Geochemistry of the Scourian complex: petrogenesis and tectonic models

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SUMMARY: The segmentation of the Lewisian complex which occurred at the end of the Archaean resulted in the juxtaposition of zones representing different crustal levels, including deep crustal granulites. Critical aspects of the geochemistry of Lewisian gneisses are considered in relation to their petrogenesis and tectonic models. Most of the Lewisian is made up of a bimodal suite of tonalitic–trondhjemitic gneisses enclosing mafic–ultramafic enclaves and layered complexes which are often associated with metasediments. These mafic components have a low-pressure petrogenesis and represent tectonic inclusions incorporated into the complex at the time of gneiss generation. They are more abundant in the granulites. The silicic gneisses have a higher pressure petrogenesis and their geochemistry is consistent with hydrous partial melting of amphibolites in a shallow-dipping subducting slab with hornblende as a residual phase.

Granulite metamorphism occurred or peaked some 200 Ma or so after the phase of gneiss generation at 2.9 Ga and terminated about 2.5 Ga. Concentrations of U, Th and Rb are at least an order of magnitude lower in the granulites than in the equivalent amphibolite-facies gneiss zones, and those of K and Pb about 50% lower. Concentrations of other elements are not significantly different. Currently popular models for the removal of these elements from the deep crust through partial melting or transport in a carbonic fluid phase are examined, but each has problems in accounting for all the available data. An alternative model, considering the low contents of heat-producing elements as a primary feature of silicic magmas emplaced into the deep crust, is explored. This relies more on fluid processes removing heat-producing elements from the ocean crust during subduction, followed by hydrous melting of the amphibolite to produce water-saturated magmas which cannot rise to high crustal levels. In contrast to modern subduction zone magmas, which are largely derived from the mantle wedge that has been modified by addition of a fluid-transported, LIL element, 'subduction component', Archaean sodic tonalite–trondhjemite magmas result from hydrous melting of amphibolite which has lost that component. Granulite-facies metamorphism is an independent process that is commonly superimposed on deep crustal rocks which are already low in heat-producing elements, and during which further loss of these elements may occur, but granulite-metamorphism may also be superimposed on more normal crustal rocks without significant chemical changes.

Introduction

A considerable number of geochemical studies have been carried out on the Lewisian—probably more than on any other Precambrian high-grade terrain—and it is the aim of this paper to focus on critical aspects of Lewisian geochemistry with relevance to the petrogenesis of the gneisses and their subsequent evolution. An immediate issue of concern is whether the Lewisian can be considered as typical of Archaean high-grade terrains and can therefore be used as a model for Archaean crust-generating processes in general (*eg* Weaver & Tarney 1984), or whether it is somehow unique and unrepresentative (*eg* Rudnick *et al.* 1985). In this regard it should be stressed that the period during which the Lewisian complex was generated and evolved in the late Archaean (*c.* 2.9–2.6 Ga; Hamilton *et al.* 1979) coincides with perhaps the most voluminous period of crustal generation in Earth history. On the other hand the Lewisian is unusual in the sense that, as a consequence of the major tectonic activity which terminated the Archaean (*ie* the start of the Inverian), segments representing different crustal levels were juxtaposed (Sheraton *et al.* 1973; Park & Tarney, this vol.), and are available for study.

There are some obvious lithological differences between these Lewisian segments. The northern part of the central granulite zone, between Scourie and Achiltibuie, has a particularly high proportion of lensoid mafic and ultramafic masses, and vast numbers of smaller mafic and ultramafic enclaves within the strongly banded tonalitic gneisses which have been derived from these mafic–ultramafic complexes through extreme tectonic disruption at the time of gneiss formation. Gneiss compositions thus range from Mg-rich dunite to silicic trondhjemite, and there is a higher proportion of banded gneisses of intermediate composition than elsewhere (Sheraton *et al.*

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1973). In the southern part of the central zone, at Gruinard Bay, where the complex appears to have suffered only hornblende-granulite facies metamorphism (Rollinson & Fowler, this vol.), the host gneisses are mainly silicic trondhjemites with a poorly developed foliation and whereas there are numerous inclusions, these are dominantly mafic rather than ultramafic in composition. In the flanking zones of amphibolite-facies gneisses, in the south at Torridon and in the north at Laxford, the gneisses are generally more potassic, and mafic enclaves are much less common. These observations are at least compatible with geophysical models for the crust which imply an increasingly more basic composition with depth.

Two other points are worth making. The first is that although the average composition of Lewisian granulites is rather more basic than that of the amphibolite-facies segments (Weaver & Tarney 1984, 1985) silicic quartz-bearing tonalites and trondhjemites are still represented. It is the higher proportion of the mafic–ultramafic component which makes the mean granulite composition somewhat more basic than average continental crust estimates. The second is that although obvious metasedimentary gneisses form only a small proportion of the early gneiss complex, these metasedimentary components mostly occur in close association with the mafic–ultramafic bodies. Just as the latter are more common in the granulite zone, metasediments thus appear to be an important, though minor, component of the deeper Lewisian crust, a feature which must have tectonic significance for the evolution of high-grade terrains. Suggestions that the whole Lewisian complex may have a sedimentary parentage (Lowman 1984) find little support from geochemical studies (Weaver & Tarney 1980); the rocks are dominantly orthogneiss.

