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A partisan review of Proterozoic anorthosites[1]

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Abstract

Most anorthosites of the massif type crystallized in the episode 1.7–1.2 Gyr, with a pronounced peak of the age distribution near 1.4 Gyr. They were emplaced anorogenically at depths as shallow as 7 km, where the ambient temperature of country rocks was probably less than 250°C. Depths of emplacement may have been as great as 25 km or more in rare cases; the greater depths of equilibration estimated from granulite facies metamorphism may be incorrectly interpreted as emplacement depths, but in any case they are demonstrably not required or characteristic of anorthosite emplacement. Penetrative deformation and metamorphism of anorthosites are post-emplacement accidents of the local geologic history, and are not directly caused by the presence of anorthosites.

Granitic rocks (mangerite–charnockite suite) associated with anorthosite are in general later or contemporaneous products of crustal anatexis, with chemical and isotopic signatures distinct from the anorthosites and their residua. Such granitic rocks should not, therefore, be summed with the anorthositic rocks to obtain bulk compositions.

The magmas that produced most anorthosites were dry, as shown by high-temperature mineralogy and anhydrous mineral assemblages in contact aureoles. Residua from their crystallization are ferrodiorites to ferrosyenites typical of closed-system fractionation. These residua were locally and frequently ejected into contemporaneously molten granite, where they formed pillows and cooled rapidly.

The overall chemistry of anorthosites and residua is broadly tholeiitic and consistent with derivation from the mantle. Olivine-bearing magmas locally ranged from leucotroctolite (anorthosite) to later but coexisting picrite or melatroctolite in the same pluton, confirming a wide spectrum of magma types. A signal feature of troctolitic and noritic magmas is their low augite content, implying high content of spinel component. Large anorthosite complexes such as Nain and Harp Lake consist of many plutons representing repeated injections of separate magma batches with varying chemistry.

The abundant true anorthosites, richer in plagioclase than magmas cosaturated with a mafic phase, must represent plagioclase enrichment by either mechanical or chemical processes or both. The role of kinetics in nucleation and solidification of such rocks may be centrally important. It is proposed that hyperfeldspathic (plagioclase-supersaturated) liquids were generated by quasi-isothermal extraction of mafic minerals from tholeiitic magma enroute to and at the site of emplacement, and that such a kinetic process was uniquely permitted in an environment of aborted continental rifting. Anorthositic rocks may have much to say about the episodic versus continuous geochemical evolution of the earth's mantle.

Introduction

Known anorthosites in the Solar System can easily be classified as lunar, Archean, Proterozoic, and Phanerozoic. Most but not all massif anorthosites[2] are Proterozoic in age, and all major occurrences of Proterozoic anorthosite are of the massif type (possible minor exceptions include Pikes Peak,

[1]Dedicated on behalf of all students of anorthosite to the memories of A. F. Buddington and E. P. Wheeler, who sustained us all so long by their marvelous example and enthusiastic support.

[2]The term "Adirondack-type" used by some U.S. authors is not well suited to a characterization of massif anorthosite in general (Emslie, 1980, p. 80) and seems a bit provincial in view of the overwhelmingly greater abundance of massif anorthosite in Québec and Labrador; its use could profitably be abandoned.

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Barker and others, 1975; and pre-Gardar xenoliths, Bridgwater and Harry, 1968). Lunar anorthosites occur solely as fragments forming a major component of the lunar highlands (Wood *et al.*, 1970; Walker *et al.*, 1973; Smith, 1979, 1980). The lunar investigators have introduced the useful term ANT (anorthosite–norite–troctolite) suite, which applies very well to many terrestrial occurrences. The popular view is that lunar anorthosites represent flotation cumulates (earliest crust) from a lunar magma ocean (see Smith, 1979). A peculiar feature of the lunar ANT suite is that it shows two trends on an En–An diagram; references are given by Raedeke and McCallum (1980), who find similar trends in the terrestrial Stillwater Complex and explain them in terms of modal relations and the fractionation of trapped intercumulus liquid. Students of massif anorthosites may be able to shed further light on such trends and processes, but many chemical and genetic differences between lunar and terrestrial anorthosites require the exercise of caution in comparisons.

Terrestrial Archean anorthosites such as those at Fiskenaesset, Limpopo, Sittampundi (Weaver *et al.*, 1981; Windley *et al.*, 1979), Shawmere (Simmons and Hanson, 1978) and Okhakh-Tessiuyakh (Wiener, 1981) occur mainly (solely?) as minor layers in complexes of overall basaltic composition, which contain gabbroic and ultramafic rocks as well. Spherulitic plagioclase megacrysts resembling those in younger mafic rocks (Berg, 1980) occur in some layers. The setting of the anorthosite-bearing complexes appears to be that of ocean floor subducted at continental margins (Weaver *et al.*, 1981; Windley *et al.*, 1979).

Phanerozoic anorthosites include an important Lower Paleozoic massif anorthosite at Sept Iles, Québec (Higgins and Doig, 1977, 1981) and olivine-bearing anorthosite in the Lower Paleozoic ring complexes of Air, Niger (Husch and Moreau, 1981), thus demonstrating that anorthosite, even of the massif type, is not limited to a Proterozoic "event".

The Baltimore–Wilmington complex of the eastern U.S. seaboard contains anorthosite layers and segregations, and may possibly turn out to be a Lower Paleozoic equivalent of Archean anorthosite.[3]

[3]I am indebted to M. L. Crawford for calling attention to the current work of Allan Thompson (University of Delaware) which suggests this notion to me, and to R. W. Bromery for pointing out that aeromagnetic and gravity observations permit the correlation of the Baltimore "gabbro" with the Wilmington complex.

