
Geochemical Constraints on the Evolution of the Early Continental Crust [and Discussion]

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Geochemical constraints on the evolution of the early continental crust

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The most important process affecting both major and trace-element concentrations in the mantle and crust is melting producing silicate liquids which then migrate. Another process whose effects are becoming more apparent is the transport of elements by CO₂- and H₂O-rich fluids. Due to the relatively small amounts of fluids involved they have but little effect on the major-element abundances but may severely affect minor- and trace-element abundances in their source and the material through which they travel.

The Archaean crust was a density filter which reduced the possibility of komatiite or high FeO melts with relative densities greater than about 3.0 from reaching the surface. Those melts retained in the lower crust or at the crust–mantle boundary would have enhanced the possibility of melting in the lower crust. The high FeO melts may have included the Archaean equivalents of alkali basalt whose derivatives may form an important component in the Archaean crust. The occurrence of ultramafic to basic to alkaline magmas in some Archaean greenstone belts is an assemblage most typical of modern ocean-island suites in continental environments. The rock types in the assemblage were modified by conditions of higher heat production during the Archaean and thus greater extents of melting and melting at greater depths. If modern ocean-island suites are associated with mantle plumes, which even now may be an important way to transport heat upward from the deeper mantle, it is suggested that during the Archaean mantle plumes were an important factor in the evolution of the continental crust.

It appears that the Archaean continental crust was of comparable thickness to that of the present based on geobarometric data. If the freeboard concept applied then, this would suggest that plate tectonics was also an active process during the Archaean. If so, it is probably no more realistic to assume that all Archaean greenstone belts had a similar tectonic setting than to assume that all modern occurrences of basic rocks have a common tectonic setting.

INTRODUCTION

In discussing geochemical constraints on the evolution of the early continental crust, this paper will place most emphasis on the processes occurring within the mantle and crust. In table 1 are presented estimated abundances of selected elements for the primitive mantle and upper continental crust. It can be seen that the relatively incompatible elements such as K, Rb, Ba and U are all strongly enriched to approximately the same extent, whereas the more compatible elements such as the rare-earth elements (r.e.e.) are less strongly enriched, with the light r.e.e. (Ce) more enriched than the heavy r.e.e. (Yb). Table 2 gives selected trace element ratios for primitive mantle, upper continental crust and normal ocean-ridge basalts. Normal ocean-ridge basalt is the basalt that is depleted in the more incompatible elements, suggesting that its mantle source has undergone previous loss of these elements. From tables 1 and 2 it can be seen that to a first approximation those elements that are more enriched in the crust are also more depleted in the source for ocean-ridge basalts, suggesting that similar processes may be responsible.

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If heterogeneous accretion or condensation of a silicate atmosphere were important factors in the development of the present continental crust, the volatile elements should be more strongly enriched than the less volatile. Thus, the alkalis might be expected to be more enriched than higher-temperature condensates such as Ba and U. Yet these elements are all enriched to approximately the same extent. This suggests that heterogeneous accretion or a silicate atmosphere are not major factors in the origin or evolution of the present upper continental crust.

TABLE 1. ELEMENT CONCENTRATIONS IN PRIMITIVE MANTLE† AND UPPER CONTINENTAL CRUST‡ (The primitive mantle has concentrations of about twice that of chondrites for the lithophile, early condensate elements such as Ba, rare-earth elements, Sr, etc.)

	primitive mantle $\mu\text{g g}^{-1}$	upper continental crust $\mu\text{g g}^{-1}$	enrichment factor
Rb	0.67	110	164
K	240	27 400	114
Ba	7.1	700	99
U	0.026	2.5	96
Ce	1.64	64	39
Nb	0.79	25	32
Zr	13	240	18
Sr	22.1	350	16
Yb	0.413	2.2	5.3

† Sun & Nesbitt (1977).

‡ Taylor & McLennan (this symposium).

TABLE 2. ELEMENT RATIOS IN PRIMITIVE MANTLE†, UPPER CONTINENTAL CRUST‡ AND NORMAL OCEAN-RIDGE BASALT (o.r.b.)†

	primitive mantle	upper crust	normal o.r.b.
K/Rb	360	249	1060
K/Ba	34	39	88
K/U	9200	10960	10600
Rb/Sr	0.030	0.31	0.0081
K/Zr	18.5	114	10.6
Zr/Nb	16.5	9.6	32.3

† Sun & Nesbitt (1977).

‡ Taylor & McLennan (this symposium).

The existence of a magma ocean during differentiation would presumably form an ultramafic or komatiitic crust with the possible rafting of plagioclase. However, early mafic, ultramafic or anorthositic crusts do not appear to be major components of our present continental crust (Taylor 1979). Thus it would appear that the major geochemical features of the crust have resulted since differentiation. The most evident process affecting the earth since then is convection of the mantle and its concomitant melting, resulting in the removal of heat from the mantle and core and also the continuous removal of the heat-producing components K, U, and Th from the mantle and their emplacement into the crust.

To place limits on the evolution of the early continental crust this paper will consider the sources of some of the major components making up the Archaean continental crust and the processes involved. There appear to be two types of Archaean terrane, the greenstone belts and

gneiss belts. The greenstone belts consist predominantly of volcanics and immature sediments which are intruded by granitoid rocks, the belts having formed over a period of 10s to 100s of millions of years. The gneiss terranes consist predominantly of deformed as well as undeformed plutonic granitoid rocks with intercalated supracrustal sequences. The rocks making up the gneiss terranes have generally formed over periods of 100s to 1000s of millions of years. Because the mantle is probably the ultimate source of the present crust, it is important to recognize the types of material that may have been derived from the mantle and the processes that result in the transfer of material from the mantle to the crust.

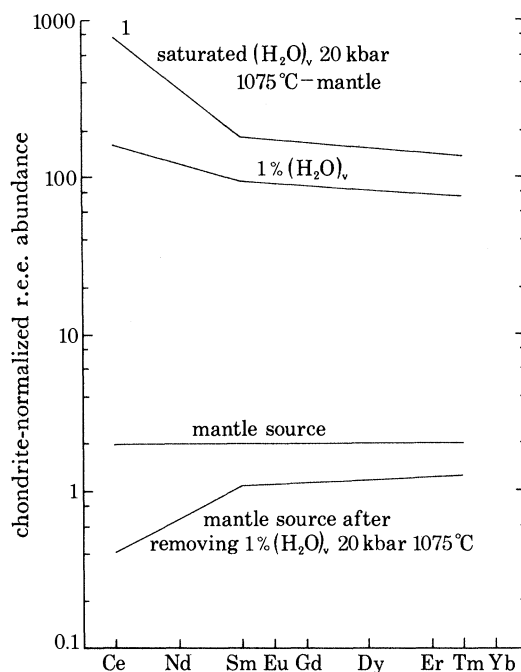


FIGURE 1. Calculated chondrite-normalized r.e.e. patterns for water vapour at 1075 °C, 20 kbar, in equilibrium with a lherzolithic mantle consisting of 55% olivine, 25% orthopyroxene and 20% clinopyroxene and with an original concentration of the r.e.e. of two times chondritic. The 'saturated $(\text{H}_2\text{O})_v$ ' curve is the concentration of the r.e.e. in an infinitesimal fraction of H_2O in equilibrium with the mantle. The '1% $(\text{H}_2\text{O})_v$ ' curve is the concentration of the r.e.e. in water representing 1% of the mass of the mantle source. The r.e.e. concentration is shown before and after removing 1% H_2O from the mantle source. Distribution coefficients are selected from Mysen (1978).

In considering possible processes the emphasis will be on the use of the trace elements, principally the r.e.e., as petrogenetic indicators of igneous rocks. The r.e.e. are particularly useful because: (1) they are geochemically very similar; (2) variations in their relative abundances can be used to place limits on the parent r.e.e. composition, the minerals present during melting or crystallization and the processes involved; (3) distribution coefficients are available for mathematical modelling; and (4) numerous high-precision data are available. For the model calculations presented here the distribution coefficients for the appropriate conditions are those given in Hanson (1980a).

There are two main ways that matter may be transferred from the mantle to the continental crust: (1) as silicate melts; and (2) as fluids. Transfer by silicate melts will affect major-, minor- and trace-element abundances in the mantle source and crust. Transfer by migrating fluids,

however, will selectively affect the minor and trace elements and eventually radiogenic isotope ratios because the radiogenic isotopes and their parents generally occur as trace elements. The reason fluids may affect mainly minor and trace elements is because fluids form such a small fraction of the source that they may not significantly affect the major-element abundances even though the major elements may have high concentrations in the fluids. Those trace elements that are strongly enriched in a fluid, however, may be significantly reduced in the source and enriched in the crust because they have such low original abundances in the mantle and crust.