Lewisian granulites have mineral assemblages which record P–T conditions of at least 8 kb and 800°C (Cartwright & Barnicoat, this vol.; Sills & Rollinson, this vol.), conditions which are appropriate to the deep crust. However, the fact that these granulites are now exposed at the surface, and the present crustal thickness must be at least 25 km, inevitably raises the question of what underlies the granulite terrain. One possible model is to appeal to exhumation of a thickened crustal keel of Andean type, in which case the present granulites might be expected to be underlain by similar material, perhaps with a higher proportion of mafic–ultramafic components. Alternatively the major shear zones which terminated the Archaean may have been listric in form (see Park & Tarney, this vol., fig 2*b*) leading to ramping of deep crustal granulites above mid-crustal amphibolite-facies gneisses along the Laxford front, but ramping of granulites above granulites in the middle part of the central granulite zone in Assynt. Of interest in this regard is the distribution of the potash-rich 'Scourian' pegmatites (Sutton & Watson 1951) which were emplaced at *c.* 2.5 Ga (Giletti *et al.* 1961; Evans & Lambert 1974) shortly before the intrusion of the Scourie dykes at *c.* 2.4 Ga (Chapman 1979). These are relatively abundant in the granulite terrain just to the south of the Laxford front, but decrease in abundance southwards and are both rare and more potash-poor in the Assynt region (Sheraton *et al.* 1973). The granulite gneisses are too potash-poor to yield K-feldspar rich pegmatites on partial melting (their melting products tend to be trondhjemitic in composition). However pegmatites in the granulites south of the Laxford front could have resulted from the thermal effects of ramping deep-crustal granulites over mid-crustal amphibolite-facies gneisses, the source of the pegmatites (which have low K/Rb ratios) being the underlying amphibolite-facies gneisses. Thus although these early pegmatites have traditionally (Sutton & Watson 1951) been regarded as the last stage of the Scourian (Badcallian) cycle, they may in reality be linked to the tectonic effects which initiated the Inverian and produced the segmentation of the Lewisian.

Available Sm–Nd and Pb–Pb whole rock geochronological data (Moorbath *et al.* 1969; Hamilton *et al.* 1979; Whitehouse & Moorbath 1986) suggest that the gneisses in both the granulite- and amphibolite-facies zones were generated at 2.92 Ga and that most suffered uranium loss at this time which led to their present unradiogenic Pb compositions. However Sm–Nd mineral isochrons for the granulites give ages of 2.49 Ga (Humphries & Cliff 1982) indicating that grain-size re-equilibration at granulite facies occurred until this time, some 400 Ma after crustal generation. This time of ultimate closure to REE diffusion may mark the uplift associated with Inverian segmentation of the Lewisian complex. Further analysis of the Sm–Nd mineral data led Humphries & Cliff (1982) to suggest that the peak of granulite-facies metamorphism occurred rather earlier at *c.* 2.62 Ga, an age supported by U–Pb zircon data (Pidgeon & Bowes 1972). A Pb–Pb whole rock isochron for the granulites (2.68 Ga: Chapman & Moorbath 1977) is also younger than that for the Lewisian complex as a whole (see also Holland & Lambert 1975), which suggests that U loss from the gneisses continued until this time. It would appear that the peak of Badcallian granulite-facies metamorphism occurred some 250 Ma after the time of formation of the gneisses. When

the deeper Lewisian gneisses first attained granulite grade is much less easy to define. Rollinson's data on ilmenite-magnetite intergrowths in late trondhjemite and granite sheets at Scourie (Rollinson 1980) are consistent with more hydrous fluid conditions before attainment of granulite-facies equilibrium.

The above are the main geochronological and geotectonic constraints relating to the age and distribution of different gneiss compositions in the Lewisian within which a petrogenetic discussion must be framed. It is preferable to consider the possible geochemical effects associated with granulite-facies metamorphism after those connected with the primary stage of gneiss generation.

Petrogenesis of Lewisian gneisses

There are a number of comprehensive geochemical studies of Lewisian gneisses, beginning with Sheraton (1970). Tarney *et al.* (1972) and Sheraton *et al.* (1973) first drew attention to the anomalously low contents of heat producing elements U, Th, Rb and (to a lesser extent) K in the granulite-facies gneisses compared with amphibolite-facies equivalents in the Lewisian and elsewhere. Because of the potential mobility of other lithophile elements during the granulite-facies event and later retrogressive phases, emphasis had necessarily to be placed on immobile elements such as the rare-earths in discussing the petrogenesis of the gneiss precursors (Weaver & Tarney 1980, 1981). For instance, although REE mobility occurred on a mineral scale during the granulite-facies event (Humphries & Cliff 1982), the fact that whole rock Sm–Nd isotope data on both granulite- and amphibolite-facies Lewisian rocks faithfully recorded the time of gneiss generation (Hamilton *et al.* 1979), suggested there had not been widespread REE mobility within the gneiss complex. Indeed the 2.92 Ga Sm–Nd isochron in itself tends to argue against significant melting of the Lewisian rocks during the granulite event which, on the evidence cited above, occurred or peaked *c.* 250 Ma later. Thus it can reasonably be assumed that the REE distributions reflect those of the gneiss precursors.

