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Archaean mantle heterogeneity: evidence from chemical and isotopic abundances in Archaean igneous rocks

By B.-M. Jahn,[†] P. Vidal[†] and G. R. Tilton[‡]
[†] *Université de Rennes, Institut de Géologie, Avenue du Général-Leclerc,*
35042 *Rennes Cedex, France*
[‡] *Department of Geological Sciences, University of California, Santa Barbara,*
California 93106, *U.S.A.*

The problem of compositional heterogeneity in the Archaean upper mantle is evaluated from the existing data of isotopic compositions (Sr, Pb and Nd), trace element abundances (rare earth elements (r.e.e.)) and some critical major element chemistry in Archaean igneous rocks. To assess a large-scale mantle heterogeneity, we have emphasized the chemical characteristics of undoubtedly mantle-derived komatiites.

The initial isotopic compositions of Sr and Pb (I_{Sr} and I_{Pb}) in Archaean igneous rocks show a significant variation, suggesting that a long-term heterogeneity in Rb/Sr and U/Pb ratios might have existed in their mantle sources. In contrast, the available I_{Nd} values show little variation from the expected evolution in a chondritic uniform reservoir. These contrasting results, i.e. the variable I_{Sr} and I_{Pb} values as against the uniform I_{Nd} values, are in part attributed to the fact that mobile–refractory pairs in the Rb/Sr and the U/Pb systems but only the refractory–refractory pair in the Sm/Nd system have been involved in geochemical differentiation.

The r.e.e. distributions in komatiitic rocks are not uniform. Many show strong depletion in light r.e.e. The degree of light r.e.e. depletion in these komatiitic rocks is comparable with that found in present mid-ocean ridge basalts. Regarding the major element chemistry, we have demonstrated that the CaO/Al_2O_3 ratios are significantly different in the komatiitic suites of two adjacent cratons (Kaapvaal and Rhodesia) in southern Africa. Both trace and major elements data suggest that the Archaean upper mantle was heterogeneous, but whether the heterogeneity truly represents a long-term phenomenon remains to be resolved by more isotopic analyses.

1. INTRODUCTION

Partial melting of the upper mantle and the subsequent extraction and transport of magmatic liquids to the crust naturally lead to a compositional heterogeneity in this segment of mantle. So long as there are active melting processes and magmatism, there must exist a certain degree of mantle heterogeneity. The heterogeneity of the mantle at any given time is related to the extent of previous melting events and to the efficiency of later reverse processes that efface the degree of heterogeneity. These reverse processes include elemental diffusion and the more efficient mass convection and mixing. It is necessary, therefore, to distinguish the short-term and the long-term heterogeneity of the mantle. We are here more concerned with the long-term phenomena and effects that have been registered in geologic materials.

The evidence for the chemical and isotopic heterogeneity of the modern mantle is very well documented. Commonly the long-term effect of chemical heterogeneity is best indicated by certain radiogenic isotopic abundances, such as those of Sr, Pb and Nd (see related papers in this symposium). Stable isotopes such as oxygen are not produced by radioactive decay processes. However, a recent interesting study of oxygen isotopes in oceanic basalts by Kyser & O'Neil

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(1978) has shown a meaningful difference between the $\delta^{18}O$ values of tholeiitic rocks (+5.0 to +6.2) and alkali basalts (+5.8 to 7.5). These authors argued that the alteration (hydration) process and chemical differentiation cannot account satisfactorily for the difference in $\delta^{18}O$ values; rather, if isotopic equilibrium is maintained during melting, alkali basalts and tholeiites are derived from heterogeneous sources.

