abundant grains of ilmenite. Larger crystals of ilmenite and olivine may give the rock a porphyritic appearance, and in the variety called alnöite there are conspicuous plates of brown mica. Melilite is confined to the groundmass, along with augite and perovskite. Partial modes of some South African melilite-basalts have been furnished by M. S. Taljaard, as follows :—

	(1)	(2)	(3)
Olivine	28·5	32·5	31·3
Augite	26·0	30·5	27·4
Melilite	13·2	14·6	15·2
Perovskite	4·0	1·0	5·2
Iron ore	9·2	15·2	14·6
Irresolvable	19·0	6·2	6·3

The alnöite of Alnö, Sweden, is a micaceous type of melilite-basalt which has many insets of biotite, together with titanomagnetite, pyroxene, olivine and apatite, in a groundmass of mica, melilite and calcite. Perovskite, chromite, pyrrhotite and melanite are present in some cases (A. C. Högbom). The rock is very variable in composition, but Tröger gives the mode of one example as follows :—melilite 33, biotite 30, augite 17, olivine 5, ore, perovskite, carbonates, etc., 15.

The alnöite of Ile Cadieux, Quebec, contains the same minerals as the Swedish alnöite, with the addition of scapolite and monticellite. N. L. Bowen has given the following mode :—

biotite	30 (by weight)	melilite	3
monticellite	25	perovskite, ore	7
olivine	15	carbonates	6
augite	10		

Kimberlite, the "blue ground" of the South African diamond fields, is usually described as a "serpentinized porphyritic peridotite," but it is clear that even in its freshest condition it was never a eucrystalline rock. It is a dark greenish-black rock which for the most part is obviously a breccia; it encloses angular lumps and fragments of shale, diabase, quartzite, gneiss and granite, also rounded boulders or nodules of pyroxenite and garnet-pyroxenite (griquaite). The groundmass in which these lie is a scarcely resolvable mixture of serpentine and carbonates, with grains of more or less completely serpentinized olivine, others of pyroxene, garnet, ilmenite, perovskite, scales of biotite, and amorphous decomposition-products. The smaller olivine crystals, the perovskite, and the chromite are idiomorphic, but the large olivines, the garnets and the pyroxenes are always rounded and have been deeply corroded by the magma. It is concluded that the pyroxenes and garnets, and perhaps part of the ilmenite and olivine, are foreign bodies that have been derived from the enclosures of pyroxenite and griquaite; and that the endogenous minerals of kimberlite are olivine, serpentine, biotite, ilmenite and perovskite (S. J. S., 4). Since these minerals are nearly free from lime, and all chemical analyses of kimberlite show 15 per cent or more of calcium carbonate, we must conclude that in an earlier stage of its history kimberlite contained some easily decomposed lime mineral. Considering the associated minerals, melilite is at once suggested. No fresh melilite has been identified in any South African kimberlite, but C. H. Smyth was able to identify it in one of the dyke-peridotites of New York State, which are the rocks that most closely resemble kimberlite. The conception that kimberlite is a highly altered melilite-basalt was first put forward by H. C. Lewis, and it has been supported in a recent study of the question by M. S. Taljaard.

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CHAPTER XXIII

THE UNDERSATURATED ROCKS (FELDSPATHOIDAL DIVISION)

(DESCRIPTION)

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A.—Mineralogy

The feldspar that occurs in this group of rocks includes every variety from the most potassic microcline, as in certain foyaites of the Transvaal, to pure albite, as in the canadite of Ontario, and to almost pure anorthite as in some of the leucitic lavas of the Roman region. Nepheline and the other soda-feldspathoids are associated particularly with soda-orthoclase, anorthoclase and albite, and even in the group of nepheline-plagioclase rocks the feldspar is not usually more calcic than medium labradorite. Leucite may be found in conjunction with a purely potassic feldspar, as in the Leucite Hills of Wyoming, and sometimes with a more or less sodic orthoclase or sanidine, but not with albite or soda-rich plagioclase. When leucite occurs together with plagioclase, the latter is always a calcic one, from labradorite to anorthite. The explanation may lie partly in the relatively low temperature of soda-rich magmas and partly in the greater stability of orthoclase than of albite when silica is deficient.

When the proportion of nepheline in a rock is small, the mineral crystallizes late and occupies the interstices between the feldspar crystals; when it is abundant, nepheline forms beautifully idiomorphic crystals which may either be enclosed in large feldspar tables or separated from each one another by thin plates of feldspar. Sodalite too may crystallize early and appear as little dodecahedrons entirely enclosed in feldspar, or it may be interstitial, or again it may form extremely irregular, branching masses that seem to replace feldspar. Analcime too behaves in this contradictory way, sometimes forming euhedral insets, sometimes filling angular interstices between the feldspars, and sometimes replacing feldspar. Cancrinite, when it is abundant, always replaces nepheline; only in a few instances has it been recorded as a primary constituent of rocks.

Leucite is entirely restricted to lavas and near-surface rocks. The missourite of the Highwood Mountains, Montana, is often cited as a plutonic rock, but it occurs in a volcanic plug and is at most hypabyssal. The leucite-theralite described by Bowen and Ellestad, in the Lake Kivu region, is also regarded as a remnant of a volcanic plug. In most lavas leucite is a mineral of very early crystallization, yet in some of the lavas of the Leucite Hills, Wyoming, the only visible crystals are of diopside and mica, and the leucite is retained in the glass base.

Orthoclase melts incongruently at 1150° ± 20° (Schairer and Bowen), giving leucite and a siliceous liquid. If a lava rich in orthoclase began to crystallize above that temperature and were then chilled rapidly, it would give rise to a rock consisting of crystals of leucite in a glassy matrix containing an excess of silica. This abnormal condition is realized in some of the rocks of the Leucite Hills, and the same condition has been observed in the newly discovered leucite rocks of the West Kimberley area, Western Australia (Wade and Prider).

In soda-rich rocks the place of leucite is taken by "pseudo-leucite." This name is given to certain aggregates, having the shape of trapezohedral crystals such as those of leucite and analcime, which have been found especially in hypabyssal rocks of syenitic or foyaitic character. The chief constituent of the aggregates is orthoclase, with which there may be associated either nepheline, white mica or zeolites. To account for the generation of these aggregates it has been supposed that a somewhat sodic leucite has broken down according to the equation

$$\underset{\text{(leucite)}}{2(K,Na)AlSi_2O_6} = \underset{\text{(orthoclase)}}{KAlSi_3O_8} + \underset{\text{(nepheline)}}{NaAlSiO_4}$$

But Bowen and Ellestad have thrown doubt upon the existence of soda-rich leucite; they claim that potash in leucite is replaceable by soda only to a very limited degree, and that for the development of pseudo-leucite it is necessary for the leucite crystals to react with the residual liquid, which contributes the necessary soda. They have described a leucitophyre of central Africa in which the insets of normal potash-rich leucite (K_2O 18·3%, Na_2O 1·8%) have developed mantles of orthoclase and nepheline, thus clearly demonstrating the action of a soda-rich liquid upon the potash-rich crystals.