Petrogenetic models for the Lewisian must take account of the bimodal character of the gneisses, which is a feature of Archaean high-grade gneiss terrains worldwide. This bimodal character is less well marked in the granulites owing to extreme tectonic mixing of the two components.

The granulite REE data (Weaver & Tarney 1980) suggest that these two (mafic and silicic) components have a different petrogenesis. The mafic gneisses show a rather large range of Fe/Mg ratios which correlate positively with REE abundance and have essentially parallel REE distributions, all of which show moderate enrichment in light REE relative to chondrites (Fig. 1). Such patterns are typical of tholeiitic low pressure crystal fractionation involving olivines, pyroxenes and plagioclase. Some have slight negative Eu anomalies. The relative light REE enrichment would be compatible with melting of an 'undepleted' mantle source with chondritic REE distributions. The closely associated ultramafic gneisses also have 'undepleted' REE distributions but most have REE abundances more in keeping with komatiitic liquids than with residual mantle (Weaver & Tarney 1980; Sills *et al.* 1982). The close spatial association of mafic–ultramafic bodies with metasedimentary horizons in the gneisses is also consistent with a low pressure tectonic environment for the primary evolution of these bodies. Mafic gneisses with essentially similar REE characteristics and associations have been described in granulites from Fiskenaesset, West Greenland (Weaver *et al.* 1982), Madras, India (Weaver 1980) and NE China (Jahn & Zhang 1984). The apparently ubiquitous occurrence of such bodies in high-grade terrains, including also genetically-related layered gabbro–anorthosite complexes (Weaver *et al.* 1982), is suggestive of a common origin. The most realistic solution (Weaver & Tarney 1980) is to regard many of them as tectonic inclusions of Archaean ocean floor incorporated penecontemporaneously with crust generation (see Park & Tarney, this vol., Fig. 2*a*). Mafic bodies in the Laxford amphibolite facies gneisses of the Lewisian have essentially the same REE distributions as those in the granulites (Weaver & Tarney 1981) and appear to have a similar origin.

The silicic granulites have very different REE distributions (Fig. 1). They have strongly fractionated REE patterns showing first (compared with the mafic gneisses) enhanced light REE enrichment in the intermediate gneisses, then progressive heavy REE depletion in the tonalitic compositions, followed by both light REE and heavy REE depletion in the trondhjemites. The more silicic gneisses develop a marked positive Eu anomaly and show increased concave-upwards curvature in their heavy REE distributions (Weaver & Tarney 1980). A comparison of granulite- and amphibolite-facies tonalites at equivalent silica contents (Weaver & Tarney 1981) showed that there was no significant difference in their REE distributions. Very similar REE characteristics have now been described from most Archaean high-grade ter-