The degree of mantle heterogeneity in Archaean times is the subject of this article. The Archaean mantle is generally believed to be hotter and the geothermal gradient was probably substantially higher compared with that of the present day. Two major sources of heat might have contributed to the higher temperatures: (1) greater abundances of long-lived heat-producing radioactive elements (^{238}U, ^{235}U, ^{232}Th and ^{40}K) in the Archaean mantle, and (2) residual heat that was earlier transformed from the gravitational energy as a result of core formation. A causal effect of greater heat reservoir in the Archaean mantle is the more rapid mass convection within it. We could envisage that because of the significant size of the Earth's core, the heat generated must be efficiently dissipated through convective motion. This convective motion can only be enhanced when counting together the heat contributed from the radioactivity. Furthermore, in the early Archaean ($t \geqslant 3.8$ Ga), some major meteoritic impacts onto the Earth's surface are predicted as inevitable events. The impact events might have induced pressure-released partial melting of the upper mantle and augmented the geothermal gradient which in turn increased the rate of convection within the mantle.

The common occurrence of peridotitic komatiites in Archaean terrains may add supporting evidence for a rapidly convecting Archaean mantle. The experimental study of Green (1975) suggests that the extrusive temperature of periodotitic komatiites is higher than 1600°C and the diapirs producing these magmas would have deep origins (not less than 200 km). Whichever acceptable geotherm model of Green (1975) is chosen, the sublithospheric convective cell would be deeper in the Archaean than today. The rapid ascent of mantle diapirs, which is required to generate a high degree of partial melting for peridotitic komatiite magma before partial melt is segregated, also suggests a more active convective motion in the mantle. A rapid convective motion should be capable of erasing the compositional heterogeneity with greater efficiency. Consequently, the Archaean mantle composition might be presumed to be more homogeneous despite the fact that partial melting events might have occurred more frequently. We shall examine if this presumption is indeed valid from the available chemical and isotopic data of Archaean igneous rocks.

It should be emphasized that isotopic ratios are the first order indicators of long-term effect of heterogeneity. Trace element data are the second order of useful indicators, whereas major element abundances are of the third order. This can be argued from the fact that the modern oceanic basalts are relatively similar in major element chemistry, but trace element contents are distinctly different among the major rock types, such as mid-ocean ridge basalts (m.o.r.b.) oceanic island tholeiites and alkali basalts. We shall present in the following our evidence for heterogeneity in the Archaean mantle in the order of isotopes, trace elements and major elements.

2. Isotopic data

(a) *Strontium*

Figure 1 is a summary of the available initial $^{87}Sr/^{86}Sr$ ratios (termed I_{Sr} values hereafter) mainly obtained by the whole-rock isochron method for many Archaean igneous and meta-

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igneous rocks. In this Sr isotopic evolution diagram, we note that the most abundant data come from rocks of 2.5–2.7 Ga of age. Evidently, this suggests that this is a period of time when active continent forming events took place (see also the argument from Nd isotopic data of McCulloch & Wasserburg (1978)). The data also show a 'wide' range of I_{Sr} values for rocks formed in this period of time. Older rocks from Greenland and southern Africa show the same degree of scattering of I_{Sr} values, taking the time of evolution into account.

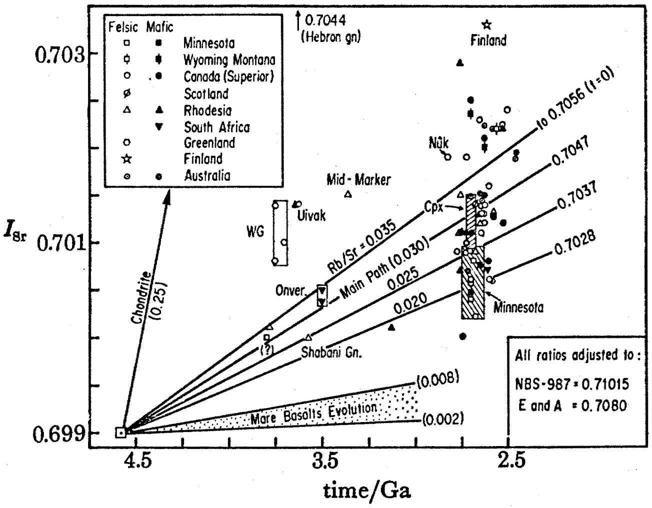

FIGURE 1. Sr isotopic evolution (I_{Sr}–T) diagram for Archaean igneous rocks. Data sources from the literature.