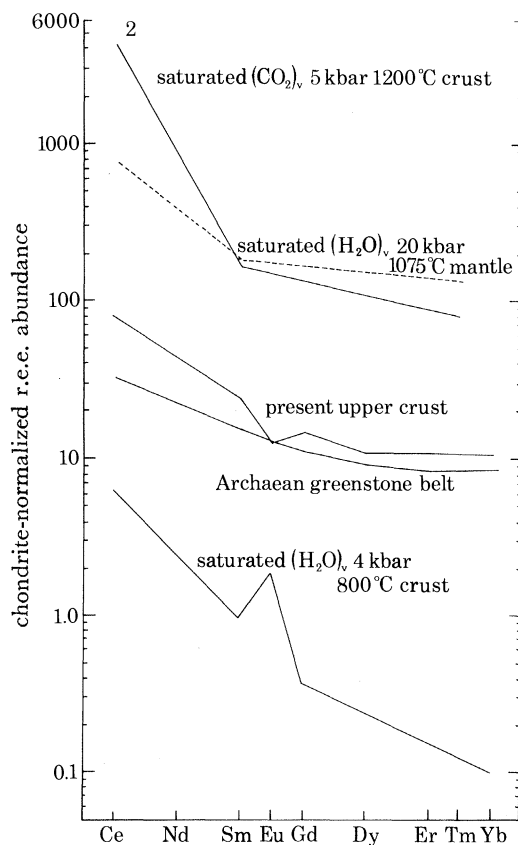


FIGURE 2. Calculated chondrite-normalized r.e.e. patterns for H_2O vapour and CO_2 vapour in equilibrium with an average greenstone-belt composition leaving a granulite residue with 40% feldspar, 40% clinopyroxene, 10% orthopyroxene and 10% quartz (all percentages by mass). For comparison 'saturated $(\text{H}_2\text{O})_v$ ' from figure 1 is also shown. Removal of 1% $(\text{H}_2\text{O})_v$ will increase the concentration of each of the r.e.e. of the average Archaean greenstone-belt composition by approximately 1%. Distribution coefficients are from Mysen (1978) and Wendlandt & Harrison (1979). Present upper crust and Archaean greenstone-belt data are from Taylor & McLennan, this symposium.

As an example to show the possible effect of fluids on trace elements, figure 1 shows the relative concentration of r.e.e. in an infinitesimal amount of water (saturated $(\text{H}_2\text{O})_v$) and 1% water vapour in equilibrium with a mantle with a r.e.e. abundance of two times chondrites at 20 kbar† and 1075 °C. These calculations are based on selected silicate melt – fluid distribution coefficients from Mysen (1978) and Wendlandt & Harrison (1979). Also shown is the effect on the r.e.e. abundance of the mantle of batch removal of 1% water vapour from the mantle. It

† 1 kbar = 10^8 bar = 10^8 Pa.

can be seen that 1% water will remove 75% of the Ce, leading to a strong depletion of the light r.e.e. in the source. If the water were removed incrementally in infinitesimally small steps, rather than in one step, the same effect on the Ce concentrations in the mantle will result after the removal of 0.36% water.

Figure 2 shows the relative concentration of the r.e.e. in an infinitesimal fraction of water at 20 kbar, 1075 °C, in equilibrium with the mantle, from figure 1, and in an infinitesimal fraction of water in equilibrium with an average greenstone belt composition at 4 kbar, 800 °C, based on data from Flynn & Burnham (1978). Thus water vapour at 20 kbar in the mantle may have a much greater concentration of the r.e.e. than has water vapour in the crust, so that water vapour released from the mantle at 20 kbar could enrich the overlying mantle or the crust. To determine whether the overlying mantle or crust is enriched it is necessary to know the amount of H₂O vapour removed from the mantle source and how the water vapour – rock distribution coefficients for the trace elements considered vary as a function of temperature (T), pressure (P) and X . If the water vapour – rock distribution coefficients decrease approximately linearly from 20 to 4 kbar for the conditions given, and an infinitesimal fraction of water vapour is removed from the original mantle source, then as the water vapour rises both the overlying mantle and the crust will be enriched. If, however, for example, 1% water vapour is removed as a batch from the original mantle source and the distribution coefficients decrease linearly, the migrating fluid will move upward and deplete the overlying mantle of the trace elements in a zone-refining process. This depletion will continue until the fluid has become so enriched in trace elements that under the P , T conditions given the fluid has a concentration of the trace elements equivalent to that of an infinitesimally small fraction of fluid under the same conditions, i.e. it is saturated. The smaller the fraction of original fluid relative to the source the sooner the fluid will be enriched to saturation. The fluid will then continuously enrich the remaining overlying mantle or crust, because for this example the water vapour – rock distribution coefficients are decreasing with lower T and P . So, depending on (1) how the distribution coefficients vary as a function of T , P and X , (2) the fraction of fluid removed from the original source, and (3) the chemical abundances and mineralogy of the source and overlying mantle and crust, some elements may be removed from the overlying mantle or crust by the migrating fluid, while other elements are being deposited.

Also in figure 2 is a plot of the concentration of r.e.e. in an infinitesimally small fraction of CO₂ vapour at 5 kbar, 1200 °C, in equilibrium with an average greenstone-composition crust. The ‘saturated’ CO₂ vapour has very large concentrations of the r.e.e. relative to the crust. Removal of 1% CO₂ from the crust should strongly deplete it in the r.e.e. and lead to a strong depletion of the light r.e.e. If, like water, CO₂ had greater concentrations of r.e.e. in equilibrium with a mantle composition as compared to a crustal composition, CO₂ vapour could potentially be very effective at transporting r.e.e. from a degassing mantle to the overlying mantle and crust.

Fluids may play an important role in transporting trace elements from the mantle to the overlying mantle and crust. Migration of fluids in the mantle may also explain metasomatic effects observed in lherzolite nodules derived from the subcontinental mantle which have secondary amphibole and phlogopite and which have apparently been enriched in Ti, Al, Fe, Mn, Ca, Na, K, H₂O, CO₂, Rb, Sr, Y, Zr, Nb, Ba and La (see, for example, Lloyd & Bailey 1975). Fluid migration may also be important in depleting the mantle sources of normal ocean-ridge basalts and would help explain why some mantle sources that are strongly depleted in the more incompatible elements (table 2) are apparently enriched in basaltic components (Langmuir &

Hanson 1980). To effectively evaluate the role of fluids more experimental data on distribution coefficients for trace elements between silicate minerals or melts and fluids must be obtained.

There is little doubt that silicate melts must play an important role in transporting major, minor and trace elements from the mantle to the crust. The principle sites for melting of the mantle at present are: (1) spreading centres, where melting most likely is associated with the adiabatic uplift of the mantle that is filling a void left by spreading; (2) zones of subduction, where possible melting of the mantle is a result of interaction with a down-going slab which is itself probably degassing and melting; and (3) ocean islands and continental rift zones associated with hot spots or mantle plumes. It would seem that during the Archaean equivalent régimes modified by the greater production of radiogenic heat may also have been major sources of magmas.

PETROGENESIS OF ROCKS IN GREENSTONE BELTS

One of the interesting features of the continental crust is that while it appears, based on Nd isotope ratios in oceanic basalts, that the mantle has had on average an essentially chondritic to slightly light-r.e.e.-depleted abundance, the crust is significantly enriched in the light r.e.e. (see Taylor & McLennan, this symposium). The reason the enrichment of the light r.e.e. is interesting is that the melting of a mantle with chondritic r.e.e. ratios leads to significant enrichment of the light r.e.e. only if very low extents of melting are involved (less than about 5%; see, for example, Hanson 1980*a*). Phanerozoic melts, which are strongly enriched in the light r.e.e. and presumably represent small fractions of melting of the mantle (alkali basalt, nephelinite, kimberlite etc.), are generally silica-undersaturated, whereas the upper continental crust is oversaturated in silica. Also these rock types appear to be relatively less abundant than those derived by larger extents of melting, such as tholeiitic basalt. In Archaean greenstone belts, one gets the impression that rocks representing lower extents of melting of the mantle are rare while melts representing larger extents of melting, such as tholeiites and komatiites, are more common. Thus it would appear that the continental crust must include a large fraction that is a result of reworking of earlier mantle-derived material. Because the greenstone-belt assemblages include the majority of those rocks that may be considered as having been directly derived from the Archaean mantle, it would seem reasonable that those rocks may be considered as likely candidates for the parental material of the continental crust. The r.e.e. as well as other trace-element abundances and isotope ratios in rocks within Archaean greenstone belts have been reviewed by Jahn & Sun (1979).