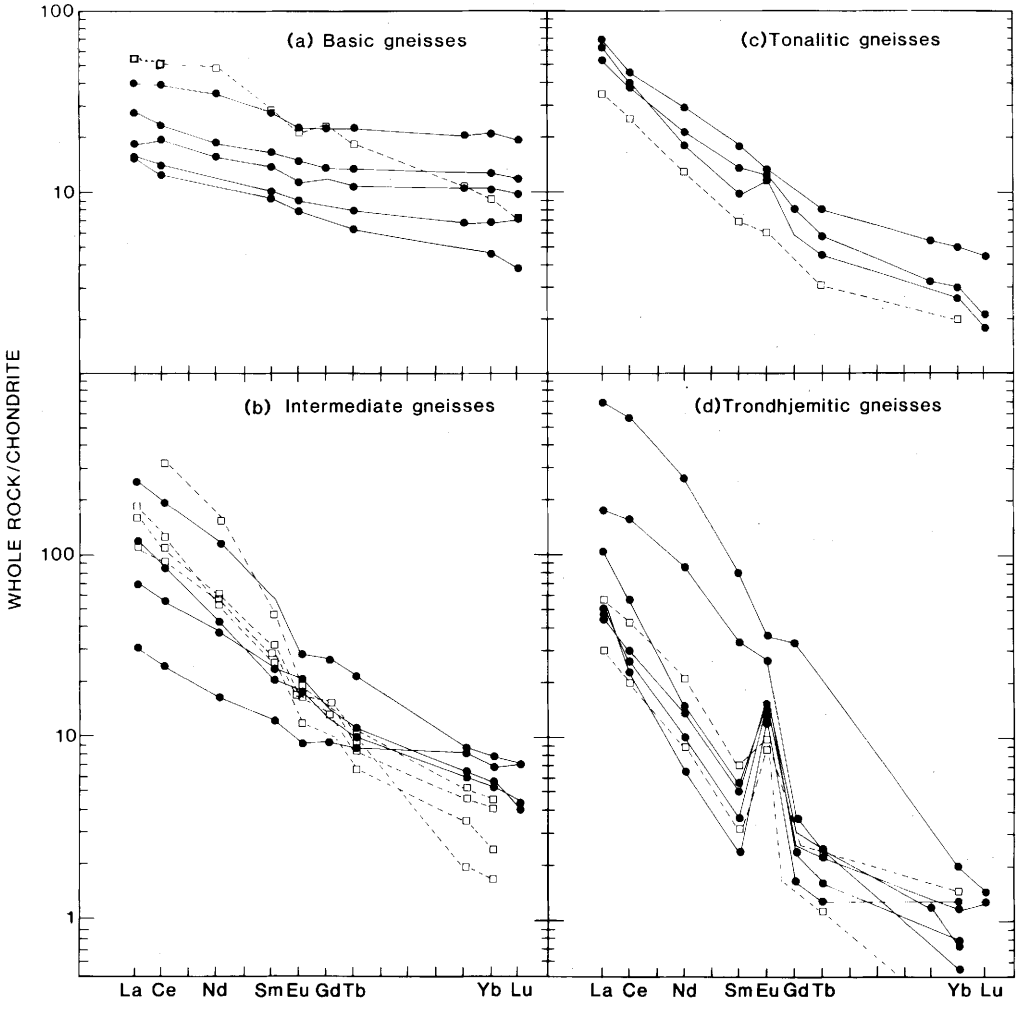

FIG. 1. Rare earth patterns for mafic, intermediate, tonalitic and trondhjemitic Lewisian granulites (full lines—after Weaver & Tarney 1980) and amphibolite-facies gneisses (dashed lines—after Weaver & Tarney 1981).

rains, irrespective of whether they are potash-poor like Lewisian granulites or moderately potash-rich such as the type charnockites from Madras (Weaver 1980; Weaver & Tarney 1983*a*). By contrast, high level granites from Archaean low-grade terrains tend dominantly to show negative Eu anomalies and have less pronounced curvature in their heavy REE distributions (*cf* Condie *et al.* 1986). Most post-Archaean high-level granites have negative Eu anomalies, which contribute to the marked negative Eu anomaly shown by sedimentary REE patterns (Taylor & McLennan 1981).

Rare earth distributions in gabbro–diorite–tonalite–trondhjemite complexes have been attributed to hornblende fractionation (Arth *et al.* 1977). However it is equally possible to reproduce the same REE patterns by differing degrees of partial melting of a mafic source with hornblende and/or garnet in the residue (Tarney *et al.* 1979; Weaver & Tarney 1980, 1981; Rollinson & Fowler, this vol.). Similar petrogenetic schemes have been proposed for other Archaean high-grade terrains in India (Condie & Allen 1984) NE China (Jahn & Zhang 1984) and East Antarctica (Sheraton & Black 1983; Sheraton & Collerson 1984). The significant petrogenetic point to be made is that the silicic gneisses must be generated under moderately high pressure conditions, in contrast with the low pressure

petrogenesis of the mafic gneisses. To explain the voluminous production of tonalitic magmas at this period in earth history, subduction melting of ocean crust is the only viable way of supplying the necessary thermal energy. The higher thermal regime in the Archaean would permit hydrous melting of the downgoing slab in a more active convecting mantle (whereas thermal models of present day subduction zones do not favour significant melting, and element transfer into the mantle wedge is accomplished during dehydration: Anderson *et al.* 1978). Melting would be possible at much shallower depths than at the present day. This provides an important constraint on the evolution of such magmas. Hydrous magmas cannot rise high in the crust because they rapidly congeal as pH_2O falls; hence they will tend to be emplaced into the deeper crust, and crustal growth is accomplished by underplating (*cf* Holland & Lambert 1975). A consequence of continued underplating of tonalitic magmas into the deeper crust is that the crust would remain ductile for long periods, thus being very susceptible to deformation. Mafic material emplaced tectonically into the deeper crust would undergo severe disruption, and the whole complex would develop gneissic characteristics. Moreover, as new crustal material is continually added from beneath, all crustal material in effect passes through a high-grade state (Weaver & Tarney 1980). Such a model does not however produce the thermal conditions for granulite-facies metamorphism.

Modern geochemical equivalents of Archaean tonalites and trondhjemites, with positive Eu anomalies and marked heavy REE depletion are relatively rare. Tonalites with very similar major element compositions are perhaps the dominant pluton in the Andean magmatic belt (*cf* Tarney 1976), but the majority display either no, or slight negative, Eu anomalies and few have the marked heavy REE depletion seen in the tonalites of Archaean high-grade terrains. The majority, of course, are high level plutons. Locally, however, deep sections of both the late Palaeozoic and Mesozoic batholith are exposed (Bartholomew & Tarney 1984) and include a proportion of tonalitic rock types with very similar REE characteristics—positive Eu anomalies and strong heavy REE depletion—to Archaean high-grade tonalitic gneisses (Fig. 2). Moreover such uplifted deep sections tend to be bimodal in nature and the rocks are variably foliated. This suggests that the processes of crustal generation at Cordilleran margins may produce broadly similar hydrous tonalitic magmas which are emplaced into the deeper crust, but which are only rarely exposed.