We also note that the data for granitic rocks (*sensu lato*) and for mafic rocks (mainly obtained by dating greenstones) do not differ much in the ranges of I_{Sr} values; this was earlier noticed by Hart & Brooks (1977). The data points are most densely populated along the main path of evolution line characterized by a Rb/Sr ratio of 0.029 ± 0.003, which leads to a present-day I_{Sr} values of about 0.7045 ± 0.0006. This value of about 0.7045 has been considered to represent best the Sr isotopic ratio of non-fractionated, or pristine, terrestrial upper mantle (Jahn & Nyquist 1976; DePaolo & Wasserburg 1976*a*, *b*; Hart & Brooks 1977; O'Nions *et al.* 1977). In figure 1, the data on primary clinopyroxenes (Hart & Brooks 1977) separated from metavolcanic rocks lie within the main path of evolution. The narrow yet significant range of I_{Sr} values was interpreted by Hart & Brooks (1977) to be a result of source heterogeneity. We must also note that the data ensemble for Minnesota and for some southern African and western Australian rocks show mean values lower than the clinopyroxene data, although their error bars overlap. Higher I_{Sr} values, evolved in the source(s) with Rb/Sr > 0.040, are not uncommonly observed, and the interpretation of these 'high' values is often disputed (see, for example, Moorbath 1978; Hurst 1978). In summary, the scatter of I_{Sr} values could have resulted from (1) metamorphic and alteration effects leading to open-system behaviour of analysed whole-rock samples, (2) anatexis of long-lived sources with high Rb/Sr ratios, particularly for rocks of granitic compositions (however, as stated earlier, the data for granitic and mafic rocks are not distinguishable in figure 1), or (3) true compositional heterogeneity in the mantle sources in terms of Rb/Sr abundance ratio.

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(*b*) *Lead*

Figure 2 illustrates the evolution of common Pb in many Archaean rocks. In this diagram we have assumed a two-stage evolution model. Since most data show that their secondary isochrons pass through the field of modern Pb isotopic composition as defined by Stacey & Kramers (1975), the assumption of a two-stage evolution may have some validity.

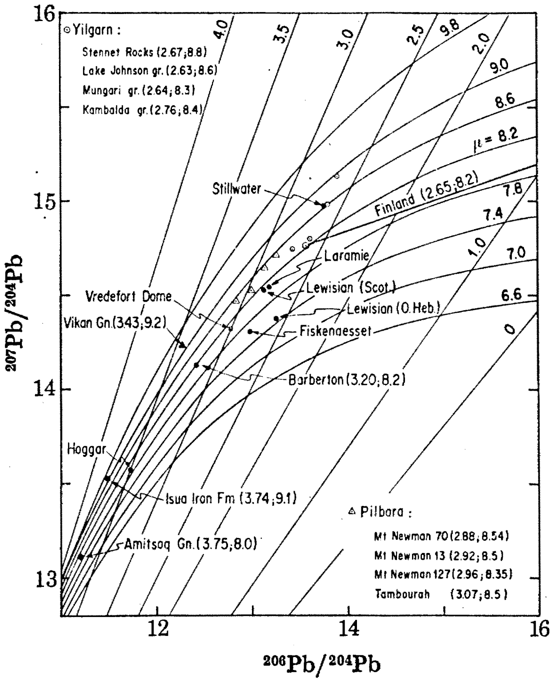

FIGURE 2. $^{207}Pb/^{204}Pb$ plotted against $^{206}Pb/^{204}Pb$ for Archaean rocks. Data points indicate initial Pb (I_{Pb}) isotopic compositions assuming a two-stage evolution model. Secondary isochrons are not shown for clarification except for the Finnish data. Data sources from the literature.