Proterozoic anorthosites

Excellent reviews have appeared recently (Duchesne and Demaiffe, 1978; Emslie, 1978a, 1978b, 1980; Ashwal and Seifert, 1980), and should be consulted for many details. In the present review I focus particularly on some of the more tenacious mythology of anorthosites and try to sum up the state of knowledge using the recent evidence that has come to my attention, particularly from Labrador. A central purpose here is to restate the constraints on anorthosite genesis and to redefine the anorthosite problem in terms that may help to focus future research.

Age

The geochronology of massif anorthosites is far from adequate despite significant advances. The maximum age range would appear to be from as old as 2.3 Gyr in the USSR (Moskin and Dagelaiskaya, 1972) to 0.5 Gyr at Sept Iles, Québec (Higgins and Doig, 1981). The rapakivi/anorthosite suites of northern and eastern Europe are typically 1.7 Gyr old (Emslie, 1980). The Mealy Mountains anorthosite, which lies well south of the Grenville Front in Labrador, is at least 1.65 Gyr old (Rb–Sr and zircon; R. F. Emslie, personal communication, 1981), and an age of 1.4–1.5 Gyr characterizes most other Labrador anorthosites as well as Laramie, Wolf River, and a large suite of rapakivi and similar rocks in North America (see for review Emslie, 1980). The crystallization of the Adirondack anorthosite has been dated at 1.1 Gyr by Silver (zircon, 1968) and 1.2 Gyr by Ashwal *et al.*, (Sm–Nd, 1980). The San Gabriel anorthosite–syenite body of California also has a crystallization age of 1.2 Gyr (zircon: for references see Carter and Silver, 1972). Crystallization ages near 0.9 Gyr are reported for the Rogaland anorthosites (Pasteels *et al.*, 1979). Some residual doubt remains as to whether the younger ages are metamorphic rather than igneous (Emslie, 1980; Turner *et al.*, 1981).

If the Sm–Nd results for the Adirondack massif are accepted as persuasive, the range of most anorthosite ages in North America is evidently 1.65–1.2 Gyr; the range for European occurrences is apparently similar (1.7–0.9 Gyr). The total range is comparable to that of the entire Phanerozoic Era. Although the idea of an anorthosite "event" must be discarded on the evidence, one may speak of an anorthosite *episode* as an incidental passage in the history of the world in which more anorthosite was emplaced than in any other age; but much more information is needed to characterize this episode.

Depth of emplacement

Berg (1977a, 1977b) found a pressure arch from 3.7 to 6.6 kbar for contact metamorphic aureoles of the Nain complex, based chiefly on cordierite–garnet (hypersthene) barometry. The entire range of pressures has since been revised downward by about 1.5 kbar (Berg, 1979 and personal communication, 1981; see also Bohlen and Boettcher, 1981), a result which leaves adequate room for the observed andalusite–sillimanite type of metamorphism and the lack of any known kyanite in the high-pressure central position of the arch. Other Labrador bodies are associated with aureoles and chilled margins that also suggest low to moderate pressures of intrusion (Emslie, 1980). Shallow emplacement was shown for Laramie (Frost and Lindsley, 1981) and suggested for San Gabriel (Carter and Silver, 1972) and for the Adirondacks (Tracy *et al.*, 1978; Valley and O'Neil, 1981).

The En = Fo + Q reaction can be crossed with An = CaTs + Q to infer crystallization pressures of 6 to 6.5 kbar in the Mealy Mountains anorthosites, Labrador; olivine–orthopyroxene equilibria in ferromonzonites suggest even higher pressures of 7–8 kbar (R. F. Emslie, personal communications, 1979, 1981), close to the 9 kbar pressure inferred for the contemporaneously crystallized Red Wine metamorphic complex (Emslie, 1981). Similar high pressures (6–9 kbar) were also found for the Wolf River batholith, Wisconsin (Anderson, 1980).

Based on these findings, the depth of emplacement of massif anorthosite in Labrador and elsewhere was most commonly in the range 5–13 km but rarely as great as 23–27 km. The frequently-cited genetic "relationship" between anorthosites and granulite facies metamorphism is simply wrong and should be abandoned. Anorthosites and granulites do occur together and *may* be related but they *need not* be as a condition for anorthosite genesis (see also Turner, 1980).

Metamorphism

The anorthosites of northern Labrador, California, Wyoming, and Wisconsin are undeformed and unrecrystallized by post-anorthosite events, and even the Mealy Mountain bodies, which lie well within the Grenville Province of Labrador, are hardly recrystallized and still carry the assemblage olivine + plagioclase (Emslie and Bonardi, 1979). Gabbroic assemblages in central Labrador have cooled from perhaps 9 kbar and 1000°C with only minor development of garnet near contacts with paragneiss (Emslie and others, 1978; Emslie, 1981). The retrograde or autometamorphic development of metamorphic mineral assemblages (for example Yoder, 1969; Martignole and Schrijver, 1970; Ashwal *et al.*, 1981; Woussen *et al.*, 1981) is not an inevitable consequence of the cooling of igneous bodies at high pressure, and the observed textures could better be interpreted as due to later prograde metamorphism. In any event, metamorphism is clearly not an important constraint on anorthosite genesis.