H. S. Washington (2) made the suggestion that "pseudo-leucite is not a pseudomorph after original leucite, but represents a distinct soda-potash mineral, as yet unknown, which was not stable under the conditions following its period of crystallization." Larsen and Buie have since found a potassic variety of analcime (K_2O 4½%) in dyke-rocks of the Highwood Mountains, Montana, in association with leucite rocks; both analcime and leucite are partly transformed into orthoclase and zeolites, hence Larsen and Buie conclude that pseudo-leucite may be formed "by the unmixing of either soda-rich leucite or potash-rich analcime."

A very remarkable example of "pseudo-leucite" was recorded by E. Hussak, from Serra de Caldas, Brazil. The pseudomorphs have a narrow mantle of orthoclase laths, within which they consist of orthoclase and nepheline with a little spinel, corundum and pyroxene. They lie in a matrix of orthoclase, quartz, and a little biotite or ægirine. Quartz is never found in the pseudomorphs or nepheline in the groundmass. In another rock from the same locality Hussak found both large pseudo-leucites of the usual sort and microscopic crystals of analcime; he concluded that both are pseudomorphs after leucite, but that the large pseudomorphs were formed under intratelluric conditions and the microscopic ones by the action of water after crystallization had been completed. The pseudoleucites of Brazil have lately been described and illustrated by D. Guimaraes.

Another type of "pseudo-leucite" consisting of orthoclase and muscovite has been described by A. Zavarickij in alkaline rocks of the Ishim River, West Siberia. The following reaction is suggested :—

$$\underset{\text{(leucite)}}{6KAlSi_2O_6} + CO_2 + H_2O = \underset{\text{(orthoclase)}}{3KAlSi_3O_8} + \underset{\text{(muscovite)}}{KH_2Al_3Si_3O_{12}} + K_2CO_3$$

The *dark minerals* of the feldspathoidal rocks include both saturated and unsaturated species, hence there are few exclusions. Muscovite is scarce, but it occurs along with corundum and biotite in the nepheline rocks of Canada. Biotite is the principal dark mineral in the Canadian occurrences and it is important in the Ural Mountains and Madras, but in the larger foyaite bodies biotite is quite unimportant and frequently absent. Even tourmaline is a constituent of certain very coarse nepheline-pegmatites on the York River, Ontario.

Many varieties of amphibole occur in these rocks, especially the sodic amphiboles arfvedsonite, riebeckite and hastingsite, the dark brown barkevikite, and the related ænigmatite or cossyrite. These soda-amphiboles have often been formed at the expense of earlier pyroxenes, and one may observe crystals of diopside or ægirine partly transformed into blue amphibole. But although individual occurrences of amphibole are common enough, the characteristic mineral of the feldspathoidal rocks is not amphibole but pyroxene. This ranges in composition from pure ægirine through more and less sodic varieties of diopside and hedenbergite, or diopside with sodic mantles, to common augite and deep purple titanaugite. On the whole the more sodic pyroxenes are associated with nepheline, and leucite with the less sodic varieties. Titanaugite may show the unusual character of an excess of calcium atoms over magnesium atoms. The type of pyroxene in which magnesium exceeds calcium (the enstatite-augite series) is characteristically absent from feldspathoidal rocks; invariably so from nepheline-bearing rocks and usually so from leucite rocks. No ortho-pyroxene has ever been recorded in a feldspathoidal rock.

Olivine of the usual basaltic character is common, especially in the more melanic lavas. Melilite is scarce, but it is fairly often recognized in leucite rocks of low silica content. The lime-garnet, melanite or schorlomite, is a not uncommon associate of both nepheline and leucite, especially in regions where limestone occurs. In deep-seated rocks it is a mineral of late crystallization, seldom showing its characteristic crystal form. In the original ijolite of Finland, and in rocks of the same character in Sweden and South Africa, melanite has crystallized after every other mineral except cancrinite, and in the melanite rocks of north-west Scotland it has crystallized after pyroxene and usually after biotite. Yet in certain effusive rocks—trachytes, phonolites, nephelinites and leucitites—melanite appears in beautifully idiomorphic crystals and seems to have been one of the earliest minerals to crystallize.

Sphene is unusually abundant in many nepheline rocks, as at Ice River, B.C.; Beemerville, New Jersey; and Sekukuniland, Transvaal. Certain foyaites at Umptek, Kola, hold nearly 50 per cent of this mineral. Apatite is another common accessory mineral that reaches an extraordinary concentration in many nepheline rocks, the outstanding locality being Umptek, where enormous bodies of nepheline-apatite rock are exploited commercially. Concentration on a smaller scale is conspicuous in Alnö and Sekukuniland.

Many rare species containing especially titania and zirconia are found in nepheline rocks. Among the titanosilicates there are astrophyllite, lamprophyllite (molengraaffite), mosandrite, besides perovskite and knopite. The zirconosilicates include eucolite (eudialyte), catapleiite, rosenbuschite, wöhlerite, lavenite.

B.—CLASSIFICATION AND NOMENCLATURE

With a large number of light minerals present, the rational subdivision of this group becomes a difficult matter. Even if all the soda-feldspathoids (nepheline, sodalite, haüyne, nosean, cancrinite and analcime) are grouped together in opposition to the potash-feldspathoid (leucite), that leaves us still with two kinds of feldspathoid in addition to the usual three kinds of feldspar. As regards the feldspathoids, a sharp line of division can be drawn between rocks which contain leucite and those which do not; and since the inversion point of orthoclase-leucite falls within the range of magmatic temperatures, it is important that such a line should be drawn; but no such precise separation can be effected with reference to the kind of feldspar present. If we continue our former practice of comparing the ratio of orthoclase to anorthite molecules in the feldspar, the results will not be comparable with those reached in the earlier classes, for in many lavas a large part of the potash is present in leucite instead of orthoclase. It thus becomes necessary to replace the orthoclase-anorthite ratio by the ratio of *orthoclase + leucite* to anorthite. This gives us the following primary classification:—

Group I.—$Or + Le > An$	(a) Leucite present (Symbol α)
	(b) Leucite absent (Symbol β)
Group II.—$Or + Le < An$	(a) Leucite present (Symbol γ)
	(b) Leucite absent (Symbol δ)

The symbols α, β, γ, δ, now have a slightly different meaning from that which was attached to them in the more feldspathic classes, but α still indicates a dominantly potassic character, β a dominantly sodic character and δ a dominantly calcic character in the light minerals.

For the first group, when eucrystalline, the general name of *syenoid* will be used. This name is derived quite legitimately from syenite, just as feldspathoid is derived from feldspar; but petrographers who tolerate so many uglier and less justifiable rock-names have not made use of it. The writer finds it useful and inoffensive. For those who dislike it the alternative is "feldspathoid-syenite." For Group II the corresponding term would be either *dioroid* or *gabbroid* according to the colour index.