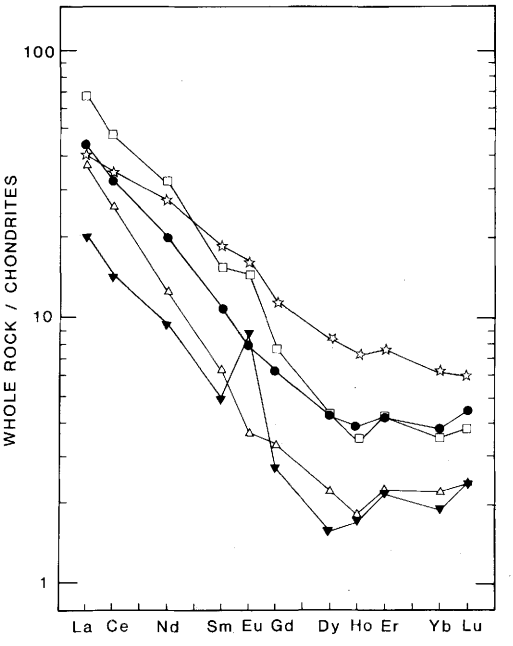

FIG. 2. Rare earth patterns for deep-seated tonalitic granitoids from the Mesozoic Patagonian batholith, Aysen, S. Chile (after Bartholomew & Tarney 1984).

Relationship between granulite-facies metamorphism and element depletion

The relationship between granulite-facies metamorphism and the commonly observed very low concentrations of selected lithophile elements in some (particularly Archaean) granulite terrains (Heier 1973) has been a subject of considerable debate. The issue is confused by two uncertainties. The first is whether granulite-facies metamorphism closely accompanies crustal generation and can be considered a continuation of the crust generating cycle, or whether granulite metamorphism is essentially a separate event which can occur at any time after crustal generation (see Ben Othman *et al.* 1984 and discussion by Moorbath 1984). The second concerns which particular elements are removed during granulite-facies metamorphism and—more important—what processes control this element removal. For instance some granulite terrains, such as the Madras charnockites (Weaver 1980) and the Jequie Complex, Brazil (Sighinolfi *et al.* 1981) show little evidence of removal of Rb, K, *etc.* nor the development of high K/Rb ratios. Studies of various granulite terrains (*eg* Heier & Thoresen 1971; Tarney *et al.* 1972; Barbey & Cuney 1982; Condie & Allen 1984; Jahn & Zhang 1984; Sheraton & Collerson 1984) show that they have

experienced various degrees of depletion of U, Th, Rb and K, relative to mean crustal values, a point emphasized by Rudnick *et al.* (1985). Additionally some amphibolite-facies terrains which have not apparently experienced granulite-facies metamorphism have suffered severe U depletion (Taylor *et al.* 1980; Moorbath & Taylor 1981), including segments of the Lewisian (Whitehouse & Moorbath 1986), and others such as the Gruinard gneisses (Rollinson & Fowler, this vol.) are notably Rb poor. This begs the question of whether granulite-facies metamorphism and element depletion are in fact intimately connected, or whether they are two separate processes which, because they relate to or affect deep crustal rocks, frequently coincide. This issue will be addressed further below.

Lewisian granulites have very low Rb, Th and U contents (Tarney *et al.* 1972; Moorbath *et al.* 1969), and have been ranked as the most extreme of all granulites in terms of their very high K/Rb ratios (Rudnick *et al.* 1985), which are well above those of normal igneous rocks (Shaw 1968). Although their major element compositions are broadly calc-alkaline (Sheraton *et al.* 1973) they differ from all calc-alkaline plutonic and volcanic suites in that there is very little change in either K or Rb concentrations between basic and silicic gneisses. Major differences between Lewisian amphibolite- and granulite-facies compositions are apparent for Rb, Th, U and K, and to a lesser extent for Pb, simply by comparing average analyses of the two terrains (Sheraton *et al.* 1973), but this is less sensitive for other elements because the granulite average is biased towards the more abundant basic components. Nonetheless a detailed comparison of element abundances in equivalent tonalitic rock types in the two terrains (Weaver & Tarney 1981) provided no significant additional evidence for major differences in other elements: Rb, Th and U are all more than an order of magnitude lower in the granulites and K and Pb about 50% lower (Fig. 3).

Currently there are two popular models which attempt to provide an explanation for these differences:

Partial melting

Classical models for granite production (*eg* Fyfe 1973) invoke extraction of granite from the lower crust (along with heat-producing elements) leaving a granulite-facies residue depleted in such elements. Extension of this process to crustal development in general (Taylor & McLennan 1981) provides a mechanism for generating upper crustal granites (and derived sediments) with a negative Eu anomaly, whilst the compensating

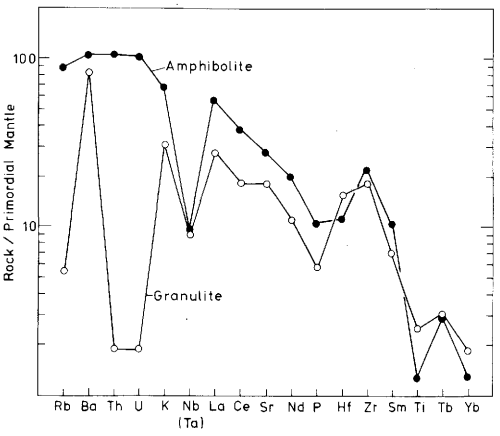

FIG. 3. Element abundances of Lewisian amphibolite-facies tonalites and granulite-facies tonalites compared on primordial mantle normalized multi-element plots (after Weaver & Tarney 1981).