The major rock types preserved in greenstone belts, which may have formed by melting of the Archaean mantle, are peridotitic komatiites, basaltic komatiites, and tholeiites (selected r.e.e. patterns are shown in figure 3). The tholeiites have essentially flat r.e.e. patterns, basaltic komatiites are enriched to slightly depleted in light r.e.e., and peridotitic komatiites have flat to light-r.e.e.-depleted patterns. Nd isotope data for peridotitic komatiites and tholeiites (O'Nions *et al.* 1979) suggest that their mantle source has had a long-term Sm/Nd ratio that is chondritic. This suggests that the light-r.e.e. depletion expected for the peridotitic komatiite source is a short-lived phenomenon, perhaps related to the processes involved in melting (Arth *et al.* 1977; Sun & Nesbitt 1978). Arndt (1977) and Hanson & Langmuir (1978) suggest that, in an adiabatically uprising mantle, melting and melt removal occurred continuously and to different extents, a process called dynamic melting by Langmuir *et al.* (1977). This process may lead to melts derived by large extents of melting that are light-r.e.e.-depleted relative to their

parent. Because the composition of the parent is becoming more refractory during melting, the later derived melts have chemical characteristics suggesting up to 60% melting of the mantle, whereas only 30% melting is necessary.

The basaltic komatiites could be explained as a result of smaller extents of melting leaving garnet in the residue, which leads to heavy-r.e.e. depletion. They might be considered to be Archaean equivalents of alkali basalts except that their high FeO and low Al_2O_3 abundances suggest that they were derived from a much greater depth than were modern alkali basalts (Langmuir & Hanson 1980).

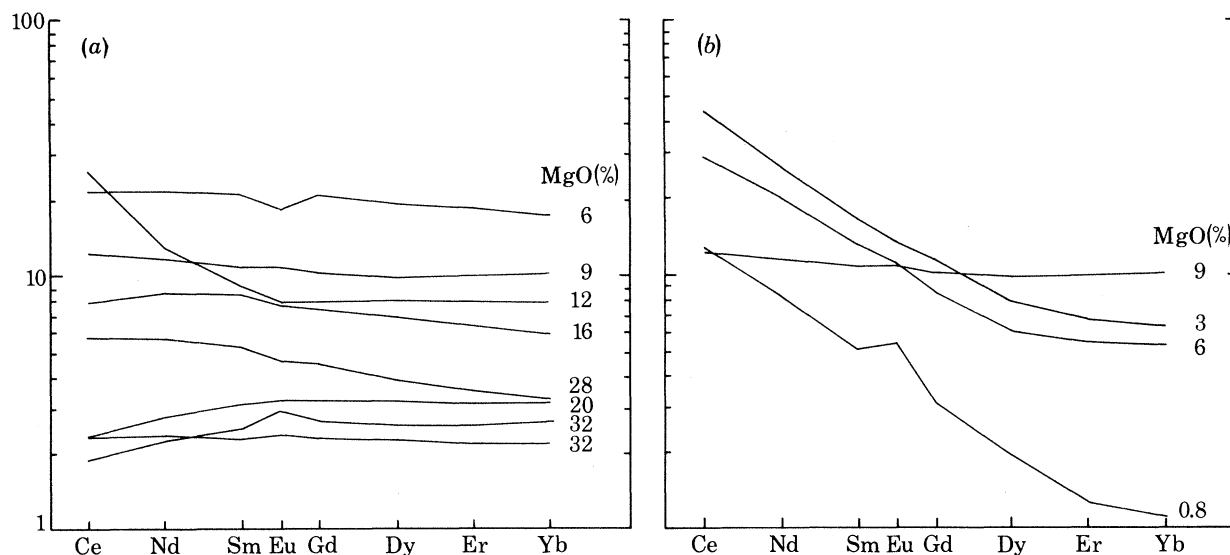


FIGURE 3. (a) Chondrite-normalized r.e.e. pattern for Archaean peridotitic komatiites (MgO 20–32% (by mass)), basaltic komatiites (MgO 12 and 16% (by mass)) and tholeiites (MgO 6 and 9% (by mass)). Data are from Sun & Nesbitt (1978) and Arth & Hanson (1975). (b) Chondrite-normalized r.e.e. patterns for calc-alkaline volcanics from Hawksworth & O’Nions (1977) and the tholeiite from figure 3a (Arth & Hanson 1975).

While some tholeiites may be differentiates of a komatiitic parent, the essentially flat r.e.e. pattern cannot be explained by differentiation of a light-r.e.e.-depleted peridotitic komatiite involving principally olivine and pyroxene.

A sequence of igneous rocks that has not received much attention in the literature occurs as mafic tonalites, hornbländites, trachyandesites, lamprophyres, syenites, monzonites syenodiorites, granodiorites and diorites (Arth & Hanson 1975; Schultz *et al.* 1979). These alkalic to potassic igneous rocks are characterized by high Sr and Ba abundances (1000–3000 $\mu\text{g g}^{-1}$), low Rb/Sr ratios (generally < 0.2) and high light- to heavy-r.e.e. ratios (figure 4). These are generally silica undersaturated on a normative basis, but may have modal quartz because of the occurrence of relatively large fractions (10 to 20%) of hornblende. These rock types generally occur as late to post-kinematic dykes or plutons. In the Vermilion District, northeastern Minnesota, for example, the Giants Range Granite east of Virginia is mainly a rock with this character (Sims & Viswanathan 1972) and has a surface area equivalent to or greater than that of the basic and mafic volcanics in the greenstone belt that it intrudes. In the district there are also a number of smaller alkalic bodies of a similar nature, e.g. the Linden Syenite, the Snowbank stock and the Icarus pluton (Sims & Mudrey 1972; Arth & Hanson

1975). The Sr and Pb isotope data available for these rock types, although limited, are consistent with their direct derivation from the mantle (Hanson *et al.* 1971; Arth & Hanson 1975; Schultz *et al.* 1979). These alkalic rocks have some similarities to modern alkali basalts and their derivatives (see, for example: Zielenski & Frey 1970; Kay & Gast 1973; Schwarzer & Rogers 1974). They are also very similar to rock types in the Cortlandt Complex, New York State, which include hornblendites, diorites, gabbros, norites and granodiorites, and are best explained as alkali basalts and their derivatives that have undergone various extents of interaction with continental rocks (Bender *et al.* 1979).

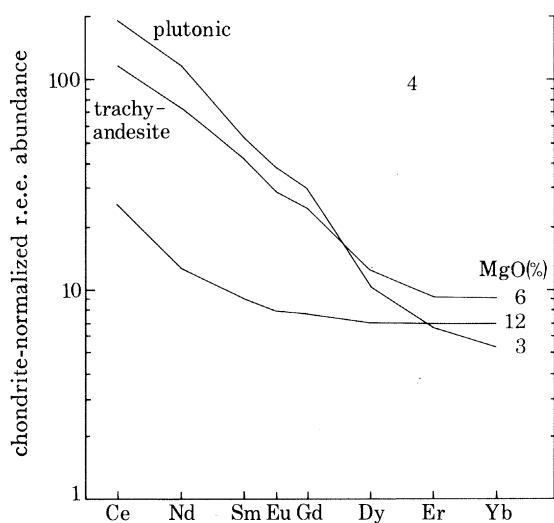


FIGURE 4.

FIGURE 4. Chondrite-normalized r.e.e. patterns for alkalic rocks from Arth & Hanson (1975) and a basaltic komatiite (MgO 12% (by mass)) from figure 3a (Sun & Nesbitt 1978).

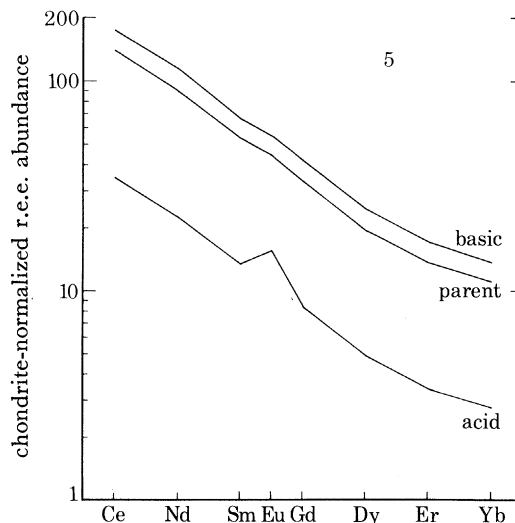


FIGURE 5.