For lavas of Group I the name *leucite-phonolite* is available. The typical phonolite, however, is a leucocratic rock, and many members of this group are mesotype or melanocratic. With increasing colour index it is the content of alkali-feldspar that diminishes rather than the content of leucite, so we may find the melanic equivalents of leucite-phonolite among the *leucitites* and *leucite-tephrites* (the former almost free from feldspar, the latter containing plagioclase) or the corresponding olivine-bearing types, *leucite-basalt* and *leucite-basanite*. There is no single name which could be used to cover all these melanic leucite-phonolites.

As regards Group IIa, the number of rocks known to fall in this group is so small that we can dispense with group names for the present.

C.—LEUCITE-SYENITE AND RELATED ROCKS. XUa, XWa

The only known example of an intrusive eucrystalline rock containing unaltered leucite, if we exclude ejected blocks like those in the agglomerates of the Alban Hills and Monte Somma, is the "missourite" of the Highwood Mountains, Montana. This is a subaluminous to metaluminous rock for which Weed and Pirsson gave the following mode:—

leucite	16 (weight)	augite	50
analcime	4	olivine	15
zeolites	4	ore	5
biotite	6	Colour index	(76)

The "Italite" and "missourite" described by H. S. Washington (3) occur in the form of ejected blocks in agglomerate at Villa Senni, Alban Hills. The largest block measured only 60 cm. and there is no certainty that any body of rock of similar composition underlies the volcano; it seems possible, from the description, that these blocks are merely facies of the normal leucitite of the Alban volcano, which have been locally enriched with leucite by flotation or otherwise. The following modes were given by Washington.

	Italite	*Missourite*
leucite	93·5	49
haüyne	1·5	2
biotite	·5	4
melanite	·5	·5
augite	3·0	43
magnetite, apatite	1·0	1·5
Colour index	(5)	(49)

Peralkaline Type

A holocrystalline porphyritic rock containing large trapezohedral pseudomorphs, supposed to be derived from leucite, was described by J. F. Williams at Magnet Cove, Arkansas. The pseudomorphs consist of orthoclase, nepheline and cancrinite; the groundmass holds nepheline, orthoclase, pyroxene, garnet and biotite. Assuming that the pseudomorphs were originally leucite, H. S. Washington (1) works out the composition of the rock as follows:—

orthoclase	3·9 (weight)	diopside and ægirine	19·2
leucite	36·9	melanite	14·5
nepheline	25·5	Colour index	(34)

L. V. Pirsson gave the name "fergusite" to a rock in the Highwood Mountains which contains 65 per cent of pseudoleucite, 24 of soda-diopside, 8 of ore, and trifling amounts of olivine, biotite and apatite; colour index 35. Another occurrence of fergusite has since been described by V. Nikolajew in Turkestan. Two modes are given, as follows:—

orthoclase	45·2	37·0
nepheline	8·8	16·9
orthoclase and nepheline intergrowths	9·7	7·0
sodic diopside	23·5	12·0
biotite	6·0	12·8
melanite	4·1	10·8
rest	2·7	3·5
Colour index	(36)	(39)

D.—LEUCITE-PHONOLITE AND RELATED ROCKS. DUa, DWa

No peraluminous rocks are known in this group.

Metaluminous Type

The first three illustrations relate to rocks from the volcanic region about Rome; they have been described by H. S. Washington (4) under the name of leucite-trachyte. In our terminology they are leucite-phonolites.

	(1)	(2)	(3)
soda-orthoclase	39·6	47·7	30·0
labradorite	17·5	16·3	24·5
leucite	31·4	16·9	15·7
nepheline	—	—	0·9
biotite	2·3	3·7	2·3
augite	5·3	10·3	18·3
magnetite, apatite	3·9	4·3	3·4
Colour index	(11)	(18)	(27)

The next two rocks are also from the Roman region; they carry more plagioclase than orthoclase, for which reason Washington (4) has called them leucite-tephrite. They fall just on the boundary between the leucocratic phonolites and the mesotype tephrites.

soda-orthoclase (Or_2Ab_1)	26·0	(Or_3Ab_1)	10·5 (weight)
labradorite (Ab_2An_3)	28·0	(Ab_1An_1)	26·1
leucite	15·8		29·0
nepheline	1·0		2·8
biotite, hornblende	3·4		2·0
augite	16·8		21·6
olivine	2·0		5·0
magnetite, apatite	7·0		3·0
Colour index	(29)		(31)

In the curious rock which W. Cross has named "madupite," from the Leucite Hills of Wyoming, no leucite is developed at all, and the rock consists of diopside, phlogopite and a little perovskite in a glassy base which has the chemical composition of leucite. If the magma had crystallized completely, this rock would have been composed of:

leucite	20·3	diopside	46·1
nosean	6·5	accessories	8·2
phlogopite	18·9	Colour index	(73)

To the metaluminous type we must add the melilite-bearing leucitites, for which H. S. Washington (4) has proposed to revive the old name *cecilite*. The composition, in a simple case, was leucite 60, augite 20, melilite 18, magnetite 2; colour index 40. In other instances described by Washington there are small amounts of nepheline, olivine and anorthite in addition. In a "nepheline-melilite rock" from the Alban Hills Washington (3) found leucite 12, nepheline 40, melilite 40, haüyne 1, melanite 4, augite 2, apatite 1; colour index 47.

A melanocratic rock of this character has been described by A. Holmes (2) under the name "katungite." The calculated mineral composition is leucite 16, melilite 40, olivine 22, perovskite 7, magnetite 6, nepheline and rest 9; colour index 75.

Subaluminous Type

The following examples are leucite phonolites from the Roman region:—

	(1)	(2)	(3)
soda-orthoclase	60·7	55·2	22·2
labradorite	11·3	20·1	35·6
leucite	8·8	1·7	25·5
nepheline	3·5	4·0	—
augite	8·9	13·1	9·1
olivine	1·1	1·7	2·3
magnetite, apatite	5·7	4·2	5·3
Colour index	(16)	(19)	(17)

Some of the lavas of the Roman region are largely composed of glass. The apparent composition of two such rocks is given below, and it is followed in each case by the calculated composition, obtained by chemical analysis.

	(observed)	*(calculated)*	*(observed)*	*(calculated)*
soda-orthoclase	1·1	48·3	1·1	47·8
labradorite	2·9	18·3	14·1	19·9
leucite	16·6	16·8	38·1	17·2
nepheline	—	6·5	—	—
augite	4·9	6·2	1·4	8·2
ore	1·4	3·9	—	5·4
groundmass	73·1	—	45·3	—
Colour index		(10)		(14)

The following examples will serve to show the range of variation among the leucite-tephrites and leucite-basanites of the Roman region. All are described by Washington (4).

soda-orthoclase	8·5	—	—	—
labradorite-anorthite	13·9	14·1	10·5	13·1
leucite	36·2	35·0	32·3	17·3
nepheline	8·3	5·1	3·4	3·4
augite	30·4	41·0	47·1	48·0
olivine	—	4·0	4·0	15·9
magnetite, apatite	5·7	0·8	2·1	2·3
Colour index	(36)	(46)	(53)	(66)

A. Rittmann has given a number of modes of Vesuvian lavas which are entirely comparable with those cited above.

Peralkaline Type

For the pyroxene in a leucite rock to be a sodic one is distinctly uncommon, but Washington (4) gives two examples from the Roman region in which the usual augite is replaced by a bright green "ægirine-augite" (soda-diopside),

soda-orthoclase	68·5 (weight)	19·2
labradorite	—	7·0
leucite	13·4	37·0
nepheline	4·7	11·7
haüyne	1·0	10·0
soda-diopside	7·8	11·5
melanite	—	2·0
biotite	—	1·3
ore, apatite	4·6	0·3
Colour index	(12)	(15)

Washington described the first rock as leucite-phonolite, the second as leucite-tephrite, but both rocks are strongly alkaline in character and both should be called leucite-phonolite.