positive Eu anomaly is carried in the plagioclase-rich basic residual granulites. Unfortunately this model is difficult to apply to the Archaean for several reasons: *a*) Archaean sediments in general do not have a negative Eu anomaly and, by implication, neither does the average Archaean upper crust, thus there is *a priori* no reason to extract a granitic melt, *b*) Whereas the Lewisian average granulite does display a positive Eu anomaly, in detail this is carried entirely by the silicic components, not the mafic gneisses (which tend to have negative Eu anomalies): thus melting of these silicic granitic components would provide liquids with positive Eu anomalies, *c*) Removal of granitic liquids would help deplete the residue in Th, U, Rb and K, but would be expected to produce other element fractionations, for instance of Ba relative to Sr; these are not observed, *d*) Significant melt extraction at the time of granulite-facies metamorphism would have severely perturbed the Sm–Nd isotope systematics established *c.* 250 Ma earlier at the time of crust generation; this has not happened.

For these and other reasons Weaver & Tarney (1981) concluded that major melt extraction was not the primary cause of the marked depletion in heat producing elements in Lewisian granulites. This is not to say that melting did not occur on a minor scale during the evolution of the granulite complex. There are numerous trondhjemitic veins, and Weaver & Tarney (1980) proposed that the minor intrusive anorthositic bodies which commonly occur in the Lewisian and other granulite terrains may have been generated through wet melting of the mafic granulites. Rock types with potentially higher aH_2O, such as the

metasediments, provide ample evidence of local melting (Cartwright & Barnicoat, this vol.).

Flushing by CO_2-rich fluids

This has become an increasingly popular model in recent years, inspired by the observation that fluid inclusions in granulites are CO_2-rich, and by the spectacular exposures in Kabbaldurga Quarry in southern India where granulite- and amphibolite-facies assemblages co-exist (*cf* Hansen *et al.* 1984). The essence of the model is that as CO_2-rich fluids penetrate the deeper crust from the mantle, the higher CO_2 activity will reduce aH_2O, thus promoting dehydration of hydrous minerals, and at the same time raising the rock solidus to inhibit melting. Elements released as the rock recrystallises are carried upwards by the rising CO_2 front. Whereas the evidence from Kabbaldurga seems beyond dispute as far as the amphibolite–granulite transition is concerned (Friend 1983), the efficacy of the carbonic ichors in element removal has yet to be fully verified experimentally, and their potential may easily be overestimated (*eg* Collerson & Fryer 1978). Geochemical studies at Kabbaldurga (Weaver & Tarney 1983*a*) demonstrated little change in K, Rb contents or K/Rb ratios for the silicic gneisses in the transition to granulite, although the associated basic rocks showed a trend toward high K/Rb ratios.

The carbonic fluid mechanism may inhibit melting, but it does not necessarily prevent it. For instance Weaver (1980) and Weaver & Tarney (1983*a*) proposed that the advancing CO_2 front would be preceded by a H_2O-rich front resulting from the breakdown of hydrous phases which would actually promote K-metasomatism and melting at intermediate crustal levels. Further advance of the CO_2-rich front would convert these K-rich rocks to granulites and account for the potassic nature of the Madras charnockites. This model has been extended by Friend (1983) and Condie *et al.* (1986) to account for the production of granitoid bodies such as the Clospet granite which appear to have been emplaced contemporaneously with charnockite formation. A possible equivalent in West Greenland might be the Qôrqut granite and the associated migmatite terrain.

While CO_2-flushing models may potentially be adequate to account for the conversion of gneisses to granulite facies and to permit local melting, this is not necessarily proof of their efficacy in accomplishing the depletion of the residual granulites in heat-producing elements. For instance there are no comparable granitoid bodies or K-rich granitoid zones in the Lewisian to complement the K-poor granulites. Moreover, Powell (1983) suggests that CO_2 inclusions in granulites might be a natural consequence of the partitioning of H_2O into a melt phase under internal buffering, and do not necessarily provide evidence of widespread CO_2 flushing. More important perhaps is that whereas CO_2 flushing might promote biotite dehydration and release of the Rb (and K) contained in biotite, there is no obvious reason why this should also promote the simultaneous breakdown of U- and Th-bearing phases. Nor indeed why some of these elements are lost at amphibolite facies. At the opposite end of the scale there is the problem of why, if CO_2 fluids are the sole agency of element removal and conversion of rocks to granulite facies under appropriate P–T conditions, some granulite terrains such as the Jequie Complex, Brazil (Sighinolfi *et al.* 1981) remain K- and Rb-rich.

Taking an entirely crustal perspective, the above two models of partial melting and fluid transfer are the only ones which are realistically capable of accounting for the production of deep crustal granulites depleted in heat-producing elements. Unfortunately neither fully explains all the available geological and geochemical data. Thus it is worth exploring the extent to which some of the geochemical characteristics of depleted granulites might be primary. This has not been seriously considered previously simply because volcanic rocks with the geochemical characteristics of Archaean granulites do not exist. A necessary consequence of a 'primary' model must therefore be that magmas with these characteristics are constrained by their physical mode of generation to the deeper continental crust: *ie* they must be underplated.

Primary nature of 'depleted' granulites?