Tectonic setting

Most massif anorthosites are anorogenic (Emslie, 1978a, 1978b, 1980; Berg, 1977a). The evidence in Labrador is impressive: the last major regional recrystallization in north-central Labrador occurred 2.7–2.4 Gyr ago; granites as old as 2.3 Gyr are undeformed; the ≥1.8 Gyr-old Snyder group of supracrustal rocks was unmetamorphosed and undeformed until emplacement of the Nain anorthosites and Kiglapait intrusion at about 1.4 Gyr (for references see Morse, 1979b). Similar criteria apply to other parts of Labrador (Emslie, 1980). Anorogenic emplacement seems so clearly demonstrated for these large bodies as to require special care in the interpretation of deformed anorthosites elsewhere.

A rifting environment for anorthosite genesis was suggested by Bridgwater and Windley (1973) and Berg (1977a). Emslie (for example, 1978; 1980) points out that rifting locally postdated the emplacement of anorthosite; but such events would seem not to preclude an earlier, failed continental rifting event. All the early-rifting ideas rest on circumstantial evidence (for example, Berg, 1977a), but receive support from analogy with the Duluth Complex, which is associated with the midcontinent rift, and Sept Iles in the St. Lawrence Graben (Higgins and Doig, 1981). The concept of *aborted* continental rifting has much in its favor as a setting for anorthosites; at very least, it makes room for the voluminous plutons.

Gravity data

The popular thin-sheet model for the Adirondack anorthosite (Simmons, 1964) rested on a standard density for anorthosite corresponding to a color index (CI) of only 4. For a more realistic value of

CI~15, a relatively high content of mafic minerals is implied for the surroundings, and it is not clear from the data that the anorthosite massif is either a thin sheet or unaccompanied by a mafic counterpart (Morse, 1968, p. 185). Indeed, wavelength-filtered Bouguer anomaly maps that integrate over most of the crustal thickness consistently show the Adirondack massif as a gravity high of 10 mgal or more (Simpson and others, 1981, Kerr, 1982; Arvidson and others, 1983), indicating excess mass in the crustal column. Similarly, Tanner (1969) identified six large local positive Bouguer anomalies over known anorthositic intrusions in eastern Québec and Labrador, and showed that these bodies were relatively thick sheets with inward-dipping margins. Although no such positive anomalies were detected from unfiltered data over anorthosites west of about 73°W, it appears likely from the Adirondack example that wavelength filtering would reveal them.

In sum, the gravity data suggest that mafic counterparts to anorthosite may presently reside in the crust or in subjacent shallow mantle.

Role of water

Yoder's (1969) idea of generating anorthosite from hydrous melts proved to be not generally valid for massif anorthosite. The case for wet melts presupposes that H_2O would eventually be a fugitive component, but the contact aureoles of anorthosite are conspicuously dry, particularly where they contain osumilite (Berg and Wheeler, 1976; Berg, 1977b; Maijer *et al.*, 1977), which has an exceedingly low tolerance for H_2O (Olesch and Seifert, 1981). Late stage fractionation products of troctolite and anorthosite should reveal any H_2O that remained in solution, yet these are either conspicuously dry, with hypersolvus or high-solvus pyroxenes and feldspars (Morse, 1979b, 1981b; Huntington, 1979, 1980; Ranson, 1981) or at least far below saturation in H_2O (Wiebe, 1978). A fugitive component cannot flee if it remains in solution in residual liquids and crystallizes in hydrous minerals, as seems to be the case in the Fongen–Hyllingen intrusion of Norway, where dry assemblages at the base give way progressively to wet assemblages at the top (Esbensen, 1978; Wilson *et al.*, 1981).

Wet magmas, in short, cannot and do not explain massif anorthosite. Quite the contrary, most evidence points to magmas extremely poor in H_2O, like modern mid-ocean ridge basalts (Presnall *et al.*, 1979).

Mineralogy

The mafic silicates of anorthosite are dominated by hypersthene, followed by olivine and augite. In Labrador, orthopyroxene compositions in anorthosite span the range En_{80}–En_{20}, and generate a triangular envelope with apex at En_{80}, An_{80} on an En–An[4] diagram (Fig. 1). The steep and shallow En–An trends of the lunar and Stillwater rocks (see Introduction) are not clearly reproduced by the terrestrial massif anorthosite data, which tend to scatter throughout the envelope.

Many large to giant hypersthene megacrysts in anorthosite are aluminous and now contain exsolved plagioclase lamellae, either because they crystallized at high pressure (Emslie, 1975) or crystallized rapidly, in place, from plagioclase-rich magmas (Morse, 1975a, 1975b; Gromet and Dymek, 1981). Plagioclase in the anorthosites of the Nain complex ranges in composition from An_{90} to An_{34} in a roughly gaussian distribution having maxima in the high 40's and low 50's (Morse, 1977). The range and shape of the distribution recalls Bowen's (1928) gaussian distribution for basalt centered near An_{50}; modern data for basalts in RKNFSYS (Chayes, 1975) are similar.

Although delicate oscillatory zoning of iridescent plagioclase is well preserved in some unmetamorphosed anorthosites, isocompositional cumulus crystals in sizes from millimeters to a meter across are common over wide areas, and signify adcumulus growth (for example, Emslie, 1980, Fig. 25a). Reversed rims on plagioclase occur (Speer and Ribbe, 1973; Emslie, 1980) and are probably common (Morse and Nolan, 1981; Dymek, 1981); they require explanation.

The rate of An fractionation by plagioclase feldspar appears to be a function of the augite content of the liquid (Morse, 1979a): in augite-poor liquids the equilibrium plagioclase composition changes much more slowly with crystallization than in augite-rich liquids. This effect may help account for the relatively limited plagioclase composition range found in many individual anorthosite bodies which are poor in augite.