FIGURE 5. Calculated chondrite-normalized r.e.e. patterns for immiscible alkaline liquids on the assumption that the basic melt – acidic melt distribution coefficient is 5 for all of the trivalent r.e.e. and 3.5 for Eu (Watson 1976; Ryerson & Hess 1978). Of the parent 25% (by mass) is acid melt, 75% (by mass) basic melt. Note that while the acid melt develops a positive Eu anomaly the corresponding negative Eu anomaly in the basic melt is imperceptible.

Fyfe (1978) noted that komatiitic melts may have relative densities of 3.0 or greater and thus such melts would have a difficult time invading rock sequences with lower densities without an appropriate hydrostatic head. Densities for silicate melts at 1 atm† can be calculated from the data of Bottinga & Weill (1970) over the range of 1250–1600 °C. As can be seen from table 3, of the major oxide components FeO has by far the largest effect on increasing the density of a melt. Oxidation of FeO to Fe₂O₃, however, will lead to a large reduction in the effect of iron content on the density of a melt. SiO₂, Na₂O, K₂O, and H₂O have the largest effect in reducing the density of a melt. Peridotitic komatiite melts with relatively low SiO₂, Na₂O, and K₂O, probably low *f*_{O₂} and high MgO and FeO will have densities of about 3.0 g/cm³. If the pressure effects on density described by Kushiro *et al.* (1976) are appropriate, a peridotitic komatiite melt at the base of a crustal sequence could have a density of about 3.10 g/cm³, more than for most crustal rocks. Similar values for the densities of melts are found for the most

† 1 atm = 101325 Pa.

iron-rich basalt komatiites and tholeiites found in greenstone belts. This would suggest that melts with higher FeO contents, representing lower extents of melting (see Hanson & Langmuir 1978, fig. 8), may have existed as partial melts of the Archaean mantle, but are probably not found because they were too dense to reach crustal levels. Stolper *et al.* (1981) suggest that dry basic magmas may even become denser than their residual peridotite at some pressure.

TABLE 3. PARTIAL MOLAR DENSITIES (M_i/v_i) FOR OXIDE COMPONENTS IN SILICATE MELTS AT 1300 °C AND 1 atm

(From Bottinga & Weill (1970).)

	$\frac{M_i/v_i}{\text{g/cm}^3}$		$\frac{M_i/v_i}{\text{g/cm}^3}$
SiO ₂	2.24	TiO ₂	3.89
Al ₂ O ₃	2.69	MgO	3.56
FeO	5.70	CaO	3.46
Fe ₂ O ₃	3.06	Na ₂ O	2.20
MnO	5.18	K ₂ O	2.11

TABLE 4. SELECTED BASIC LIQUID – ACIDIC LIQUID DISTRIBUTION COEFFICIENTS FOR IMMISCIBLE LIQUIDS

(From Watson (1976) and Ryerson & Hess (1978).)

element	basic/acidic	element	basic/acidic
K	0.5	Ca	3
Na	0.5	Mn	3
Ba	1.5	Ti	3
Sr	1.5	Fe	3
Mg	2	r.e.e.	5
Zr	2.5	P	10

Iron-rich melts representing low extents of melting could be the parents of alkalic and potassic sequences that are a result of differentiation and interaction with crustal rocks deeper in the crust or at the crust–mantle boundary. The possibility of large volumes of melts accumulating in the mantle even in more recent times is inferred from studies of nodules in alkali basalts and kimberlites which suggest that the upper mantle includes magmatic pyroxenites and dunites with large cumulate bodies occurring near the crust–mantle boundary (Mercier 1977). If large volumes of dense, iron-rich alkalic melt could have existed at or near the crust–mantle boundary, large-scale liquid immiscibility may have been an important process. This is because alkalic melts are susceptible to immiscibility, and increased iron contents and increased pressure enhance the possibility of immiscible liquids forming (Roedder 1978; Philpotts 1976; Watson 1976; Bender *et al.* 1980; Naslund 1976; Watson & Naslund 1977; Visser & Koster Van Groos 1979).

Such immiscible liquids would consist of basic and acid liquids (selected element coefficients are shown in table 4). The basic liquid would most likely be silica-undersaturated, and significantly enriched in iron, phosphorous, r.e.e., Ca, Mg, the transition metals and other high-field-strength elements. The acid liquid would be silica oversaturated and would have potassium and sodium contents similar to or greater than that of the parent. Thus for an alkalic parent the acid liquid may be a granodiorite or granite with a r.e.e. pattern subparallel to that of alkali basalt but with a lower abundance (figure 5). The actual abundance will depend on

the extent of differentiation before separation of the liquid and the fraction of acid liquid formed. If the acid liquid is an infinitesimal fraction of the parent, it will have a subparallel r.e.e. pattern with approximately one-fifth of the total r.e.e. abundance of the parent or total liquid just before separation. If it represents a larger fraction, the r.e.e. abundances will approach those of the parent liquid. If there is significant Eu^{2+} , the acid liquid will have, relative to the basic liquid, a positive Eu anomaly whose extent will be a function of the $\text{Eu}^{2+}/\text{Eu}^{3+}$ ratio. Due to its relatively low specific gravity the acid liquid would probably

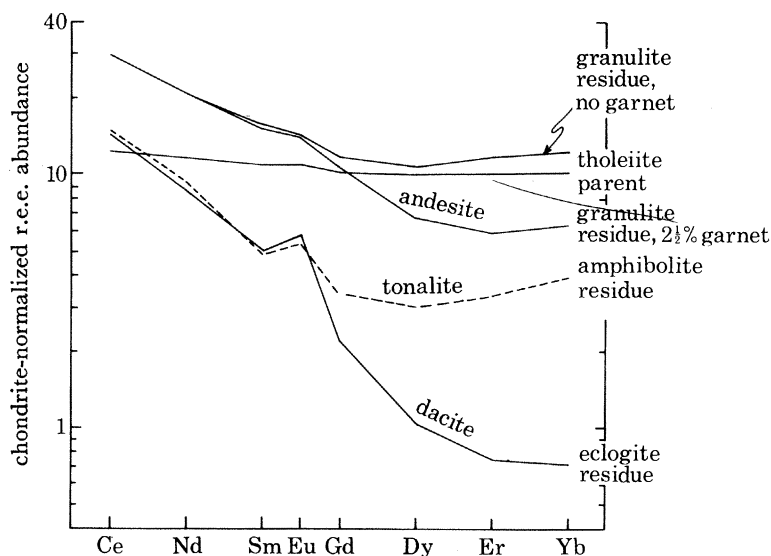


FIGURE 6. Calculated chondrite-normalized r.e.e. patterns for melts representing 20% melting of a tholeiite parent (MgO 9% (by mass), from figure 3*a*) leaving granulite residue with and without garnet, amphibolite residue and an eclogite residue (distribution coefficient data are given in Hanson (1980*b*)).

separate very rapidly, and because it would be relatively dry it would probably occur as shallow intrusions or extrusions. It might be difficult to distinguish from granitic rocks derived by partial melting of greywacke or lower crust, as they would all have light-r.e.e.-enriched patterns with low heavy-r.e.e. contents, of the order of two to six times chondritic. However, low total r.e.e. abundance, positive Eu anomalies, and relatively low abundances of high-field-strength elements such as P, Zr, Nb in acid rocks would be indicative of immiscibility (Bender *et al.* 1980). The low total r.e.e. abundance, positive Eu anomaly and granitic composition could also be similar to rocks derived by precipitation from aqueous fluids at high grades of metamorphism (see figure 2).

In greenstone-belt sequences peridotitic komatiites seem to predominate near the base of the sequence. Higher sections generally show a predominance of tholeiites, dacites and volcanogenic sediments. It is suggested that komatiites could be forming throughout the lifetime of a greenstone belt, but that later komatiites and iron-rich basalts would have formed intrusions near the base of the crust or greenstone-belt assemblage. Komatiitic or basaltic melts with lower densities could form intrusions at sediment–greenstone interfaces or occur as volcanics. These melts might have been able to supply enough heat to enhance the possibility of melting within the thickening greenstone belt. The most likely rocks to undergo melting would be thick tholeiite sequences, sediment–tuff sequences, or mixtures of the two.