Two factors cast some doubt on whether element depletion was achieved *in situ* during granulite-facies metamorphism itself: (1) whole rock Pb isotope data link the timing of major U loss from all Lewisian gneisses with the period of crustal generation rather than with the granulite facies peak P–T conditions some 250 Ma later—indeed the U contents of Lewisian zircons are abnormally low (Pidgeon & Bowes 1972); (2) although K/Rb ratios in Lewisian granulites of igneous parentage are uniformly high, the metasediments within the granulites have normal upper crustal K/Rb ratios of about 250 (Sheraton *et al.* 1973). Tarney *et al.* (1972) ascribed this to compositional and mineralogical control; recrystallization reactions leading to Rb and K loss can only take place in diopside-normative rock types (which includes the vast majority of orthogneisses) and are

inhibited in corundum-normative metasediments. This difference in K/Rb ratios could however also be interpreted as a primary feature.

Mineralogy exerts a vital control on rock chemistry during extensive fluid activity. For instance Beach & Tarney (1978) demonstrated that during hydrous retrogression of the Lewisian granulites the whole-rock compositions readjusted completely to the new mineralogy, implying element migrations over distances of in some cases above several metres. Minor mineral phases may play an important role, in particular with elements such as U and Th.

The important point to stress however, is that element fractionation resulting from fluid–rock interactions can operate at any one of the several stages involved in the evolution of a magma or its source. The generation of a tonalitic magma for instance, is a product of multistage processes: (1) mantle melting at an ocean ridge to form basalt; (2) hydrothermal activity at the ridge to produce a vertically-zoned series of hydrous assemblages; (3) dehydration during subduction of the ocean lithosphere when the expelled fluids may be in equilibrium with a progressive series of greenschist-, blueschist-, amphibolite- or eclogite-facies mineral assemblages, each with different minor mineral phases; (4) subduction zone melting, in equilibrium with possibly different major and minor phases. Loss of K, Rb, Th or U because of the absence of a suitable retentive mineral phase during any one of these stages could provide a suitable source for the generation of tonalites depleted in heat-producing elements. Unfortunately processes in subduction zones are not well constrained, so that it is difficult to provide more than a qualitative assessment of the likely processes.

During ridge hydrothermal activity there is intake of K and Rb from seawater in the upper layers of the ocean crust where clay minerals and zeolites are stable but, conversely, major loss of K, and particularly Rb, in the deeper layers where hydrothermal activity is more intense and where hornblende is stable. Hornblende accepts some K into its structure, but discriminates against Rb. This results in the deeper layers of ocean crust having very high K/Rb ratios (Sanders *et al.* 1979). However there appears to be a net gain of U by the ocean crust (Aumento 1979).

During subduction there is general agreement that this hydration process is reversed and that as dehydration occurs, selected LIL elements (Pearce 1983), especially Rb and K (Hole *et al.* 1984), are transferred from the subducted ocean crust into the overlying mantle wedge. This LIL element component may either be stabilized as mantle phlogopite or removed as calc-alkaline magmas. Clearly there may be a very great difference in the K and Rb contents and K/Rb ratios of magmas generated from melting of hydrothermally altered ocean crust, after this further stage of dehydration, compared with those in magmas generated from the mantle wedge.

Thermal models for present day subduction zones (Anderson *et al.* 1978, 1980) and experimental data for subduction zone melting under different thermal conditions (see review by Wyllie 1983) indicate that extensive melting of the subducted ocean crust is unlikely, particularly under continental regions, and that small-degree melts that might form after transformation of the ocean crust to eclogite would react and equilibrate with the overlying mantle wedge. They are water-undersaturated. Modern calc-alkaline magmas then are generated from the acceptor regions of the LIL-element enriched subduction component, and this is evident from their trace element patterns (Pearce 1983).

Melting of the donor regions (subducted ocean crust) is much more likely in the Archaean when mantle isotherms were considerably higher than today and, with smaller, more rapidly convecting plates, the subducting ocean crust was warmer. If the subducted slab induces convection in the overlying mantle wedge, the thermal conditions for melting of the ocean crust are even more easily attained. Melts would be water-saturated and generated in equilibrium with hornblende and perhaps residual garnet and, if generated at shallow levels, could be quite silicic. If, as suggested by Green (1975), thermal gradients in the Archaean were too high for the formation of dense eclogite in subducted ocean crust, warm ocean crust would be subducted at a shallow angle. One effect of this would be to extend the zone over which melting with residual hornblende occurs, thus leading to more voluminous magma production under these conditions. Another would be that shallow melting provides less opportunity for magmas to react with the overlying mantle, thus they are more likely to remain water-saturated. Water-saturated magmas would be unable to rise to high crustal levels and would congeal in the deep crust.

Conversely, with a cooler subducting plate, the Benioff zone would be steeper, with melting displaced to deeper levels. Magmas generated from a more steeply inclined subduction zone have more opportunity to react with the higher temperature mantle wedge, take up the LIL element subduction component already transferred to the wedge and, being water-undersaturated, rise to higher crustal levels. This model

implies a general correlation between the depth of magma generation and the height to which the magma can eventually rise in the crust. This is dominantly controlled by the degree of water saturation, but is also compatible with the general hydrostatic principle that the height to which a magma column can rise is dependent on the supporting weight of crust + mantle. Modern calc-alkaline magmas, being generated at greater depths and from the mantle wedge, are mostly water-unsaturated and can rise to high crustal levels.