The leucite-phonolite of Perlerkopf, in the Eifel district of Germany, contains leucite 30, nepheline 30, soda-diopside 32, melanite and sphene 8; colour index 40 (R. Brauns).

E.—Nepheline-Syenites and Related Rocks. $xu\beta$, $xw\beta$

These rocks contain a soda-feldspathoid which is generally nepheline but sometimes sodalite, cancrinite or analcime, without any leucite. In this group the tendency to sub-divide and to create specific names has been carried to extreme lengths, over forty names being in use for the eucrystalline varieties alone. Most of these names are unnecessary, and all lack quantitative definition.

It will be sufficient for our present purpose to divide the syenoids into two series, one in which alkali-feldspar is abundant and another in which feldspathoids greatly exceed or entirely replace feldspar. The characteristic members of the first series are leucocratic, and from the many names in use we select *foyaite* (derived from the typical locality at Foya, Portugal) as a generic name. For mesotype varieties of this series the name *malignite* (from Maligne River, Ontario) is suitable. Most examples of the second series are mesotype, and from their occurrence on a large scale at Ijo, Finland, they are known as *ijolite*, while the less common leucocratic varieties have received the name of *urtite* (Lujaur-Urt, Kola). When we consider the melanocratic syenoids, the separation of the more feldspathic from the less feldspathic varieties ceases to have much significance, for the light minerals form the minor part of the rock; the name *jacupirangite* (from Jacupiranga, Brazil) is therefore suitable for all such rocks, no matter what the proportion of feldspar to feldspathoid may be.

Peraluminous Type

The soda-syenoids of Ontario are mostly peraluminous, and some of them contain corundum. The feldspar is always a very sodic variety, partly albite and partly soda-orthoclase or microperthite. P. D. Quensel has proposed to give the name "canadite" to all albite-rich foyaites, but since there is an excess of albite molecules over orthoclase molecules in nearly all foyaites, the distinction is not a very significant one. (It would be much more useful to make canadite a generic name for all peraluminous foyaites, to distinguish them from the commoner peralkaline type.) Two analyses of the foyaites of Ontario are given below, after Adams and Barlow:—

	(1)		(2)
orthoclase	16·1	(weight)	4·4
albite	53·4		50·8
anorthite	1·3		1·2
nepheline	18·2		8·0
muscovite	8·0		—
biotite	1·3		29·6
ore, apatite, calcite	0·9		4·4
Colour index	(10)		(31)

Others of the Ontario syenoids have nepheline in large excess over feldspar, and belong to the urtite division. The first of the following examples is from Craigmont (Adams and Barlow), the second from Monmouth (W. G. Foye):—

	(1)		(2)
albite	—		9·2
oligoclase	29·7	(weight)	—
nepheline	63·2		56·7
scapolite	—		10·2
muscovite	4·4		—
biotite	0·5		22·5
corundum	0·5		—
rest	1·5		1·4
Colour index	(7)		(23)

The foyaites of Litchfield, Maine, and Salem, Massachusetts, are remarkably similar in composition, according to data furnished by W. S. Bayley and H. S. Washington respectively. The figures are

	Litchfield	*Salem*
orthoclase	27·0	27
albite	46·9	43
nepheline	17·0	20
cancrinite	2·0	1
biotite	6·9	6
accessory	—	3
Colour index	(7)	(9)

The peraluminous type of foyaite is also found at Ditró, Hungary, and Miask, Ural Mountains, along with metaluminous and peralkaline types; but quantitative information is lacking as regards both the bulk and the composition of these types.

Metaluminous Type

The "monmouthite" of Monmouth, Ontario, is a hornblende-syenoid; it is much rarer than the mica-syenoids of this region. The hornblende is a dark green variety with peculiar optical properties, called hastingsite; it carries about five per cent of potash and soda, which raises the question whether this rock ought not to be referred to the peralkaline type. The fact is that hastingsite and barkevikite are intermediate between common aluminous hornblende and the real soda-amphiboles, arfvedsonite, riebeckite and crocidolite, which have seven to twelve per cent of alkalies and little or no alumina. Taking both alkalies and alumina into consideration, it seems that hastingsite and barkevikite are nearer to common hornblende (which nearly always holds two or three per cent of alkalies) than to the soda-amphiboles.

The mode of typical "monmouthite" is as follows (Adams and Barlow):—

albite	1·8 (weight)	calcite	3·1
nepheline	72·2	hornblende	15·1
sodalite	0·3	hematite, etc.	0·6
cancrinite	5·1	Colour index	(16)

L. V. Pirsson and H. S. Washington describe hornblende-foyaite at Red Hill, New Hampshire. The hornblende is called catophorite but it is evidently an aluminous hornblende with no excess of alkali, for acmite does not appear when the norm is calculated. It is stated that ægirine is sometimes present, but the modes and chemical analyses that are furnished are those of metaluminous rather than peralkaline rocks. The following modes are given :—

	(1)	(2)
orthoclase	36	32
albite	37	43
nepheline	14	7
sodalite	5	2
hornblende	7	12
accessory	1	4
Colour index	(8)	(16)

Coming now to rocks with a higher colour index, we include a malignite from Monmouth (Adams and Barlow), and a hornblende-jacupirangite from Tamaulipas, Mexico (G. I. Finlay).

	Monmouth	*Tamaulipas*
orthoclase	2·8 (weight)	12
albite	22·3	—
anorthite	1·7	—
nepheline	26·2	18
hornblende	39·7	40
titanaugite	—	23
magnetite, etc.	5·8	5·9
Colour index	(45)	(69)

A few syenoids contain a lime-iron garnet without the soda-pyroxene that usually accompanies it; they are therefore placed in this group. The "borolanite" of J. Horne and J. J. H. Teall, from Assynt, Scotland, is a melanite-biotite syenoid which typically holds no pyroxene at all. The rock exhibits white spots, of rounded or rudely polygonal outline, which may be "pseudo-leucite," set in a base of alkali-feldspar, melanite and biotite. The white spots consist of feldspar and zeolites, and the mode of the rock is as follows (S. J. S., 1) :—

feldspar, etc., of white spots	33·7 (volume)
feldspar, etc., of groundmass	32·6
melanite, sphene	25·5
biotite	8·2
Colour index	(34)

Melanite-biotite foyaite has been described in the Kruger Mountain area, British Columbia, by R. A. Daly (1), who gives the following mode :—

soda-potash-feldspar	63·9 (volume)	melanite	8·8
nepheline	15·1	apatite	0·6
biotite	11·1	sphene, ore	0·5
		Colour index	(21)