The conditions for melting of a tholeiitic parent could be: at high partial pressures of water, leaving an amphibolite residue; at low partial pressures of water, leaving a granulite residue; or at relatively high pressures, leaving an eclogite residue. If tholeiites were melting at relatively high water pressure ($P_{\text{H}_2\text{O}} > 0.6P_{\text{T}}$), they would form trondhjemites or tonalites, leaving an amphibole-rich residue (Barker & Arth 1976; Helz 1976) which would have r.e.e. pattern such as those in figure 6. Because these tonalites or trondhjemites would be nearly water-saturated, they would probably occur as deep-seated intrusions. These types of rocks do not presently seem to be common in greenstone belts.

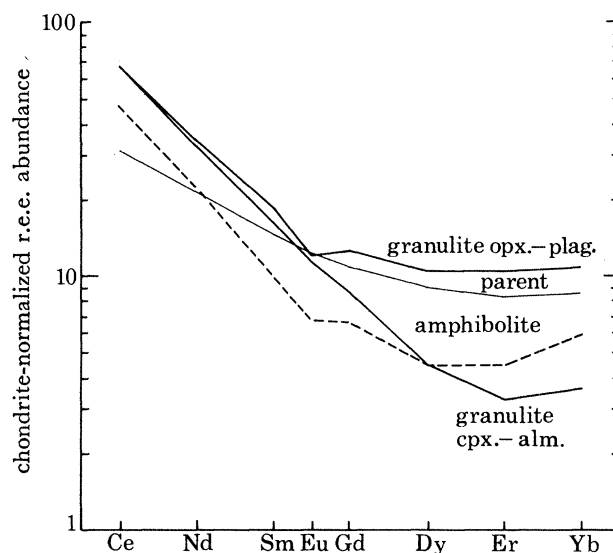


FIGURE 7. Calculated chondrite-normalized r.e.e. patterns for a melt derived by 20% melting of an average Archaean greywacke (Nance & Taylor 1976) leaving an amphibolite residue and for melts derived by melting an average Archaean greenstone-belt composition leaving an orthopyroxene-plagioclase subfacies and a clinopyroxene-garnet subfacies granulite-grade residue.

If the tholeiites were melting leaving a granulite residue, they would form andesites (Green & Ringwood 1968). There are two conditions for such melting, at the plagioclase-orthopyroxene grade, in which plagioclase and pyroxene would be the principle residue phases, or at the clinopyroxene-garnet grade, where plagioclase, clinopyroxene and garnet would be the principle residual phases (figure 6). Because these melts would be relatively dry they should form shallow dioritic intrusions or andesitic volcanics. Andesites in greenstone-belts have been reported with patterns similar to those calculated for the clinopyroxene-garnet grade (see figure 3*b*). These types of melts would be forming at depths of the order of 30 km or more (Winkler 1979, pp. 266-268). Due to their low water content these rocks would probably form shallow intrusions or extrusions. With increasing depth, as plagioclase becomes less stable, the minimum melt composition may be that of an anorthositic gabbro (Simmons & Hanson 1978). The r.e.e. patterns would be similar to that for the granulite residue with garnet in figure 6. Although these rocks may be common in the Proterozoic, particularly in the Grenville Province, they have not been described in the Archaean except as small dykes (Simmons *et al.* 1980). If melting occurs at a great enough depth, so that the residue is eclogitic (i.e. pyroxene plus garnet), dacites or tonalites (figures 3*b*, 6) may become products of melting (Green & Ringwood

1968). Leucocratic quartzo-feldspathic veins resulting from precipitation from an aqueous fluid leaving a granulite residue would have r.e.e. patterns similar to that for the saturated H_2O , at 4 kbar, 800 °C, in figure 2.

If greywackes are involved in melting, the melting may take place at amphibolite grade leaving a residue of plagioclase, amphibole, biotite, quartz and garnet or at granulite grade leaving a residue of plagioclase, pyroxene, biotite, quartz, garnet (Arth & Hanson 1975), in either case forming a granodiorite or quartz monzonite (figure 7). The melt formed leaving an amphibolite residue will probably be wetter and will probably not rise much above its zone of melting whereas the melt leaving a granulite grade residue will be drier and may rise significantly above the zone of melting and form a higher-level intrusion or even extrusion. Because large volumes of sediments would be involved in the melting process and greywackes tend to have similar compositions, the average compositions of the parent sediments should be quite similar and essentially independent of local conditions (Nance & Taylor 1976). Because garnet plus clinopyroxene at granulite grade or amphibolite \pm garnet at amphibolite grade have the strongest control of the r.e.e. abundances in the melt, the melts from the two different environments may be difficult to distinguish. The melts leaving an amphibolite residue will probably have lower Sr contents and larger negative Eu anomalies because at the lower temperature of melting the plagioclase–melt distribution coefficient are significantly greater than those for granulite-grade conditions (Drake & Weill 1975). The melt leaving an amphibolitic residue may be more concave upward than a melting leaving a granulite residue, due to the larger K_d for hornblende, particularly in the middle to heavy r.e.e. (compare patterns in figure 7).

In the consideration of the sequence of events in a greenstone belt it should be recognized that some apparent sequences of igneous rocks may be strongly controlled by the physical conditions during the development of the sequence. One possible model is a greenstone-belt assemblage that occurs in a rift zone maintained by an upwelling mantle. The mantle-derived basic and ultramafic volcanics and shallow intrusions are reworked by melting and sedimentary processes. The base of the sequence would consist of basic and possibly ultramafic volcanics and intrusions. As the sequence thickens or if there is a significant thickness of continental crust underlying the belt, the peridotitic and iron-rich magmas will extrude or intrude only if they have a sufficient hydrostatic head. Tholeiitic basalts and perhaps basaltic komatiites as well as dacitic and andesitic volcanics and intrusions derived by melting of tholeiites will then be emplaced or added to the sequence. Only after the sedimentary rocks are buried deep enough that they can melt will granitic melt form. Those melting at amphibolite grade will not rise very far above the zone of melting and will cut all of the previous rocks. Although in that section they would appear to be the youngest rocks, they are not the youngest rocks in the belt, because at the same time basalt and dacite may be added to the top of the sequence and tonalite magmas will still continue to intrude at shallow levels while melting takes place at depth. When upwelling of the mantle ceases, the production of komatiitic and basaltic melts ceases and maintenance of the rift also ceases. The type of rocks presently found in the belt will be a function of the depth of exposure. Basal sections will have mainly basic and ultramafic rocks, and associated sediments. Intermediate sections will have basic and dacitic volcanics, shallow tonalitic intrusions, volcanogenic sediments and greywackes, intrusive granites and gabbroic bodies representing magma chambers in which tholeiitic magma was differentiating (Simmons *et al.* 1980). Higher sections only having reached lower amphibolite grade or less would include basic, intermediate and dacitic volcanics, shallow tonalitic intrusions and associated sediments.

Alkalic rocks may be found at any level but may be more common at intermediate and higher levels. The relation of tectonic events to the volcanic, sedimentary and intrusive events is of course critical to evaluating any such model.

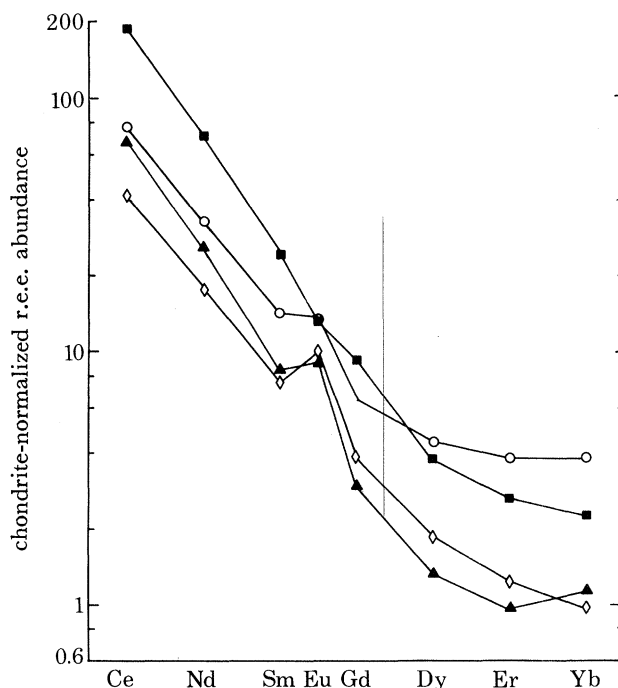


FIGURE 8. Chondrite-normalized r.e.e. patterns for Archaean gneisses from the Minnesota River Valley (unpublished data).