[4]Fo–An diagrams can be transformed to En–An diagrams by the relation $X_{En} = aX_{Fo} + b$, where $a + b = 1.0$. The data of Medaris (1969) yield $a = 0.87$ with a correlation coefficient $r = 0.999$ (I am indebted to R. F. Emslie for pointing this out to me); Morse (1979d) used $a = 0.85$ based on experimental and natural data. I suggest we call the relation Medaris's rule, and perhaps choose a value of $a = 0.86$ as a reasonable compromise for natural samples.

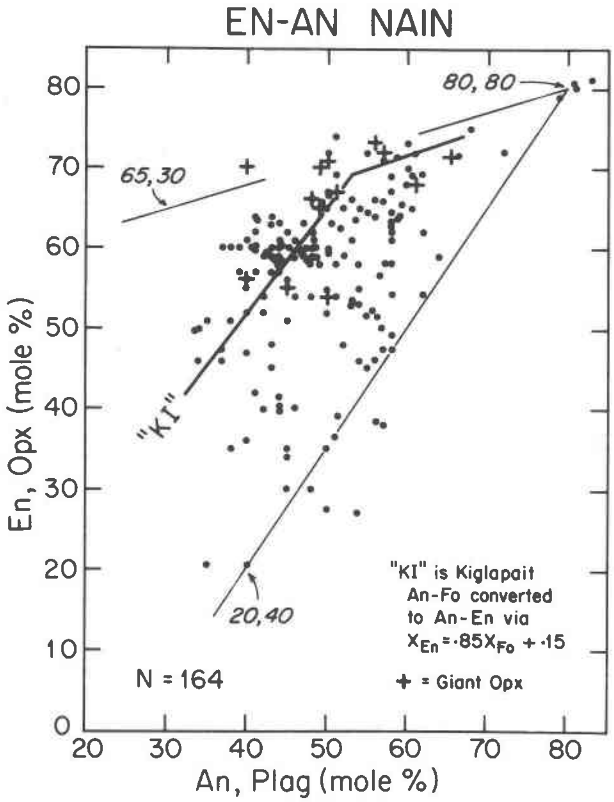

Fig. 1. En–An diagram for anorthosites, norites, and troctolites of the Nain complex, Labrador. Data from Ranson (1981) and unpublished maps of E. P. Wheeler II. Kiglapait trend (converted from Fo–An) from data in Morse (1979*b*).

The specific dichotomy between labradorite and andesine types of anorthosite proposed by Anderson and Morin (1969) does not hold up in general (Romey, 1968; Ranson, 1981; Morse, 1977), although broad compositional differences certainly exist among some individual plutons. Large anorthosite complexes are made up of dozens of such individual intrusions (Emslie, 1980; Morse, 1977).

Magmas: direct evidence

The chilled margins of anorthositic bodies are high-alumina troctolitic and noritic, more rarely gabbroic. Troctolitic examples with Al_2O_3 characteristically near 19 wt.% include Michikamau (Emslie, 1978), Hettasch (Berg, 1980), and Barth Island (Wheeler, 1968, analysis 2 of Table 3; de Waard, 1976). The summed bulk composition of the Kiglapait intrusion nearly matches the Hettasch chilled margin composition (Morse, 1981b), and similar rocks are widely distributed in North America (Nehru and Prinz, 1970).

Chilled leuconoritic margins with color indices in the range 10–18 and Al_2O_3 in the range 21–25 wt.% occur in at least two plutons of the southern Nain complex, and anorthosite dikes show similar compositions (Wiebe, 1978, 1980b). In the northern Nain complex, large leuconorite bodies have more nearly cotectic modes, whereas augite-bearing anorthosites have very low color indices (Ranson, 1981). Elsewhere in the Nain complex there is evidence for magmas as mafic as melatroctolite or picrite basalt (Berg, 1980).

High alumina marginal gabbros at Harp Lake average 18% Al_2O_3 and were considered parental by Emslie (1980), who now doubts that interpretation because of low REE in the marginal gabbros compared to the anorthositic rocks of the complex (personal communication, 1981). Low alumina marginal gabbros (15% Al_2O_3) are more fractionated but have REE contents more appropriate to equilibrium with anorthosites (Emslie, personal communication, 1981). Anorthosite, leuconorite, and leucotroctolite (28, 26, and 26% Al_2O_3, respectively) are considered by Emslie (1980) to be plagioclase cumulates.

Mixing models used by Weiblen and Morey (1980) for the Duluth complex yield a high alumina (18% Al_2O_3) parent for troctolite, norite, and olivine gabbro, but paradoxically a low alumina (9% Al_2O_3) *pyroxenitic* parent for anorthositic rocks (22%), *peridotite* (65%) and granophyre (13%).

The gabbroic anorthosite magma of Buddington (1939) closely resembles dikes and leuconorites at southern Nain in composition (Wiebe, 1979).

The color index of anorthosite dikes at Nain locally reaches very low values and at St. Urbain it reaches nearly to zero (Dymek, 1980; Gromet and Dymek, 1980). These observations imply the existence of magmas (not necessarily all liquid) extending to nearly pure anorthosite.

Cosaturation of plagioclase and mafic minerals probably occurs in the range 17–19% Al_2O_3 (color index roughly 20–25) for troctolitic and noritic magmas (Morse, 1979c, 1981b; Ranson, 1981). Whether more aluminous (21–25% Al_2O_3) compositions represent liquids (Wiebe) or crystal-laden magmas (Emslie) is debated. Both authors invoke removal of aluminous orthopyroxene at depth to produce feldspar-rich compositions. In the Emslie case, feldspar-laden magma must either be continually delivered from the deeper staging region during higher-level floor accumulation of anorthosite, or it must fill the local magma chamber and then the crystals must sink to form floor cumulates. In the

Wiebe case, the liquid must be supersaturated in plagioclase component. Problems relating to these modes of origin are discussed below under kinetic considerations.