Whereas the above model provides a conceivable explanation of why K- and Rb-poor tonalitic magmas should dominantly be emplaced in the deep crust, the explanation for their low U and Th contents needs further consideration. It may be that during dehydration processes in the subducting slab they are also transferred to the mantle wedge, hence melting of the ocean crust after dehydration would produce magmas low in U and Th (though if Pb had also been removed during dehydration, U/Pb and Th/Pb ratios may not have been seriously affected). Much however will depend on the nature of the minor mineral phases (*eg* sphene, epidote, apatite, zircon) stable during the dehydration stage and also the melting stage. Sphene is stable well into the melting range of hydrous tholeiite compositions (Hellman & Green 1979) as is zoisite. Both sphene and (at higher pressures) rutile are important sinks for high field-strength elements such as Nb and Ta, and the fact that all subduction-zone magmas (and the continental crust generally, irrespective of age—Weaver & Tarney (1985)) display marked negative anomalies for these two elements on mantle-normalized multi-element diagrams (Weaver & Tarney 1983*b*) demonstrates the important role they play in subduction processes. If U and Th can be retained in minor mineral phases in the same way during melting of amphibolite in shallow subduction zones, it might begin to account for the low U and Th contents of high-grade gneisses. Unfortunately experimental data are lacking.

Discussion

Past discussions of the formation of the Scourian gneiss complex have centred on four main issues: *a*) the petrogenesis of the gneisses, *b*) their tectonic development, *c*) the age and causes of the granulite-facies metamorphism, and *d*) the reasons for the anomalously low concentrations of Rb, U, Th and K in the granulites. The traditional approach of linking *d*) with *c*) is not without problems, particularly in accounting for the pervasive manner in which U, Th, Rb and K have been removed from the granulites, either through melting or fluid transport (later fluid activity affecting the granulites for instance is extensive but very localized). The alternative approach of linking *d*) with *a*) as a primary characteristic has potential, though uncertainties regarding Archaean subduction processes leave it poorly constrained. The main feature of the model is that, in contrast to modern subduction-zone magmas which have their source in the mantle wedge (enhanced in LIL elements through fluid transfer), Archaean tonalitic magmas are derived from melting of ocean crust amphibolite which has suffered complementary loss of LIL elements.

Essentially the process of loss of heat-producing elements from deep crustal rocks is transferred from the granulite-facies event to an earlier stage. This reduces the need to appeal to melt removal from granulites or super-active carbonic fluids pervading the Lewisian deep crust, for which there is meagre evidence. On the other hand this does not mean that further loss of Rb, Th, or U did not occur with the expulsion of fluids during granulite-facies metamorphism. Sheraton (1984) has described the chemical changes in mafic dykes which cut high-grade gneisses and are metamorphosed at amphibolite- or granulite-facies in East Antarctica. The effects are variable. Dykes which were metamorphosed directly to granulite facies show no chemical changes. Most compositional change was seen in dykes which underwent major changes in mineralogy and where there was associated deformation to enhance fluid activity (see also Beach & Tarney 1978). In this regard it should be noted that a subduction zone is an ideal environment for fluid transport; there is continual deformation and recrystallization. In the Lewisian granulites there was extensive deformation before the granulite event and many of the tectonically emplaced mafic gneisses have obviously equilibrated with the silicic gneisses in terms of K/Rb ratios.

If the depletion in U, Th and Rb can be ascribed to subduction rather than crustal processes, there are a number of interesting secondary consequences. First, there is no reason why the deeper crust under many Archaean cratons should not be composed of similar U-, Th- and Rb-depleted material, which will represent a substantial reservoir of low U/Pb, low Th/Pb and low Rb/Sr material with unradiogenic Pb and Sr. Second, the high U/Pb and high Th/Pb complement to this is not necessarily in the upper crust but, if U and Th are stabilized in minor mineral phases (along with Nb and Ta) in the subducted ocean crust, this component is returned to the

mantle. Ocean island mantle sources are the only obvious complement to the lower crust in this regard: they have radiogenic Pb compositions which imply that U/Pb ratios of their sources were increased over normal mantle values at least 1.7 Ga ago (Sun 1980); additionally they have distinct positive Nb and Ta anomalies (Wood *et al.* 1981).

Finally, Archaean high-grade terrains have a distinctive tectonic style marked by strong horizontal thrusting and deformation (Bridgwater *et al.* 1974); indeed in Lewisian granulites the foliation is sub-horizontal over large areas (Sheraton *et al.* 1973), which contrasts with the widespread development of sub-vertical shear zones in the Proterozoic. A ductile lower crust is inducive to mobile horizontal tectonics; this would be facilitated by continual underplating of hydrous tonalitic magmas. However, once thermal gradients fall to the extent that voluminous melting of ocean crust is no longer possible, magma generation is restricted to the mantle wedge and is necessarily much less voluminous. Melts produced are then water-undersaturated and can rise to higher crustal levels. The crust adopts a more stable configuration; the permobile regime is at an end. This marks the Archaean–Proterozoic boundary, and is characterized by distinct compositional changes in the sedimentary record (Taylor & McLennan 1981) which largely reflect this change in mode of magma production.

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