ARCHAEAN GNEISS TERRANES

The Archaean gneiss terranes are characterized by large volumes of intrusive leuco-granitic rocks of varying ages and extents of deformation (see, for example, McGregor 1973). Although there is a variety of granitic rock types, they are commonly trondhjemites or tonalites on a modal basis although on a normative basis they are tonalitic to granodioritic because much of the potassium occurs in biotite. Barker (1979) includes a good selection of articles regarding the geochemistry, petrography and petrology of Archaean gneiss terranes. Some typical r.e.e. patterns are shown in figure 8. These patterns have strong enrichment in the light r.e.e. with a relatively straight pattern from Gd through La with positive, negative or no Eu anomalies. The heavy r.e.e. abundances are usually about two to six times chondritic, with the shape and abundance of the heavy r.e.e. quite variable. Within a gneiss terrane some parents for the granitic bodies have been derived directly from the mantle or had a very limited history in the crust while other parents have clearly had a long crustal history (see, for example, Moorbath 1977).

While some of the granitic rocks may be derivatives of a mantle-derived basic magma, the relatively strong light r.e.e. enrichment would suggest that if they are the parent melt was alkaline. Those derived by differentiation from an alkaline basalt would tend to be relatively high in Ba, Sr, and ferromagnesian minerals and nearly silica-saturated. Those derived by

liquid immiscibility would be silica over-saturated with a lower total r.e.e. content. The gneisses cannot represent melts derived from melting a tholeiitic source with a flat r.e.e. pattern whether at amphibole, granulite or eclogite grade, because melting under these conditions will not lead to melts with enough enrichment in the light r.e.e. (see figure 6). Thus, if a basic source were involved in melting it must be light-r.e.e.-enriched, i.e. most like an alkali basalt. If an alkali basalt were the parent, melting would need to be at amphibolite or eclogite grade to allow the melt to be oversaturated in silica because amphibole and garnet may be undersaturated with respect to an alkali basalt composition while plagioclase and clinopyroxene are not.

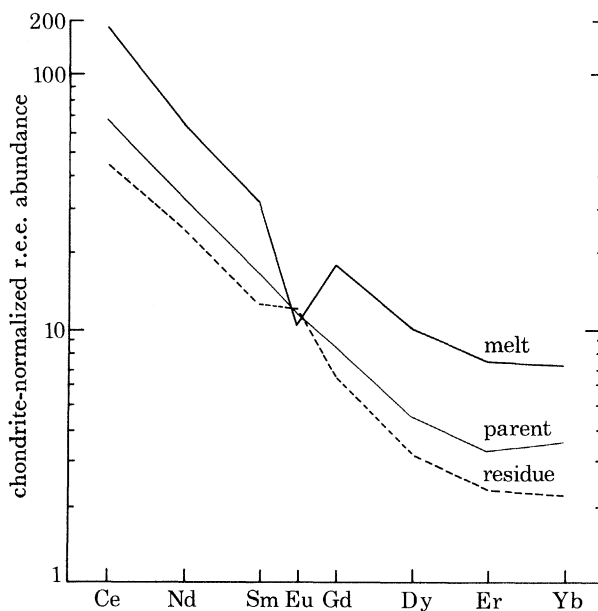


FIGURE 9. Calculated chondrite-normalized r.e.e. patterns for the melt and residue of 20% melting of a leucocratic gneiss leaving 30% (by mass) quartz, 60% (by mass) feldspar, 10% (by mass) biotite, 0.5% (by mass) apatite and 0.05% (by mass) zircon in the residue. Distribution coefficient data are given in Hanson (1980*b*).

Another potential source of granitic plutons is melting of pre-existing crustal rocks. In the Minnesota River Valley, for example, the gneisses that are about 3500–3700 Ma old, as well as the later intrusive granitic rocks at about 3100, 2600 and 1800 Ma B.P. (Goldich *et al.* 1981) all have the same approximate r.e.e. patterns (Hanson, unpublished data). The gneisses vary modally from quartz diorite or trondhjemite through quartz monzonite and the later intrusions are granodiorite to quartz monzonite. Pb isotope data by Doe & Delevaux (1981) suggest that the later intrusions are derived from related sources that were in a high $^{238}\text{U}/^{204}\text{Pb}$ environment for a significant period of time and then were in a very low $^{238}\text{U}/^{204}\text{Pb}$ environment also for a significant period of time before melting. The history of the source of these later intrusions is apparently comparable to that of the oldest gneisses in the area. Melting of the gneisses themselves to form the later intrusions would result in r.e.e. abundances for the granitic rocks that are greater than those of the gneiss and also the patterns would have significant negative Eu anomalies (figure 9).[†] This would rule out the gneisses as the parent of the later intrusive

[†] Note that in figure 9 the r.e.e. pattern for a residue of melting of a granitic gneiss parent is very similar to the r.e.e. pattern of the parent. Thus, it would be difficult to determine whether a quartzo-feldspathic gneiss was a residue of melting by means of only the r.e.e.

granites. Another possible source is melting of the lower crust which might have a history similar to that of the exposed gneisses. The lower crust probably consists of granulite grade rocks of intermediate composition (see, for example, Smithson & Brown 1977) that have been intimately mixed by folding, transposition of folds, and faulting. The volume of parent involved in melting to produce even a small intrusion must be many hundreds of cubic kilometres. Sampling of such a large volume of the lower crust would probably give a gross average of the lower crust allowing granitic intrusions derived from the lower crust to have rather similar characteristics.

Let us assume that the lower crust has the composition of an average greenstone belt and melt it leaving a granulite-grade residue. There are two conditions in the granulite grade where the melting may take place at the relatively low pressure or high temperature conditions of the plagioclase–orthopyroxene subfacies or at the relatively high-pressure, low-temperature conditions of the garnet–clinopyroxene subfacies. The main difference is that garnet is present in the residue for the garnet–clinopyroxene subfacies, but not in the plagioclase–orthopyroxene subfacies. Because garnet has a large effect on the heavy r.e.e., the r.e.e. patterns should be quite different for the two conditions of melting (see figure 7). If the gneiss and later intrusion are derived by melting in the granulite facies of an intermediate composition similar to that of average greenstone belts, it is suggested that the heavy r.e.e. depletion suggests that they are melting at the conditions of garnet–clinopyroxene subfacies. This would imply depths of melting on the order of 30 km (Winkler 1979). If melting occurred leaving an amphibolite residue or if the lower crust were peraluminous similar r.e.e. patterns could result by melting at shallower depths. If the melting is in the garnet–clinopyroxene subfacies, this would imply that during the Archaean the continents were as thick as they are now, which is compatible with geothermometry and geobarometry studies in other Archaean gneiss terranes (see, for example, Taylor 1979).

Heier (1973) among others has shown that granulite-grade rocks have significantly lower U, Th, and K concentrations than similar rocks of amphibolite grade. Also some gneisses show histories based on Pb–Pb isotope studies suggesting that originally they had relatively high $^{238}\text{U}/^{204}\text{Pb}$ ratios which were subsequently greatly reduced. These data suggest the possibility that the concentration of some elements in a gneiss terrane may have been affected by metasomatic processes. Touret (1974) discovered that, whereas fluid inclusions in amphibolite grade rocks are predominantly H_2O , in granulite-grade rocks CO_2 is common. Collerson & Fryer (1978) have suggested that selective solution of trace elements in migrating mantle-derived CO_2 vapour may be affecting the trace-element concentrations of lower crustal rocks, including the depletion of the heavy r.e.e. in the granitic gneisses. The distribution coefficient data of Wendlandt & Harrison (1979) for r.e.e. between CO_2 vapour and rock show that the r.e.e. are strongly partitioned into CO_2 vapour at 5 kbar, 1200 °C (see figure 2). Flynn & Burnham (1978), however, found that equal amounts of water and CO_2 at 800 °C and 4 kbar had the same fluid–silicate melt distribution coefficient as pure water. Their data would suggest that CO_2 with water would not be capable of removing significant amounts of r.e.e. In either case the limited data would suggest that the heavy r.e.e. would not be selectively enriched in a CO_2 -rich fluid.

If CO_2 vapour has higher vapour–rock distribution coefficients for the r.e.e. at higher P and T , mantle-derived CO_2 vapour may actually enrich the r.e.e. in the crust. The occurrence in high-grade gneiss terranes of high-grade gneisses and later essentially unmetamorphosed

granitic rocks with similar r.e.e. patterns as well as highly metamorphosed basaltic and mafic units with essentially flat r.e.e. patterns consistent with their major element chemistry would suggest that, while depletion of the heavy r.e.e. by mantle-derived CO₂ vapour may be possible, it does not appear to be a dominant process.