We should like to single out an example from Finnish Archaean rocks. Vidal *et al.* (1979) have analysed rocks of diverse composition, including peridotitic and basaltic komatiites from a greenstone belt, a post-kinematic granodiorite which intrudes the greenstone belt, and various granitic gneisses surrounding the belt (regarded as basement rocks). Their common Pb isotopic compositions are displayed in figure 3. It is clear that all data points except three fall near or on a single regression line which also passes through the modern Pb field. This data set leads us to propose the following interpretation. (1) This regression line does represent a secondary isochron, which yields an age of 2.65 ± 0.05 Ga, being interpreted as the time of formation of these rocks. This Pb–Pb age of 2.65 Ga is in close agreement with a Rb–Sr whole rock isochron age of 2.61 ± 0.06 Ga ($\lambda Rb^{87} = 0.0142\ Ga^{-1}$) as earlier reported in Blais *et al.* (1977*a*). If the pegmatite data are excluded, the Rb–Sr age becomes 2.66 ± 0.10 Ga. (2) The inferred μ_1 value for the predecessor(s) of these rocks is 8.2. This suggests that the rocks are probably derived from an upper mantle source, or from sources with similar U/Pb ratios, although they may not be necessarily comagmatic as evidenced from their trace element distributions (Jahn *et al.* 1980). (3) The fact that all rocks of different composition, both from within and without the greenstone belt, lie along the same isochron suggests that the total time span of development of both

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the basement and the greenstone belt is probably less than 100 Ma, an observation quite consistent with the data from other Archaean terrains (see, for example, Hart & Davis 1969; Jahn & Murthy 1975; Hawkesworth *et al.* 1975; Cooper *et al.* 1978). On the other hand, as in the Vikan Gneisses of Norway, the linear regression line of 3.43 Ga ($\lambda_8 = 0.155125\ \mathrm{Ga}^{-1}$; $\lambda_5 = 0.98485\ \mathrm{Ga}^{-1}$) by Taylor (1975) does not pass through anywhere near the modern Pb field, and its inferred μ_1 value of 9.2 is rather high. The line may thus be a consequence of U loss relative to Pb in a later granulite metamorphism. A recent Sm/Nd isotopic study (Jacobsen & Wasserburg 1978) demonstrates that the Vikan gneisses were formed about 2.6 Ga ago and were subject to a granulite metamorphism and U loss about 1.8 Ga ago. Recently, Taylor (1978) has reinterpreted his original data and reached a similar conclusion. There are other 'abnormal' secondary isochrons, such as the ones represented by Fiskaenesset ($t = 2.8$ Ga; $\mu_1 = 7.6$), Lewisian (2.6 Ga; $\mu_1 = 7.5$) and Stennet Rocks of Norseman, Australia (2.67 Ga; $\mu_1 = 8.8$) (Oversby 1975). The first two isochrons pass significantly below the modern Pb field, and the last passes significantly above it and it has an inferred μ_1 value of about 9.0. We suspect that the history of these rocks may be more complicated than a two-stage model could adequately describe.

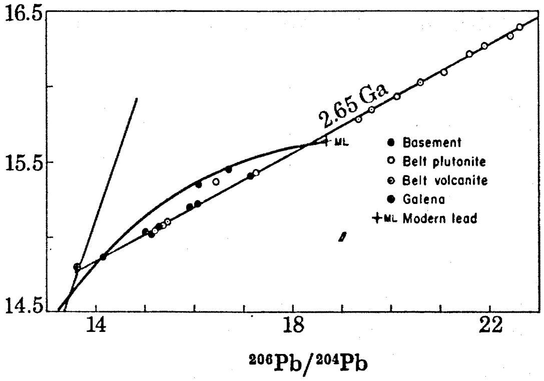

FIGURE 3. ^{207}Pb/^{204}Pb plotted against ^{206}Pb/^{204}Pb for the Finnish Archaean rocks (data from Vidal *et al.* 1979).

Nevertheless, when applying the two-stage model to most data, we obtain initial Pb isotopic compositions (I_{Pb} values) and inferred μ_1 values corresponding to individual isochrons. As shown in figure 2, these I_{Pb} and μ_1 values are not so homogeneous. The μ_1 values range from 8.6 (Stillwater and Isua Iron Formation) to 7.8 (Hoggar gneisses, $t = 3.47$ Ga; P. Vidal, unpublished data) and possibly to as low as 7.0 for the Amitsoq gneisses if the age is taken as 3.6 Ga as determined by Baadsgaard (1976) and Gancarz & Wasserburg (1977). We tend to think that this scatter of μ_1 values represents a compositional heterogeneity in terms of U/Pb ratio in the Archaean mantle sources. Oversby (1978) has reached the same conclusion mainly from the K-feldspar Pb data, but we think that her argument is weakened by the fact that the Pb in K-feldspar is often disturbed during later metamorphic episode(s). Some examples can be found in the Pb data on K-feldspar from Wyoming and on Finnish Archaean granitic gneisses (Rosholt *et al.* 1973; Vidal *et al.* 1979).