The "okaite" described by J. Stansfield, from the Oka Hills, Quebec, is one of the rare deep-seated rocks that contain melilite. The following mode is given :—

haüyne	21	magnetite	7·5
melilite	52·5	pyrites	0·5
biotite	8	apatite	2·5
calcite	2	Colour index	(77)
perovskite	6		

The similarity of this okaite, which occurs at a limestone contact, with the "turjaite" of the Kola Peninsula is sufficiently pronounced to suggest that the latter too owes its abnormal composition to assimilation of limestone. The following mode is given by E. H. Kranck.

nepheline	16·4	magnetite	7·2
melilite	41·7	apatite	3·7
biotite	18·3	calcite	1·8
melanite	2·3	Colour index	(83)
perovskite	4·1		

Subaluminous Type

Strictly subaluminous rocks are rare in this class, for the pyroxene tends either to hold a little of the ægirine molecule in solid solution or else to be mantled with zones of more and more sodic pyroxene. Many foyaitic rocks are said to contain "augite" or "diopside," but in nearly all cases the mineral proves to be more or less sodic in character. This is true of the augite in the foyaite of Tamaulipas, Mexico, and of the diopside in the ijolite of Kuolarjärvi, Finland, two rocks which were cited as examples of the subaluminous type in the first edition of this book. These rocks are transitional between the subaluminous and the peralkaline type. In many instances the formation of sodic mantles about common pyroxene is not a magmatic phenomenon at all but a deuteric reaction. R. H. Jahns has shown that this process, in an analcime-syenite of South Park, Colorado, was contemporaneous with analcitization of the feldspar.

Peralkaline Type

The great majority of syenoids, and many phonolites, are of this type. Three of the greatest masses of nepheline rocks in the world (Kola, Julianehaab, and Pilansberg) consist almost exclusively of peralkaline types; and other large syenoid intrusions such as those of Serra de Monchique, Arkansas, Cabo Frio, Serra de Tinguá, and Los Islands are mainly peralkaline.

Foyaites with only small quantities of nepheline or sodalite, which are near syenite in composition, are sometimes called "pulaskite" from their occurrence in Pulaski County, Arkansas. H. S. Washington (1) gives the following modes computed from analyses of the pulaskite of Fourche Mountain:—

	(1)	(2)
orthoclase	28·6	35·1
albite	39·0	39·8
nepheline	6·2	3·1
sodalite, nosean	1·2	1·2
ægirine	6·9	5·1
biotite, hornblende, diopside	7·9	11·0
rest	10·2	4·7
Colour index	(25)	(21)

The "hedrumite" of W. C. Brögger (1), from Hedrum, Norway, is not in any essential respect different from pulaskite, and the "sölvsbergite" of Tjose-Aklungen is a fine-grained dyke rock with trachytic texture, consisting of alkaline feldspar 71·5, nepheline 5, soda-pyroxene 15·2, biotite 7·1, melanite 0·7 and perovskite 0·3 per cent.

The "lardalite" of Brögger (2) is only distinguished from other foyaites by the peculiar rhombic habit of the feldspar (anorthoclase) and by the frequent appearance of olivine. The mode is as follows:—

orthoclase	19·0	62·3	soda-pyroxene	8·3
albite	40·6		biotite	9·5
anorthite	2·7		ore	2·8
nepheline	13·0		apatite, zircon	1·7
sodalite	2·0		Colour index	(22)

Among the commoner examples of foyaite there is a distinct tendency towards what might be called a standard composition. The writer observed that five of the most abundant foyaites of South Africa have nearly the same chemical composition, which expresses itself in terms of minerals as follows :—

soda-orthoclase (Or 32, Ab 29)	61
nepheline	23
soda-diopside (Di 5·7, Ac 4·6)	10·3
sphene, magnetite	2·4
rest (including combined water)	3·3

or in round numbers feldspar 62, nepheline 23, melanes 15. B. M. Kupletsky and T. M. Oklova have since computed the average composition of 182 nepheline-syenites and find it to be feldspar 60·4, nepheline 20, melanes and secondary minerals 19·6.

The name "lujaurite" was given by W. Ramsay to foyaites of Lujaur-Urt, in the Kola peninsula, which are characterized by the presence of abundant needles of ægirine arranged in parallel position, giving the rock the appearance and texture of a crystalline schist. This curious rock is actually one of the most important varieties of foyaite, for it forms a large part of Lujaur-Urt, it is an important element in the Ilimausak intrusion in Greenland, it is seen again in the "green foyaite" of Pilansberg, and it forms the greater part of the Sekukuniland stock. Rare minerals, especially eudialyte and lamprophyllite, are remarkably common in these rocks. Most lujaurite contains 40 to 50 per cent of feldspar, 22 to 27 of nepheline, and 23 to 30 of ægirine or soda-rich diopside, with a variable and often large proportion of eudialyte. The mode of the Sekukuniland lujaurite was computed as follows (S. J. S., 2) :—

orthoclase, albite and microperthite	44
nepheline	27
acmitic pyroxene	24
sphene, ore, apatite, combined water	5
Colour index	(27)

Among the less common varieties of foyaite we may notice the "mariupolite" of J. Morozewicz, which is extremely rich in albite, and the "juvite" of W. C. Brögger (3), which is an orthoclase-rich variety. The modes are

orthoclase (Or_3Ab_1)	—	51
albite	74	—
nepheline	13	36
sodic pyroxene	7·5	7
biotite	4	—
zircon, sphene, etc.	2	6
Colour index	(13)	(13)

Sodalite is an abundant constituent of many foyaites, from 20 to 50 per cent having been recorded in some of the sodalite-foyaites of Greenland (N. V. Ussing). In the sodalite-foyaite of St. Hilaire Mountain, Quebec, J. J. O'Neill has recorded as much as 70 per cent of this mineral.

Cancrinite replaces nepheline to a large extent in some foyaites and ijolites, especially in the neighbourhood of limestone contacts. Ramsay and Nyholm described a rock at Kuolajärvi, Finland, which carries 29 per cent of cancrinite, along with nepheline, orthoclase, ægirine and accessories. The foyaite of Diamond Jo quarry, Magnet Cove, also holds much cancrinite (H. S. Washington, 1).

Analcime is another mineral which may accompany or replace nepheline in these rocks. G. W. Tyrrell (1) has described such rocks in the west of Scotland. One of them has the following composition:—alkali-feldspar 69, analcime 16, ægirine 9, alkali-hornblende 4, ore and apatite 4; colour index 17. Foyaite containing from 15 to 30 per cent of analcime occurs in the San Carlos Mountains, Mexico, but E. H. Watson reports that analcime was the last constituent to form and to a large extent replaces earlier crystals of orthoclase and nepheline.