Magmas: indirect evidence

One would like to invert the mineral compositions of anorthosite to obtain the compositions of the magmas from which they crystallized (for example, Gill and Murthy, 1970; Duchesne, 1971; Morse, 1974, 1981a; Duchesne, 1978; Duchesne and Demaiffe, 1978). This exercise is done via the mineral–melt partition coefficient $D = C^{S}/C^{L}$ where C is concentration and S and L are the crystal species and liquid respectively. The uncertainty in values of D and how they vary with T, P, and bulk composition (Irving, 1978) is so large that any of the prime candidates for anorthosite parents can be derived by varying the choice of D's. It is an interesting and valuable exercise to extract values of D from plutonic bodies themselves, for which the magma composition can be calculated (Duchesne, 1971; Paster and others, 1974; Roelandts and Duchesne, 1979; Shimizu, 1978; Morse, 1981a, 1982). The success of such a venture is inversely related to the number and nature of assumptions required to assess the parent magma composition, particularly those assumptions related to which rocks are to be included in the whole.

It becomes increasingly clear that the composition and structure of melt, kinetics of nucleation and growth, and details of postcumulus solidification may all affect the apparent or real value of D (for example, Mysen and Virgo, 1980; Takahashi and Irvine, 1981; Morse, 1981a, 1982), and an objective assessment of the correct value of D to be used will not always be easy. The values of D found by Morse (1981a) for K and Rb (feldspar/melt) are larger than the classical literature values by a factor of about five, which is enough to make a granodiorite parent into a basalt in term of these elements! The weight of empirical evidence favors (I would even say requires) use of the larger values of D (feldspar/melt), near 1.0 for K and Rb (Ranson, 1981), but the reasons why this should be so are as yet merely conjectural (Shimizu, 1978).

Somewhat better agreement between empirical field-related and experimental evidence obtains for Sr, with the realization that $D_{Sr}^{plag/liq}$ is strongly dependent on the augite content of the liquid (Morse, 1982). Data for phosphorus (Watson, 1979; Morse, 1981b) and REE (Roelandts and Duchesne, 1979) in apatite are useful for characterizing the parental liquids of late-stage residua, but it should be pointed out that these apatite-saturated rocks are on average very much like their parent liquids (Morse, 1981b), hence inversion is hardly necessary except when fine detail is sought. Roelandts and Duchesne (1979) have attempted to calculate f_{O_2} from the distribution of Eu between plagioclase and apatite.

An early goal of the inversion process was to distinguish between the classical rival hypotheses on magma types parental to anorthosite: gabbroic versus granodioritic. Events have largely overtaken this consideration; the field, petrographic, chemical, and isotopic evidence loudly proclaims in most cases a gabbroic (broadly speaking) parentage for the ANT to ferrodiorite suite and an independent, probably crustal-anatectic parentage for the associated granitic rocks. A more refined and more ambitious goal of inversion is to characterize the parental magmas of the ANT suite in detail (for example, Simmons and Hanson, 1978) and to be able to say something quite specific about their sources. It would also be nice to be able to distinguish feldspars grown from feldspar-supersaturated liquids from those grown at saturation with a mafic phase. Conceivably this could be done with minor elements such as Mg and Fe in plagioclase (Longhi *et al.*, 1976).

Residual liquids

Rocks persuasively shown to represent residual liquids from the crystallization of anorthosite parents are characterized by iron enrichment and the presence of Fe–Ti oxide minerals and apatite. They are ferrodiorites, ferromonzonites, and ferrosyenites, not the granitic rocks described in the next section. Examples occur in the Kiglapait intrusion (Morse, 1981b) southern Nain complex (Wiebe, 1978; 1979; 1980a; Huntington, 1980); Harp Lake complex (Emslie, 1980), the Adirondacks (Ashwal, 1982); and probably in South Rogaland (Wiebe, 1980a). The high temperature hypersthene monzonite of the Laramie complex (Fountain *et al.*, 1981) may also belong to this class. The group as a whole is characterized by high temperatures of crystallization commonly demonstrated by hypersolvus or barely subsolvus pyroxenes and ternary feldspars; estimates reach to more than 1100°C at about 5 kbar (Huntington, 1980). Dry conditions are clearly indicated for such temperatures. The iron enrichment (Fenner trend) is characteristic of closed-system fractionation (Osborn, 1959; Morse, 1980), and implies relatively little interaction with oxidizing crustal rocks, as does the low quartz content.

Granitic and hybrid rocks

Quartz-rich mangerites, charnockites, adamellites (quartz monzonites), granites, rapakivi and related rocks near anorthosite are increasingly being shown to represent independent melts chemically unrelated to anorthosite residua. Grounds for these conclusions include field evidence, isotopic relations, and chemical evidence including REE. Examples include southern Nain (Wiebe, 1978, 1980a); Adirondacks (Ashwal and Seifert, 1980) and possibly Rogaland (Pasteels *et al.*, 1979). At southern Nain, ferrodiorite pillows quenched in granitic melts decisively demonstrate the contemporaneous nature yet independent parentage of anorthosite residua and granitic magma. Such quenching relations are fatal to theories of anorthosite genesis by liquid immiscibility (for example, Philpotts, 1981). However, a small subset of granitic rocks shows evidence of representing an immiscible split from ferrodiorite (Wiebe, 1979).