(*c*) *Neodymium*

The Nd isotopic composition (^{143}Nd/^{144}Nd ratios, termed I_{Nd} values hereafter) in modern volcanic rocks are demonstrably heterogeneous and they suggest that their respective sources

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are distinctly different in Sm/Nd ratios (and thus the r.e.e. patterns) for more than 10^9 years (DePaolo & Wasserburg 1976*a*, *b*, 1977, 1978; O'Nions *et al.* 1977; Carlson *et al.* 1978). Their ϵ_{Nd} values range from -14 (equivalent to $I_{Nd} = 0.5119$) for continental potassic rocks to about $+14$ ($I_{Nd} = 0.5134$) for m.o.r.b.

For Precambrian rocks, the I_{Nd} data are much more limited. Most data so far available (figure 4) suggest that the Archaean rocks have been derived from a rather uniform reservoir or source characterized by a chondritic Sm/Nd ratio (0.318 mass ratio, or $^{147}Sm/^{144}Nd = 0.1936$ atomic ratio) (DePaolo & Wasserburg 1976*a*, *b*; Hamilton *et al.* 1978; Zindler *et al.* 1978). This implies that (1) the Archaean upper mantle was indeed quite homogeneous in terms of Sm/Nd ratio and the r.e.e. abundance patterns in general, (2) the compositional heterogeneity is not yet revealed by the limited data, or (3) the true heterogeneity in the r.e.e. characteristics (see next section) as we have observed in many komatiitic rocks (see, for example, Hawkesworth & O'Nions 1977; Arth *et al.* 1977; Jahn *et al.* 1980; Sun & Nesbitt 1978) is only a short-term phenomenon. We favour the second and third implications. In fact, DePaolo (1978) most recently reported that the age of the Stillwater Complex, Montana, was determined by the Sm/Nd mineral isochron method at 2.701 ± 0.015 Ga. Its ϵ_{Nd} value from the chondritic uniform reservoir evolution line is -2.6. DePaolo emphasized that this ϵ_{Nd} value is distinctly below the chondritic reservoir evolution curve and suggested that the source probably had been enriched in light r.e.e. for some time, or that the Stillwater magmas had been contaminated with older crustal rocks.

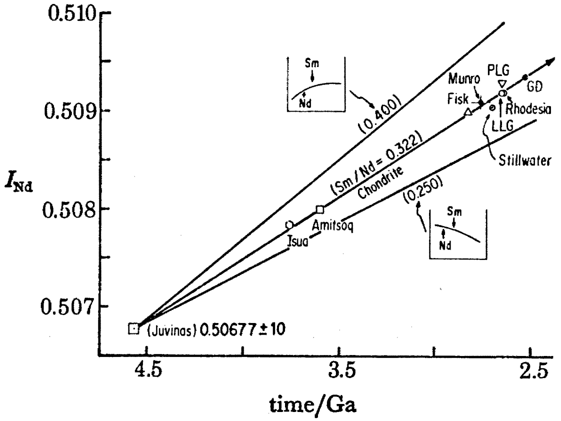

FIGURE 4. Nd isotopic evolution (I_{Nd}–T) diagram for some Archaean rocks. Sm/Nd ratio by mass. Data sources from the literature.

3. TRACE ELEMENT DATA

The problem of chemical heterogeneity of the Archaean mantle, particularly in terms of r.e.e. distribution, was recently discussed by Sun & Nesbitt (1977). New data on r.e.e. distributions in komatiites from the Munro Township and Noranda, Canada (Arth *et al.* 1977; Jahn, unpublished) and Finland (Jahn *et al.* 1980) further substantiate the existence of important heterogeneity in the Archaean mantle.