CRUST-MANTLE INTERACTION

The ultimate source of the Archaean continental crust was most likely the mantle. Because the main mantle-derived rocks preserved from the Archaean are those in greenstone belts it would seem reasonable to consider them to be potential primary sources that have been re-worked to produce the upper crust. To understand the evolution of the crust it is necessary to understand how the mantle has evolved through time.

The present mantle may be considered to consist of the suboceanic and subcontinental mantle. In the suboceanic mantle two end-member sources of magma are suggested by incompatible trace-element and isotope ratios, those for ocean islands and those for normal ocean-ridge basalts, where normal ocean-ridge basalts are those basalts that are relatively more depleted in the most incompatible elements (see, for example, Sun *et al.* 1979). Based on Rb-Sr and Pb-Pb isochron ages, Sun & Hanson (1975) suggested that those two sources have been separated for periods of thousands of millions of years and that they are not end members of mixing between a normal ocean-ridge source and an ocean-island source. This is best shown on a plot of ⁸⁷Sr/⁸⁶Sr against ²⁰⁶Pb/²⁰⁴Pb, in which the ocean-ridge basalts plot away from the main trend of data for the ocean islands and not at either end of a potential mixing curve. Although ocean-ridge basalts occur in spreading centres, whether at ocean ridges or in marginal basins, these environments are ubiquitous and encircle the globe. The ocean-island basalts are found in continental, island-arc and oceanic terranes as well as along ocean ridges, seemingly unrestricted in their geographic occurrence. Thus both sources appear to be ubiquitous but separated. Until we have better information regarding convection in the mantle, the simplest model to explain these observations is a stratified mantle in which the source for the ocean-ridge basalts is above the source for the ocean-island basalts. The relatively small variations in the major and trace element abundances and radiogenic isotope ratios for the ocean-ridge basalts compared to their relatively large variation in ocean-island basalts suggest that the ocean-ridge source is relatively well mixed or stirred while the ocean-island source is not.

Another potential mantle source for basalts is the subcontinental mantle which may have been attached to continental segments since times comparable to their times of formation (Jordan, this symposium). Because of their long residence time under the continents, the subcontinental mantle segments may have been significantly enriched in incompatible elements by fluid migration. The subcontinental mantle may be the source for some continental flood basalts, potassic lavas, and alkaline magmas with characteristics different from ocean-island assemblages, and the earliest basalts formed when spreading along a ridge or marginal basin broke up a continent.

Hager & O'Connell (1979) suggest that plate tectonics are mainly controlled by the thickening of the lithosphere, which is denser than the underlying mantle, and the viscosity of the underlying mantle. As the lithosphere cools it becomes thicker and more unstable which leads ultimately to subduction of the lithosphere and spreading at ocean ridges. If the mantle has a lower viscosity, the plate need not become as thick before subduction begins (Hager &

O'Connell 1981). In this model motion in the upper mantle is mainly controlled by the plate motion rather than being the cause of the plate motion, and neither plate motion nor the related motion in the mantle need reflect deep mantle convection. At a spreading centre, where the surrounding mantle is flowing into the 'void' created by the spreading, melting is occurring in shallow upwelling diapirs, tholeiitic magmas predominate and only a small range in magma types is found.

At the subduction zone the upper lithosphere including the ocean floor is metamorphosed, degassed and possibly melted. As discussed earlier small volumes of migrating fluids can have a large effect on the abundance of trace elements in the mantle (figure 1), whereas melting is relatively inefficient at separating incompatible elements (Hanson 1977). It is thus suggested that the relative depletion of the most incompatible elements in the mantle source for normal ocean-ridge basalts is in part a result of continued degassing of the subducted lithosphere which is stirred back into the mantle source for ocean-ridge basalts. Below the arc there may be melting of: the subducted plate to produce tonalites; the subcontinental mantle or ocean-ridge mantle to produce basalts; mafic rocks near the base of the crust to produce anorthositic or gabbroic plutons and possibly andesite; and continent crust of intermediate composition to produce granitic intrusions.

If the ocean-island source is below the ocean-ridge source, it is necessary to have that mantle raised to levels such that it may intersect its solidus. The most straightforward way is by uplift in adiabatic diapirs or mantle plumes which may be important components for whole-mantle convection. The depth at which the diapir intersects its solidus is a function of the total heat, major-element composition and volatile abundance of the diapir. If the deeper ocean-island source is not convecting with the shallower ocean-ridge source, the heat content of the deeper mantle should be greater, and a thermal boundary layer should separate the two sources. Any rising adiabatic diapir from such a deeper mantle would have a significantly greater temperature than the surrounding ocean-ridge source (B. H. Hager, personal communication, 1980). In such a rising diapir, a range in melt composition is to be expected even if the mantle is homogeneous (Oxburgh & Turcotte 1968). If the mantle is heterogeneous on every scale, as suggested by peridotitic nodules in basalts, peridotites in ophiolites, and basalt composition, then within any one diapir variations in melting may also occur as a function of local conditions. Those parts most enriched in volatile and basaltic components begin melting at greater depths; those least enriched melt at shallower depths or do not melt. Thus an ocean-island assemblage shows a range in both extents and depths of melting. This mantle source will then of course become a part of the ocean-ridge source. If an equivalent amount of ocean-ridge source is not added to the ocean-island source, the ocean-ridge source will grow at the expense of the ocean-island source.

Consideration of possible mantle sources in the Archaean in comparison to modern-day régimes must take into account any possible variations caused by different conditions during the Archaean. Hager & O'Connell (1981) conclude that the effective viscosity of the mantle is dependent on temperature, for example an increase of 100 °C will lead to a decrease of viscosity by a factor of 10, leading to a threefold increase in plate velocities. A higher heat content of the earth during the Archaean would thus lead to a lower viscosity of the mantle, higher plate velocities and subduction of younger and therefore thinner lithosphere. This would suggest that the individual plates need not have been smaller and may actually have been larger during the Archaean (Hager, personal communication, 1980). If the freeboard concept is valid (Wise

1974), the presence in the Archaean of continental crust with thicknesses comparable to those of the present crust would suggest that plate tectonics was an important factor (see Kroener (1981) for a review of models on Precambrian plate tectonics). If so, remnants of spreading centres and island arcs might be expected to be found in the Archaean.

At least in some greenstone belts there is apparently a wide range of rock types, varying from komatiites to tholeiites to derivatives of alkali basalts that represent large variations in both extents of melting and depths of melting. Possible modern analogues to Archaean greenstone-belt volcanic assemblages could be those at spreading centres, whether at ocean ridges or back-arc basins, island arcs, ocean islands and continental rifts. Tholeiites are the main magma types in normal spreading centres and a large range in melt types is generally not found. In island arcs a wide range of rock types is found, but andesite predominates, whereas in greenstone belts andesites are not a dominant rock type. The régimes that have the most similar assemblages are those in ocean islands or continental rifts. Ocean-island suites consist of tholeiites, to alkali basalts, to nephelinites representing ranges in both extents and depths of melting. The main difference may be that during the Archaean melts representing larger extents of melting were possible. If present ocean-island assemblages are associated with mantle plumes and mantle plumes are major agents for transferring heat from the deeper mantle, during the Archaean they may have been more common and may have played an important role in continental evolution.

At present there is a tendency to consider greenstone belts as having similar tectonic settings. This is, however, no more reasonable than to consider that all recent basic volcanic sequences have similar tectonic settings. Geological, petrological and geochemical criteria will have to be established to detect the subtle differences.