As an example of a mesotype foyaite or *malignite*, we quote A. C. Lawson's analysis of the original malignite of Poohbah Lake (Maligne River), Ontario:—

orthoclase	20·9	soda-diopside	48·2
nepheline	19·6	sphene	1·0
biotite	2·0	apatite	5·8
		Colour index	(57)

A malignite described by R. A. Daly (1) at Kruger Mountain, British Columbia, contains much melanite. The mode is

soda-orthoclase, perthite	36·3 (volume)	melanite	9·5
nepheline	5·4	apatite	1·0
biotite	11·0	magnetite, sphene	0·3
soda - diopside	36·5	Colour index	(58)

The original *theralite* was discovered in the Crazy Mountains of Montana, by J. E. Wolff. Through an error in the first description of the rock, the name has been generally accepted as that of a nepheline-plagioclase rock or nepheline-gabbro, although the Crazy Mountains rock contains no plagioclase. Wolff has continued to use the name theralite for these rocks, and gives the following modes from which it will be seen that the Crazy Mountains theralite is essentially what we have called malignite, or mesotype foyaite.

orthoclase	23	21
nepheline	10	22
sodalite	21	11
sodic pyroxene	36	39
magnetite, apatite	4	1
Colour index	(40)	(40)

Olivine-bearing varieties are also described, in which the colour index may exceed 60.

The typical *ijolite* is feldspar-free and contains essentially only nepheline and a sodic pyroxene, with accessory melanite, sphene and apatite. The first ijolite was recognized and named at Ijo, in the Kuusamo district of Finland, by W. Ramsay and H. Berghell. It is compared below with the remarkably similar ijolite of Sekukuniland, Transvaal (2) and with an ijolite of Magnet Cove, Arkansas (3). The first analysis is by V. Hackmann, the second by the writer, and the third by H. S. Washington.

	(1)	(2)	(3)
nepheline	51·6	54·6	38·7
cancrinite	—	7·4	—
calcite	—	0·5	—
soda-pyroxene	39·2	32·2	42·8
sphene	2·1	0·5	—
melanite	0·7	—	15·3
apatite	4·3	4·8	3·0
Colour index	(46)	(38)	(61)

W. C. Brögger (3) has described beautiful examples of ijolite in the Fen district of Norway; they occur, as usual, in close proximity to limestone. Brögger has introduced the name "melteigite" for these rocks, but they are similar in every respect to the original Kuusamo ijolite and no new name is needed.

In view of the considerable theoretical importance of the ijolites in connection with the question of limestone assimilation, we may be allowed to cite two examples from the Kola peninsula. The first mode is by T. Brenner, the second by E. H. Kranck.

nepheline	38·2	44·2
soda-diopside	16·2	34·4
soda-hornblende	32·0	—
melanite	—	10·4
sphene, perovskite	5·8	1·2
ore, apatite, etc.	7·8	3·0
Colour index	(62)	(56)

When the proportion of dark minerals rises to two-thirds or more the name *jacupirangite* may be used. The rock to which this name was originally given by C. A. Derby, at Jacupiranga, Brazil, consisted of a more or less sodic pyroxene and a large proportion of titanomagnetite, with quite subordinate nepheline. Rocks of similar character have been found in almost all the foyaite-limestone localities that were discussed in Chapter XVIII, notably Alnö, Fen and Iron

Hill. Two modes from the latter locality are furnished by Larsen and Miller, as follows:—

nepheline	29	19
soda-diopside	38	77
garnet	25	—
apatite, ore, calcite	8	4
Colour index	(71)	(81)

The complementary type is *urtite*, named after Lujaur-Urt; it consists almost wholly of nepheline. W. Ramsay has given the composition of two of these rocks as follows:—

nepheline	86 (weight)	82
ægirine	12	16
apatite	2	2
Colour index	(14)	(18)

F.—Phonolite and Related Rocks. DUβ, DWβ

No phonolite of peraluminous type has been described.

Metaluminous Type

Quantitative descriptions of metaluminous *phonolite* are very scarce. The "vulsinite" of H. S. Washington (4), from Italy, is an augite-biotite phonolite. The composition of two examples of this rock is as follows:—

soda-orthoclase	65·2 (weight)	52·1
labradorite	14·3	17·8
nepheline	5·4	7·8
biotite	3·2	1·7
augite	8·3	17·7
magnetite, apatite	3·6	2·9
Colour index	(15)	(22)

The following highly melanocratic lavas are *nepheline basalts*. The first is from Fort Hall Reservation, Idaho, and it has been described by Mansfield and Larsen. The second was described by J. Gilluly in Utah.

	(1)	(2)
nepheline, haüyne	20 (weight)	18
biotite	8	20
pyroxene	39	40
olivine	26	20
rest	5	2
Colour index	(78)	(82)

The name *monchiquite* is given to a group of dyke-rocks which have an almost isotropic base composed largely of analcime. The following modes of monchiquite at Spanish Peak, Colorado, are given by A. Knopf.

biotite	20·6	6·7
titanaugite	19·4	31·7
olivine	—	9·3
magnetite	4·6	5·3
base (analcime, anorthoclase, etc.)	52·5	47·0
Colour index	(45)	(46)

A hornblende-monchiquite in the San Carlos Mountains, Mexico, contains augite 25, hornblende 25, with analcime and alteration products 50; colour index 50 (E. H. Watson).

The next two examples are melilite-nepheline basalts, one from Winnett, Montana (C. S. Ross), the other from Kassel, Germany (T. Ernst). These rocks correspond to jacupirangite in the eucrystalline division.

	Winnett	*Kassel*
nepheline (with haüyne and zeolites)	15	15
biotite	16	—
melilite	17	7
augite	5	34
olivine	35	30
perovskite	—	2
ore, apatite	11	10
Colour index	(80)	(85)

Subaluminous Type

Quantitative descriptions are again very scarce.

The phonolite of Leeuwkraal, Transvaal, is a subaluminous type. It contains (S. J. S., 3)

microperthite	75	diopside	10
nepheline	10	titanomagnetite	2
sodalite	10	sphene	1
		Colour index	(18)

An analcime-phonolite was described by T. Barth from the Society Islands in the Pacific. Anorthoclase makes up 82 per cent of the rock, and the analcime forms "small octagonal grains" in the groundmass, amounting to 7 per cent. Other constituents are diopside 5, ore 5, apatite 3; colour index 12.

A curious lava described by R. A. Daly (1) in British Columbia, under the name "shackanite" is an analcime-phonolite. It consists of

anorthclase	33·5	olivine	2·6
analcime	29·2	magnetite	2·0
biotite	0·2	apatite, sphene	2·0
augite	5·5	glass	25·0

A rock described by H. S. Washington (5) as "analcime-basalt," in Sardinia, falls in the mesotype section of this group. The mode is

crystals	analcime	10
	augite	25
	olivine	10
	magnetite	13
	apatite	2
glass	orthoclase	15
	albite	13
	anorthite	7
	analcime	5
Colour index		(50)

Some very melanic varieties appear among the nepheline-basalts. B. Koto has described a nepheline basalt at Yingemen, Manchuria, which holds 30 per cent (by volume) of nepheline, 15 of olivine, and apparently about 50 per cent of augite, since the only other constituent mentioned is a little glass. The colour index must consequently be near 65.