Komatiitic rocks occur in many Archaean greenstone terrains, e.g. the type locality in the Komati Formation of the Barberton Mountainland (Viljoen & Viljoen 1969), western Australia (Nesbitt 1972; Nesbitt & Sun 1976), the Superior Province, Canada (Pyke *et al.* 1973; Arndt

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et al. 1977), Rhodesia (Bickel *et al.* 1975; Nisbet *et al.* 1977) and Finland (Blais *et al.* 1978; Jahn *et al.* 1980). Peridotitic komatiites are generally believed to represent very large degree of partial melting of upper mantle material (Viljoen & Viljoen 1969; Green 1975; Arndt *et al.* 1977); they could in principle provide valuable information regarding the chemical composition of the upper mantle.

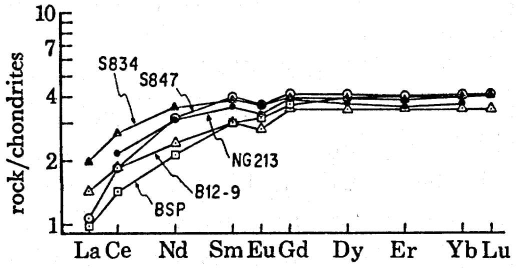

FIGURE 5. Some typical light r.e.e. depleted peridotitic komatiites from Finland (S834, S847), Canada (B12-9, BSP) and Rhodesia (NG 213). Data sources: Jahn *et al.* (1980); Hawkesworth & O'Nions (1977); Jahn (unpublished).

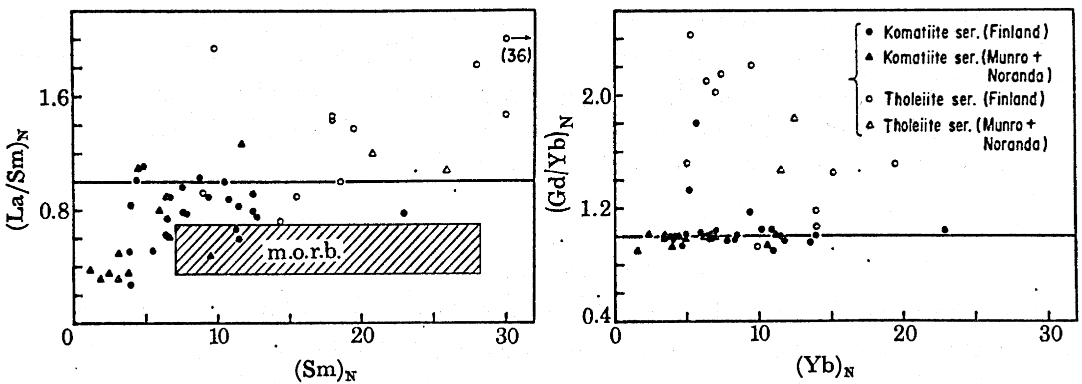

FIGURE 6. R.e.e. characteristics of some Archaean volcanic rocks (komatiites and tholeiites) from Finland and Canada. Data sources: Jahn *et al.* (1980); Arth *et al.* (1977); Jahn (unpublished).

In terms of r.e.e. geochemistry, many komatiitic rocks show unfractionated patterns relative to chondritic abundances (see, for example, Hawkesworth & O'Nions 1977; Sun & Nesbitt 1978; Jahn *et al.* 1980). Yet there are many others that show severely depleted light r.e.e. patterns (figure 5). Typical examples are found in the peridotitic komatiites of Abitibi (Arth *et al.* 1977; Whitford & Arndt 1978; Sun & Nesbitt 1978), Noranda (Jahn, unpublished), Rhodesia (Hawkesworth & O'Nions 1977), western Australia (Sun & Nesbitt 1978) and Finland (Jahn *et al.* 1980).

In figure 5 it is seen that this type of patterns is often observed in modern m.o.r.b., although the absolute r.e.e. abundances in the peridotitic komatiites are at least two times lower.