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REFERENCES (Hanson)

- Arndt, N. T. 1977 *Contr. Miner. Petr.* **64**, 205–222.
 Arth, J. G., Arndt, N. T. & Naldrett, A. J. 1977 *Geology* **5**, 590–594.
 Arth, J. G. & Hanson, G. N. 1975 *Geochim. cosmochim. Acta* **39**, 325–362.
 Barker, F. (ed.) 1979 *Trondhjemites, dacites and related rocks*. Amsterdam: Elsevier.
 Barker, F. & Arth, J. G. 1976 *Geology* **4**, 596–600.
 Bender, J. F., Bence, A. E., Hanson, G. N. & Ratcliff, N. M. 1979 *Eos, Wash.* **60**, 410.
 Bender, J. F., Hanson, G. N., Bence, A. E. & Ratcliffe, N. M. 1980 *Eos, Wash.* **61**, 411.
 Bottinga, Y. & Weill, D. F. 1970 *Am. J. Sci.* **269**, 169–182.
 Collerson, K. D. & Fryer, B. J. 1978 *Contr. Miner. Petr.* **67**, 151–167.
 Doe, B. R. & Delevaux, M. H. 1980 *Mem. geol. Soc. Am.* **182**. (In the press.)
 Drake, M. J. & Weill, D. F. 1975 *Geochim. cosmochim. Acta* **39**, 689–712.
 Flynn, R. T. & Burnham, C. W. 1978 *Geochim. cosmochim. Acta* **42**, 682–701.
 Fyfe, W. S. 1978 *Chem. Geol.* **23**, 89–114.
 Goldich, S. S., Hedge, C. E., Stern, T. W., Wooden, J. L., Bodkin, J. R. & North, R. M. 1980 *Mem. geol. Soc. Am.* **182**. (In the press.)
 Green, T. H. & Ringwood, A. E. 1968 *Contr. Miner. Petr.* **18**, 105–112.
 Hager, B. H. & O'Connell, R. J. 1979 *J. geophys. Res.* **84**, 1031–1048.
 Hager, B. H. & O'Connell, R. J. 1981 *Physics of the Earth's interior*, Enrico Fermi International School of Physics, course 78 (ed. A. M. Dziewonski & E. Boschi). Amsterdam: North Holland–Elsevier. (In the press.)
 Hanson, G. N. 1977 *J. geol. Soc. Lond.* **134**, 235–253.

- Hanson, G. N. 1980a *A. Rev. Earth planet. Sci.* **8**, 371–406.
- Hanson, G. N. 1980b In *Continental tectonics*, pp. 151–157. Washington: National Academy of Science.
- Hanson, G. N., Goldich, S. S., Arth, J. G. & Yardley, D. H. 1971 *Can. J. Earth Sci.* **8**, 1110–1124.
- Hanson, G. N. & Langmuir, C. H. 1978 *Geochim. cosmochim. Acta* **42**, 725–741.
- Hawksworth, C. J. & O'Nions, R. K. 1977 *J. Petr.* **18**, 487–520.
- Heier, K. S. 1973 *Fortschr. Miner.* **50**, 174–187.
- Helz, R. T. 1976 *J. Petr.* **17**, 139–193.
- Jahn, B. M. & Sun, S. S. 1979 *Phys. Chem. Earth* **11**, 597–618.
- Kay, R. W. & Gast, P. W. 1973 *J. Geol.* **81**, 653–682.
- Kroener, A. (ed.) 1981 *Precambrian plate tectonics*. Amsterdam: Elsevier. (In the press.)
- Kushiro, I., Yoder, Jr H. S. & Mysen, B. O. 1976 *J. geophys. Res.* **81**, 6351–6356.
- Langmuir, C. H., Bender, J. B., Bence, A. E., Hanson, G. N. & Taylor, S. R. 1977 *Earth planet. Sci. Lett.* **36**, 133–156.
- Langmuir, C. H. & Hanson, G. N. 1980 *Phil. Trans. R. Soc. Lond. B* **297**, 383–407.
- Lloyd, F. E. & Bailey, D. K. 1975 *Phys. Chem. Earth* **9**, 389–416.
- McGregor, V. R. 1973 *Phil. Trans. R. Soc. Lond. A* **273**, 343–358.
- Mercier, J.-C. 1977 Ph.D. thesis, State University of New York at Stony Brook.
- Moorbath, S. 1977 *Chem. Geol.* **20**, 151–187.
- Mysen, B. O. 1978 *Carnegie Instn Wash. Yb.* **77**, 689–695.
- Nance, W. B. & Taylor, S. R. 1976 *Geochim. cosmochim. Acta* **40**, 1539–1551.
- Naslund, H. R. 1976 *Carnegie Instn Wash. Yb.* **75**, 592–597.
- O'Nions, R. K., Carter, S. R., Evenson, N. M. & Hamilton, P. J. 1979 *A. Rev. Earth planet. Sci.* **7**, 11–38.
- Oxburgh, E. R. & Turcotte, D. L. 1968 *J. geophys. Res.* **73**, 2643–2661.
- Philpotts, A. R. 1976 *Am. J. Sci.* **276**, 1147–1177.
- Roedder, R. 1978 *Geochim. cosmochim. Acta* **42**, 1597–1617.
- Ryerson, F. J. & Hess, P. C. 1978 *Geochim. cosmochim. Acta* **42**, 921–932.
- Schultz, K. J., Smith, I. B. M. & Blanchard, D. P. 1979 *Eos, Wash.* **60**, 410.
- Schwarzer, R. R. & Rogers, J. J. W. 1974 *Earth planet. Sci. Lett.* **23**, 286–296.
- Simmons, E. C. & Hanson, G. N. 1978 *Contr. Miner. Petr.* **66**, 119–135.
- Simmons, E. C., Hanson, G. N. & Lumbers, S. B. 1980 *Precamb. Res.* **11**, 43–71.
- Sims, P. K. & Mudrey Jr, J. G. 1972 In *Geology of Minnesota: centennial volume* (ed. P. K. Sims & G. B. Morey), pp. 140–152. St Paul: Minnesota Geological Survey.
- Sims, P. K. & Viswanathan, S. 1972 In *Geology in Minnesota: centennial volume* (ed. P. K. Sims & G. B. Morey), pp. 120–139. St Paul: Minnesota Geological Survey.
- Smithson, S. B. & Brown, S. K. 1977 *Earth planet. Sci. Lett.* **35**, 134–144.
- Stolper, E., Walker, D. & Hays, J. F. 1981 *J. geophys. Res.* (In the press.)
- Sun, S. S. & Hanson, G. N. 1975 *Contr. Miner. Petr.* **52**, 77–106.
- Sun, S. S. & Nesbitt, R. W. 1977 *Earth planet. Sci. Lett.* **35**, 429–448.
- Sun, S. S. & Nesbitt, R. W. 1978 *Contr. Miner. Petr.* **65**, 301.
- Sun, S. S., Nesbitt, R. W. & Sharaskin, A. Y. 1979 *Earth planet. Sci. Lett.* **44**, 119–138.
- Taylor, S. R. 1979 In *The Earth: its origin, structure and evolution* (ed. M. W. McElkinney), pp. 353–376. New York: Academic Press.
- Touret, J. 1974 *Annls Soc. géol. Belg., Vol. spec.* (ed. P. Michot), pp. 267–287.
- Visser, W. & Koster Van Groos, A. F. 1979 *Am. J. Sci.* **279**, 1160–1175.
- Watson, E. B. 1976 *Contr. Miner. Petr.* **56**, 119–134.
- Watson, E. B. & Naslund, H. R. 1977 *Carnegie Instn Wash. Yb.* **76**, 410–414.
- Winkler, H. G. F. 1979 *Petrogenesis of metamorphic rocks*. New York: Springer Verlag.
- Wendlandt, R. F. & Harrison, W. J. 1979 *Contr. Miner. Petr.* **69**, 409–419.
- Wise, D. U. 1974 In *The geology of continental margins* (ed. C. Burke & C. L. Drake). New York: Springer Verlag.
- Zielinski, R. & Frey, F. 1970 *Contr. Miner. Petr.* **29**, 242–254.

Discussion

A. KRÖNER (University of Mainz, F.R.G.). The mantle-plume model as seen by Dr Hanson to trigger magmatic activity and Archaean greenstone-belt – granite evolution is more compatible with the apparently long time (up to 200 Ma) for the development of individual greenstone belts than the back-arc basin model since fast plate motion in the Archaean as indicated by palaeomagnetism (Dunlop, this symposium) would imply short Wilson-cycle activity, which is not observed.

However, it is difficult to visualize continuous plume activity under an evolving greenstone belt for over 100 Ma while the lithospheric plate carrying the belt moves rapidly over the asthenosphere. Could it be that the plume, once it has penetrated high enough into the lithosphere, can become decoupled from its own source and remain active for geologically significant periods?

G. N. HANSON. Plumes may have been more common during the Archaean and may have occurred as planar features or in planar arrays giving the appearance at the surface of linear features or 'hot lines'. Thus, depending on the rate and direction of travel of a plate relative to an existing array of hot spots or hot lines a greenstone belt may have a short or long history. A greenstone belt in the middle of a continent may also have been reactivated by later tectonic events, especially these associated with the break up of a continent preceding the development of an ocean ridge.