The comingling of ferrodioritic and granitic melts leads to rocks that are hybrid on scales ranging from tens of meters to millimeters (Wiebe, 1978, 1980a). The possibilities for misinterpretation of these rocks are legion until one develops an eye for them.

Kinetic considerations

Plagioclase is lighter than melts that are cosaturated with a mafic phase and suitably high in Fe (Morse, 1973; Walker and Hays, 1977; Campbell *et al.*, 1978; Kushiro and Fujii, 1977; Kushiro, 1980; Morse, 1979b; McBirney and Noyes, 1979). The possibility of generating anorthosite or feldspar-laden magma by flotation therefore exists (Grout, 1928; Morse, 1968; Kushiro and Fujii, 1977), but the process requires a suitably low yield strength of magma (Murase and McBirney, 1973) and a suitable time scale if the crystals are not very large. Persuasive field evidence for roof accumulation on a large scale is still lacking.

Plagioclase will not float in melts strongly supersaturated in plagioclase because the density contrast will be neutral or of the wrong sign; it may not sink in such melts, either, if rapid nucleation and growth of crystals cause a high yield strength (Morse, 1979b).

The barrier to nucleation of plagioclase is sufficiently high (for example, Cranmer *et al.*, 1980; Uhlmann *et al.*, 1980) that supersaturation in plagioclase component is easily accomplished by nucleation and growth of mafic minerals such as olivine or pyroxene, whose nucleation barriers are lower (Grove, 1978; Berg, 1980). Although the experimental and field settings of this phenomenon described by Grove and by Berg, respectively, occurred as a result of thermal supercooling, the same result would arguably be produced by isothermal cotectic supersaturation (Morse, 1979b), which is likely to be an important process in plutonic, feldspathic magma bodies.

Diffusion-related controls on magma dynamics and crystallization processes currently receive much attention; examples include the Liesegang process and double-diffusive convection (McBirney and Noyes, 1979; Irvine, 1980). These studies refer to the liquidus case of one crystal species plus liquid, and may apply to plagioclase-supersaturated liquids. For the more commonly considered cotectic case of two crystal species plus liquid in the slow cooling of large bodies, the diffusion-related processes become self-damping and reduce to the case of oscillation about a cotectic equilibrium.

Isothermal supersaturation of plagioclase caused by nucleation of mafic minerals will bring the solidus composition toward the liquid composition on the feldspar loop (Morse, 1979b, p. 583). The feldspar may continue to crystallize at constant (metastable) composition by the phenomenon of steady-state growth (Hopper and Uhlmann, 1974). Such a process could explain why the plagioclase feldspar compositions of many anorthosite bodies closely approach the inferred liquid composition (for example, Wiebe, 1978, 1979) and account for the high partition coefficient $D_K^{plag/liq}$ inferred from studies of natural rocks (Shimizu, 1978; Morse, 1981a).

The concept of a crustal density filter for the admission of magma into the crust (Sparks *et al.*, 1980; Stolper and Walker, 1980; Neumann, 1980) is one which seems to have obvious relevance to anorthosite emplacement. Ordinary magmas fractionated by olivine extraction reach a density minimum when they become cosaturated with plagioclase, but if plagioclase fails to nucleate, the density will continue to fall. Very feldspathic magmas should have a density less than 2.65 g/cc, but may not rise far from their site of mafic extraction because of the onset of crystallization.

Origin of anorthositic magmas

It is useful to consider two classes of feldspathic magmas: *cotectic* ones which produce equilibrium proportions of plagioclase and a mafic mineral, and *hyperfeldspathic* ones which are either mechanically enriched or supersaturated in plagioclase (plagioclasic magma of Michot, 1968).

Cotectic magmas

Although critical experiments are lacking, one can estimate with some confidence that cosaturation occurs at a color index of 25±5 for olivine or hypersthene with plagioclase having mineral compositions typical of anorthosites, at the indicated pressures (Morse, 1979c: Ranson, 1981). Because the cotectic shifts toward plagioclase with increasing pressure, magmas that are cotectic at the site of emplacement (Fig. 2) were in the primary phase field of the mafic phase, unsaturated with plagioclase, during ascent. Conversely, cotectic magmas at depth lie in the plagioclase field on ascent. Hypersthene megacrysts, noritic chilled margins and olivine-rich basal layers or chilled margins attest to magmas reaching the cotectic via the field of the mafic phase; plagioclase megacrysts and low color index may indicate magmas that ascended through the plagioclase field. The former are more obvious even if the latter may be more common.

The high-alumina troctolitic magmas associated with the anorthosite episode differ from modern high-alumina basalts in having a very low augite content. The high-pressure phase complementary to augite is spinel, and a normal, augite-rich basaltic composition can in principle be transformed to an augite-poor one by addition of spinel (Morse, 1981b). Troctolitic magmas may, then, arise by abundant melting in a source wherein clinopyroxene is exhausted and spinel consumed but not exhausted.

Noritic magmas may originate by a similar process followed by olivine extraction or some other event (for example, oxidation; Morse, 1980) that raises their silica activity enroute to the site of emplacement.