H. G. Backlund has given the mode of a nepheline basalt from General Vilkitzky Island, in the Arctic Ocean. It is

nepheline	32·1	magnetite	10·1
olivine augite	55·5	apatite, perovskite	2·3
		Colour index	(68)

Peralkaline Type

Although phonolites of this type are not uncommon, few have been described in quantitative terms. H. S. Washington (4) gives the modes of two phonolites from the Roman region, as follows:—

	(1)	(2)
soda-orthoclase	84·6 (weight)	82·7 (weight)
nepheline	3·7	5·4
sodalite, haüyne	3·9	2·1
soda-diopside	3·8	4·5
sphene, ore	4·0	4·8
Colour index	(8)	(9)

E. Lehmann (1) has calculated the composition of three phonolites from Ischia; all are so much alike that we may average them as follows:—

sanidine ...	43·3	feldspar 79·9	soda-pyroxene	7·8
albite	36·6		soda-amphibole	2·0
nepheline ...	8·2		rest	2·1
			Colour index	(12)

The "phonolitic trachytes" and phonolites that occur on oceanic islands are typically peralkaline, the melanes including ægirine and cossyrite. The phonolite of Stone Top, St. Helena, is a good example. According to the norm given by R. A. Daly (2) this rock contains

anorthoclase (Ab_2An_1)	72·5	ores	7·0
nepheline	8·5	apatite	2·4
soda-diopside	8·6	Colour index	(18)

An analcime-rich phonolite is described by A. MacGregor near Haddington, Scotland, containing

orthoclase	37·6	sodalite	2·7
albite	20·1	soda-diopside	5·7
anorthite	3·2	olivine	1·8
nepheline	4·2	ore, etc.	4·1
analcime	20·5	Colour index	(12)

An almost mesotype phonolite, in Mozambique, E. Africa, is described by A. Holmes (1). The mode is

anorthoclase	33·8	soda-amphiboles	15·8
nepheline, zeolites	31·3	ore	2·0
soda-pyroxene	15·1	calcite	2·0
		Colour index (33, weight)	

The name *tinguaite*, derived from the Serra de Tingua in Brazil, is given to a variety of phonolite which is characterized by an abundance of fine needles of ægirine which lie largely in parallel position and tend to be matted together. This rock corresponds to lujaurite in the eucrystalline division, and like lujaurite it frequently contains eudialyte and other rare minerals. The following analyses are of tinguaites from (1) Judith Mountains, Montana (Weed and Pirsson); (2) Serra de Monchique, Portugal (Kraatz-Koschlau and Hackmann); (3) Hedrum, Norway (W. C. Brögger); (4) Fen district, Norway (W. C. Brögger); (5) Bearpaw Mountains, Montana (Weed and Pirsson).

	(1)	(2)	(3)	(4)	(5)
orthoclase	54·6	20	20·5	25·3	44·4
albite	18·8	40	30	5·9	12·1
nepheline	8·8	20	30	33·6	9·3
sodalite, cancrinite	1·2	—	—	5·1	—
soda-pyroxene	12·8	8	15·2	28·6	34·2
biotite	—	3·5	2·5	—	—
ore, etc.	—	3	—	1·3	—
Colour index ...	(13)	(15)	(18)	(29)	(34)

G.—Leucite-Theralite and Related Rocks

XUγ, XWγ, DUγ, DWγ

Very few rocks satisfy the definition of this group. Among the ejected blocks of Monte Somma, A. Lacroix recognized a leucite-theralite to which he gave the name "puglianite." The mode is stated as follows by W. E. Tröger :—

sanidine	4	hornblende	2
leucite	5	augite	66
anorthite	16	Colour index	(76)
biotite	8		

Bowen and Ellestad have described a leucite-theralite from the crater of Nyamlagira in central Africa. This rock contains abundant insets of titanaugite, and some of olivine and titanomagnetite, in a groundmass of plagioclase, titanaugite and leucite. The norm indicates the following approximate composition

leucite	12	olivine	14
nepheline	6	ore	12
plagioclase (Ab_1An_2)	27	apatite	1
diopside	27	Colour index	(54)

It is likely that the proportion of anorthite indicated by the norm is too high, but there would still appear to be an excess of anorthite over leucite.

The two modes that follow relate to unusual varieties of leucite-basalt, the first from Bohemia (mode by W. E. Tröger, 2), and the second a rock from Uganda, described and named "ugandite" by A. Holmes (3).

plagioclase	albite	7	9·3
	anorthite	15	14·9
leucite		2	8·0
nepheline		—	2·0
biotite		1	—
augite		58	20·6
olivine		10·5	35·6
ore, apatite		6·5	9·0
Colour index		(76)	(65)

These very uncommon rocks, which constitute rock facies rather than rock masses, can be adequately explained by the gravitational accumulation of augite and olivine crystals, and the flotation of leucite.

H.—THERALITE AND RELATED ROCKS. xuδ, xwδ

Two names, essexite and theralite, are currently applied to eucrystalline rocks containing plagioclase and feldspathoid, but both have been vitiated by undiscriminating use: The name essexite was first given by J. H. Sears (1891) to a rock occurring on Salem Neck, Massachusetts, which contains all the minerals of a gabbro with the addition of some orthoclase, analcime and nepheline. Thirty years later C. H. Clapp showed that this rock is just a facies of the Salem gabbro which has suffered thermal metamorphism and alkali-metasomatism by the Beverley syenite-foyaite magma, whereby the plagioclase of the gabbro was largely analcitized and a little nepheline was introduced. Meanwhile, J. H. Hibsch had extended the name essexite to a rock occurring at Rongstock, Bohemia, which is essentially an orthoclase-gabbro holding only a trifling amount of feldspathoid. The consequence is that some petrographers now use the name essexite in the sense of orthoclase-gabbro and others in the sense of nepheline-gabbro or analcime-gabbro. The essexites described by W. C. Brögger in Norway are free from feldspathoids, but Rosenbusch included both feldspathoid-free and feldspathoid-bearing rocks in his essexite family.

An equally unfortunate history attaches to the name theralite, which was first given by J. E. Wolff to a foyaitic rock of the Crazy Mts., Montana, and was later applied by H. Rosenbusch to a gabbroic rock at Duppau, Bohemia. It is clear that the theralite of the Crazy Mts. does not belong to this group, and we have already discussed it among the melanic foyaites (p. 465). Custom has now established the Duppau rock as the typical theralite, but has failed to distinguish it clearly from the essexite of Rongstock, from which it seems to differ mainly by having more feldspathoid and less plagioclase.

In this chapter we are concerned only with undersaturated rocks, and it seems to the writer that the theralite of Duppau is the best representative of the nepheline-plagioclase group. We shall therefore avoid the name essexite and use *theralite* as the general name of this group. The number of rocks that fall into the group is so small that it is not necessary to subdivide them on the basis of colour-index.

Peraluminous Type

Nepheline-plagioclase rocks with excess of alumina have been described in Ontario under the names of dungannonite, raglanite, and craigmontite (Adams and Barlow). The composition of craigmontite is: nepheline 63, oligoclase 30, muscovite 4, corundum, etc.; colour index 7.