The problem of production of plagioclase-rich but cotectic magmas is evidently merely one of phase petrology, a suitably spinel-rich and cpx-poor source material, and suitably high temperatures to yield relatively large fractions of melt. Such melts

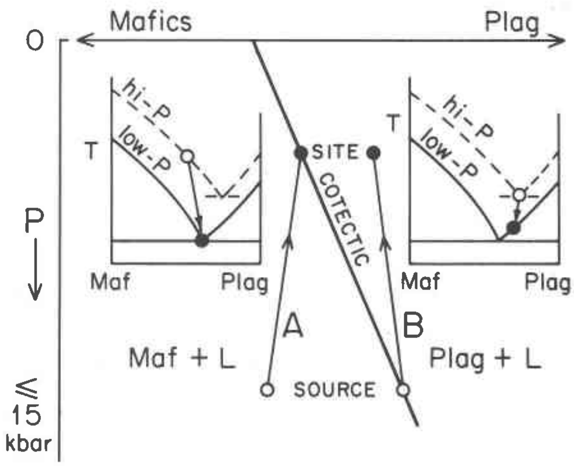

Fig. 2. A magma that is cotectic at the site of emplacement was in the field of the mafic phase on ascent (A), whereas a cotectic magma at depth rose through the plagioclase field (B). The shift of the cotectic with pressure probably reverses toward mafics at pressures greater than about 10–15 kbar.

would have low concentrations of incompatible elements by dilution; however, a depleted source is also required by element ratios such as K/Rb.

The evidence for melatroctolitic liquids (Berg, 1980) attests to very high temepratures at the source during the waning stages of anorthosite production in the Nain area.

Hyperfeldspathic magmas

Mechanical concentration of plagioclase, accompanied or followed by removal of residual liquid, can account for crystal-laden hyperfeldspathic magmas and must presumably be invoked to explain dikes or small bodies of nearly pure anorthosite. Extension of such a mechanism to the formation of large bodies encounters problems related to the inhibited motion of plagioclase, as discussed above, particularly where floor accumulation is well documented. The abundance of highly feldspathic bulk compositions represented by many dikes and plutons seems to require the existence of hyperfeldspathic, plagioclase-supersaturated *liquids*. Increased pressure and albite content will shift cotectic compositions toward plagioclase, but Fe has an opposite effect; a maximum plagioclase enrichment probably occurs near 15 kbar and would produce at most about 30% anorthosite followed by 70% cotectic material at lower pressure (Morse, 1979d).

Plagioclase supersaturation in the plutonic environment of the Nain comlex has been documented (Berg, 1980), so it is a process known to occur locally at least. It is evidently required repeatedly on the local scale by the bottom accumulation of feldspar in many layered intrusions where turbulent flow is contraindicated by plagioclase lamination (Morse, 1979b; McBirney and Noyes, 1979). Plagioclase supersaturation by oscillatory nucleation in large magma bodies is essentially an isothermal phenomenon caused by the large difference in barriers to nucleation of mafic minerals and plagioclase. Nucleation and growth of mafic minerals can drive the liquid composition metastably well into the plagioclase field. It requires, perhaps, a large leap of faith to extrapolate to a process whereby a large body of magma could become hyperfeldspathic, but that is what the field relations seem to require.

It is suggested, therefore, that the large hyperfeldspathic plutons typically having color index around 15 (or locally less) represent plagioclase-supersaturated liquids (Fig. 3) generated by the slow and steady removal of mafic minerals during ascent of the magma to its site of emplacement, followed by continued removal of mafics by sinking near the site of emplacement. The first stage of the process yields the cotectic composition or perhaps a slightly plagioclase-supersaturated composition, and the lost mafic minerals now reside in the mantle or lower crust. The second stage continues the process and yields minor amounts of lost mafics that are infrequently detected by geophysical means. The extraction process continues until plagioclase nucleation and growth proceed far enough to inhibit further settling of mafics. The sporadic occurrence in anorthosites of giant hypersthene megacrysts suggests that the sinking of mafic minerals could be very rapid because of their large crystal size.

The proposed mechanism of supersaturation could generate bulk compositions with positive Eu anomalies because Eu would be partitioned into the melt with the plagioclase component, relative to pyroxene (Sun and others, 1974; Gromet and Dymek, 1981) and it would account for the absence of aphyric anorthositic lavas because the drastic process of eruption would shear the liquid and cause rapid nucleation and growth of plagioclase (Morse, 1979b).

All these ideas about magma genesis benefit from the concept that anorthosites are associated with failed continental rifts or at least tensional settings wherein large magma bodies can be accommodated in the crust without the concurrent eruption which would deliver ordinary basic magmas to the surface. Eruption removes heat efficiently, whereas ensialic plutonism tends to bottle heat up and prolong the magmatic episode, promoting the slow processes required for extensive supersaturation in plagioclase component.

Geochemical implications

Mg, An, and Sr–isotope ratios

The molar X_{Mg} = MgO/(MgO + FeO) of anorthosite parent magmas lies typically in the range 0.45–0.6, well short of the values considered to be normal for modern MORB (Emslie, 1980). The early fractionation of mafic minerals would tend to account for this difference, assuming that the mantle source of anorthosite parents was similar to modern suboceanic mantle.

The range of X_{An} in plagioclase of MORB is 0.05 to 0.9 (Papike and Bence, 1978), greater than that found in any anorthosite complex. I have not seen

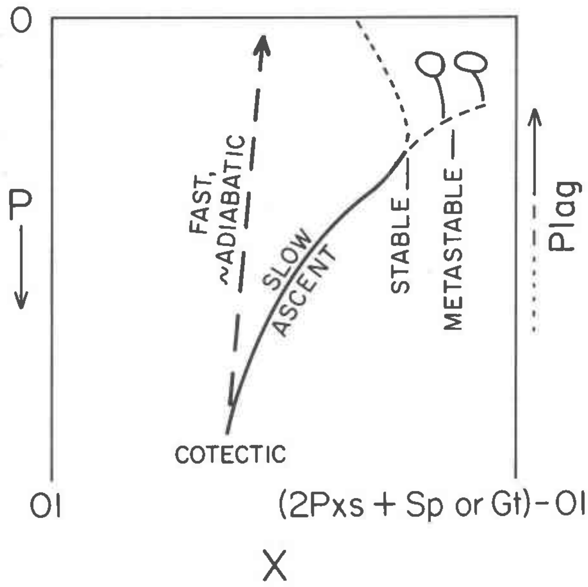

Fig. 3. Composition of liquids as a function of pressure in the system olivine–plagioclase. At high pressures the plagioclase component is represented by pyroxenes + spinel or garnet, minus olivine, and multiply-saturated melts at the source are rich in olivine. Rapid eruption or emplacement of such a deeply-generated magma would result in a mafic product such as picrite–basalt or melatroctolite (Berg, 1980). Slow ascent attended by removal of olivine ± pyroxene would cause the magma to become plagioclase-rich to some stable limit, beyond which it would become mafic-rich again if plagioclase were removed. If plagioclase failed to nucleate, however, hyperfeldspathic liquids would result from further extraction of mafics (metastable trend). These liquids would be expected to shed plagioclase on emplacement and to form plagioclase-rich leuconorites or leucotroctolites (ovals).

the frequency distribution for MORB but it is not obvious that it will be greatly different from that of Bowen (1928) or more recent compilations for basalt in general. Hence it may resemble the distribution for anorthosites. If so, we can continue to say that anorthosites are like basalts in their An content, being, like them, dominated by compositions lying at some distance from the primary end of the spectrum assumed to represent equilibrium with the mantle source. If it is true that only a small fraction of primitive melt gets erupted and that most of the magma born in the mantle dies there as well, as implied by concepts of magma mixing beneath ocean ridges (for example, O'Hara, 1977; Sparks *et al.*, 1980) it should be equally true of the magma in the subanorthosite mantle. Only the evolved magmas will routinely ascend through the crustal density filter. The trouble with this idea is that it would seem to require plagioclase fractionation in order to achieve the lower An content, a process not expected in most of the models outlined above. It is possible that the fractionated An component was removed in aluminous pyroxene (Morse, 1975; Emslie, 1975, 1980; Wiebe, 1980b; see also Gasparik and Lindsley, 1980) at depth. Alternatively, a source relatively rich in Ab or Jd component must be considered.

The well-known elevation of initial Sr–isotope ratios in anorthosites, unsupported by parental Rb (Duchesne and Demaiffe, 1978) is usually taken to imply crustal contamination. However, the assumed crustal melts represented by granitic rocks associated with anorthosite in the Nain complex tend to have initial ratios not greatly different from those of anorthosite, and inexplicably reaching to the abnormally low value of 0.7007 (E. C. Simmons, personal communication, 1980). The high Sr content and low Rb/Sr of anorthosite relative to crustal melt preclude contamination in place as a general source of the excess radiogenic Sr.

Implications for mantle evolution

Each of the characteristics discussed above admits of several explanations, and one which might fit them all is the possibility that the subanorthosite mantle was abnormally rich in Fe, Ab, and radiogenic strontium, yet depleted in Rb and other incompatible elements. Such a description would apply to an old enriched mantle (primitive or chondritic or long-before enriched with crustal material) later stripped of a relatively small low-melting fraction. Such a "mantle keel" of old enriched lithosphere has been postulated on the basis of Pb and Nd isotopic evidence for the Snake River Plain–Yellowstone Province of the Western United States (Leeman and Doe, 1982; Menzies *et al.*, 1982). The key to such investigations lies in the Pb isotope ratios which are sorely lacking for anorthositic massifs. The geochemical and isotopic integrity of many anorthositic bodies (and particularly their fresh troctolitic members which are so petrographically sensitive to alteration) is by now so well established that they should be seriously considered as geochemical windows on their mantle source regions. Any demonstration that these source regions were analogous to modern subcontinental lithospheric regions would tend to diminish the suspected role of mantle geochemistry in the origin of an anorthosite episode, and increase the suspected role of tectonics, which reflect the thermal evolution of the earth.

Summary

The mantle origin of magmas that produced anorthosite and other members of the ANT suite can hardly be doubted. The terrestrial ANT suite contains many pristine rocks that carry strong memories of their mantle parentage; their low Rb content, in particular, makes them extremely sensitive to *in situ* crustal contamination, which has therefore probably not occurred in significant degree. Volcanic rocks of comparable age are scarce and ill-preserved. The ANT-suite rocks, therefore, offer an incomparable window into the geochemistry of at least one type of the Proterozoic mantle. They should in time, with careful effort, allow us to place good constraints on the history of that mantle, and hence perhaps on the episodic versus continuous geochemical evolution of the earth (Jacobsen and Wasserburg, 1979; DePaolo, 1980).

Along with intensive geochemical investigation will have to come new understanding of the role of kinetics and melt structure in the production and evolution of magmas and the solidification of cumulate rocks. The effects of pressure and bulk composition on melt structure and element partitioning also require investigation as they relate to anorthosites. Firm evidence for the production of associated granitic magmas by crustal anatexis needs to be sought, and if found, the granitic rocks should be used as probes for the chemistry of the deeper Proterozoic crust in these regions.

The partisan message of this review, then, is that Proterozoic anorthosites afford the avenue to a

solution of major problems of earth history: the anorthosite problem becomes, potentially, the anorthosite solution.

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