Metaluminous Type

The theralite of Duppau has the following composition (Tröger):—

soda-orthoclase	12	titanaugite	33
plagioclase (about An_{60})	16	olivine, biotite	5
nepheline	15	ore, apatite	7
hornblende	12	Colour index	(57)

A theralite from Lugar, Scotland, is very close to the Duppau rock in composition. G. W. Tyrrell (1) gives the following mode:—

labradorite	16·4	augite	35·9
nepheline	16·6	olivine	8·7
biotite	6·7	ore, apatite	3·5
hornblende	12·2	Colour index	(67)

A "nepheline-monzonite" from Madagascar, described by A. Lacroix, is another example of this type of rock. The mode given by Tröger is

anorthoclase	25	barkevikite and titanaugite	25
plagioclase	26		
nepheline	17	ore and apatite	7
		Colour index	(32)

The name "essexite" has been applied to a wide variety of rocks in the Monteregian Hills, Quebec, but few of these rocks contain any nepheline. Bancroft and Howard observed that the essexite of Mount Royal develops a little nepheline in the vicinity of a limestone contact. A small proportion of nepheline was recorded in the essexite of St. Hilaire Mountain, which locally develops a theralitic facies with the following composition: nepheline 29·3, plagioclase 55·9, hornblende 3·6, pyroxene 7·5, rest 3·6; colour index 15 (J. J. O'Neill).

G. W. Tyrrell (1) has described under the name "lugarite" certain rocks from the west of Scotland which resemble the original (Salem) essexite in holding a large proportion of analcime. The following modes are given:—

	(1)	(2)
plagioclase	10·5	14·6
nepheline analcime	42·5	49·0
augite	21·7	—
hornblende	17·2	29·5
ore	5·0	2·7
apatite	3·1	4·2
Colour index	(47)	(36)

A hornblende-rich theralite from Ditró, Rumania, has been described by A. Streckeisen, who gives the following mode:—andesine 36, nepheline and cancrinite 7, biotite 8, hornblende 44, accessories 5; colour index 57. This rock has been named "orotvite." It contains the very same proportins of hornblende and nepheline as the "berondrite" of Madagascar, the only mineralogical difference between them being that the latter-rock contains some titanaugite in addition to hornblende.

The "bekinkinite" of Madagascar is an analcime-rich theralite in which the analcime has been formed at the expense of earlier nepheline (Lacroix). The colour index is over 80.

Subaluminous Type

The rock commonly called essexite at Crawfordjohn, Scotland, is a subaluminous theralite according to our definition. The following modes of the normal rock and of an augite-rich facies have been given by A. Scott.

orthoclase and labradorite ...	32·2	30·9
nepheline and analcime	13·0	9·8
augite	27·4	41·0
olivine	23·8	13·8
rest	3·6	4·5
Colour index	(55)	(59)

The name "kylite" was given by G. W. Tyrrell (1) to an olivine-rich theralite occurring in the Kyle district of Argyll, Scotland. The average of four modes is stated below.

labradorite	27·3	olivine	37·7
nepheline, analcime ...	5·1	rest	5·3
titanaugite	24·6	Colour index	(67)

An olivine-theralite from the Austral Islands, in the Pacific Ocean, which was described by W. Campbell Smith, closely resembles these Scottish examples. The mode is:—plagioclase 19, nepheline 12, augite 36·5, olivine 24·5, accessories 7·3; colour index 68.

Crinanite is an analcime-olivine-diabase originally named by J. S. Flett in Argyll, Scotland. Like all diabases it varies from coarse-grained to fine-grained and some of it is dyscrystalline. A typical example described by G. W. Tyrrell (2) contains orthoclase 10, labradorite 30, analcime 10, titanaugite 25, olivine 15, ore and apatite 10; colour index 50.

I.—NEPHELINE-TEPHRITE AND RELATED ROCKS. DUδ, DWδ

Nepheline-tephrite is the effusive rock corresponding to theralite (page 472). Tröger gives the following as a typical mode:—labradorite 42, nepheline 15, titanaugite 30, soda-orthoclase 4, ore and apatite 9; colour index 39.

The name "atlantite" was given by E. Lehmann (2) to a nepheline-basanite of Nyassaland, East Africa, containing labradorite 31, nepheline 11, titanaugite 42, olivine 5; ore and apatite 11; colour index 58 (mode by Tröger). The essexite-basalt described by Lehmann (3) from Stöffel, Germany, is a rock of similar character. The mode is

sanidine	10·2	titanaugite	35·9
labradorite	23·0	olivine	20·2
nepheline	7·8	ore, apatite	2·9
		Colour index	(59)

Basanites have been described by T. F. W. Barth from Pacific islands. The following modes refer to rocks from (1) Niihau, (2) Haleakala, (3) Kohala.

		(1)	(2)	(3)
crystals	plagioclase	40	20	45
	pyroxene	20	35	10
	olivine	5	5	10
	ore	4	10	10
	apatite	1	1	5
glass		30	29	20

No nepheline is seen among the crystals, but the norm calculated from a chemical analysis of rock no. 2 shows that the glass must contain about 13 per cent of feldspar, 10 of nepheline and 7 of iron oxides.

The analcime-basalts of Sardinia, which were formerly mistaken for leucite-basalts, have been shown by H. S. Washington (5) to have the following composition (mean of three modes):—

analcime	8	olivine	6
biotite	1·5	magnetite	15·5
augite	22·5	apatite	2
		glass	42

The glass in these rocks has the composition of a mixture of orthoclase, labradorite and analcime. The colour index is therefore 46.

The *teschenites* are a group of analcime-diabases which are sometimes fairly coarse-grained but more frequently microcrystalline. According to G. W. Tyrrell (3) teschenite is distinguished from crinanite by containing more analcime and less olivine. Soda-orthoclase is often present in crinanite, but never in teschenite. F. Walker (1) gives the following definition of teschenite:—Teschenites are dyke-rocks consisting essentially of plagioclase feldspar (which is partly or wholly analcitized), titanaugite, and analcime. Barkevikite is often present, and olivine is a frequent constituent. Walker refers especially to the "patchy replacement by plagioclase" by analcime, and considers this characteristic of teschenite.

Some teschenite is leucocratic; according to Walker (2) the teschenite of Charlestown, Fifeshire, contains plagioclase 70, analcime 10, augite 15, olivine 5; colour index 20. Most examples are mesotype, with a colour index not far from 50, and many of the teschenites of Scotland, as well as those of the type region in Moravia, are metaluminous. Both types have been recorded by A. Knopf in the Spanish Peaks region, Colorado.

The following modes of Scottish teschenites are by G. W. Tyrrell (3).

orthoclase	10·2	—	—
labradorite	23·2	28·9	6·6
analcime	16·1	13·8	12·2
biotite	3·4	2·9	3·7
hornblende	—	—	18·1
augite	28·1	39·9	24·0
olivine	10·6	4·6	32·5
ore, apatite	8·4	9·9	2·9
Colour index	(50)	(57)	(81)

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