The Negri volcanics (spinifex textured basalts) from western Australia have light r.e.e. enriched (La = 25–40 × chondrite) but unfractionated heavy r.e.e. (Yb = 6–8 × chondrite) patterns (Sun & Nesbitt 1978). In combination with the result of high pressure experimental work, Sun & Nesbitt (1978) concluded that the source region for the Negri volcanics has been enriched in light r.e.e.

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In summary, most komatiitic rocks are derived from sources characterized by either unfractionated r.e.e. or light r.e.e. depleted patterns. Few have been clearly shown to have enrichment in light r.e.e., as have the Negri volcanics. Figure 6 illustrates the r.e.e. characteristics in Archaean volcanics from Finland and Canada. Many more data can be found in Herrmann *et al.* (1976), Arth *et al.* (1977), Whitford & Arndt (1978), Hawkesworth & O'Nions (1977) and Sun & Nesbitt (1978). However, those shown in figure 6 should suffice to demonstrate the generality for Archaean volcanic rocks.

In figure 6, we see that many komatiitic rocks display a strong light r.e.e. depletion ($(\mathrm{La/Sm})_N < 1.0$) but maintain unfractionated heavy r.e.e. ($(\mathrm{Gd/Yb})_N \approx 1.0$). The degree of light r.e.e. depletion is comparable with that of m.o.r.b. It should be emphasized here that since the residual phases after a large degree of partial melting of the mantle source are likely to be olivine and orthopyroxene, and because both phases have very small K_D (r.e.e.) values, the r.e.e. distribution patterns in komatiitic rocks (particularly the peridotitic variety) should reflect those of their sources. Therefore, it is logical to argue that the Archaean mantle was rather heterogeneous in terms of r.e.e. composition. It appears at present that the widespread occurrence of light r.e.e. depleted peridotitic komatiites could suggest that the Archaean mantle was not sufficiently homogenized by the presumably more vigorous convective motion. Though heterogeneity is convincingly demonstrated by the variable La/Sm ratios, the degree of heterogeneity becomes less pronounced in terms of Sm/Nd ratios and even less so with respect to the isotopic composition of Nd, particularly if the heterogeneity was created not long before the rock was formed. We suppose that long-lived heterogeneity might have existed in the Archaean and this will have to be answered by future Sm/Nd studies.

4. Major element data

Rock types are usually classified on the basis of their mineralogical and major element compositions. Hence rocks of the same type commonly have similar major element constituents. The change in major element composition is mainly attributed to crystal fractionation plus some minor contribution from contamination, which includes assimilation and mixing. How can we then look into the possibility of source heterogeneity in terms of major element chemistry? Again, the komatiitic rocks provide us with some good samples of material to witness the heterogeneity of the mantle source(s).

The original definition of komatiites emphasizes the high $\mathrm{CaO/Al_2O_3}$ ratio (about 1.5; Viljoen & Viljoen 1969), or equivalent to the $\mathrm{CaO/(CaO+Al_2O_3)}$ ratio (termed the C value hereafter) about 0.6. The high C value is, curiously enough, found so far mainly in the peridotitic komatiites and basaltic komatiites of the Onverwacht Group. Many subsequent geochemical studies on komatiites from western Australia (Sun & Nesbitt 1978), the Abitibi (Arndt *et al.* 1977), Rhodesia (Nisbet *et al.* 1977) and eastern Finland (Blais *et al.* 1977*b*; Jahn *et al.* 1980) reveal that the high $\mathrm{CaO/Al_2O_3}$ ratio is not universal. In fact, most komatiites show $\mathrm{CaO/Al_2O_3} < 1.0$ or C values of 0.4–0.5 (figure 7). Because of the nature of residual phases, i.e. olivine and othopyroxene that bear small amounts of Ca and Al, the $\mathrm{CaO/Al_2O_3}$ ratios observed in peridotitic komatiites must be, in principle, identical to that of their mantle source(s), unless some phases capable of fractionating CaO from $\mathrm{Al_2O_3}$, such as garnet, were somehow separated from komatiitic melts. Plagioclase, though having a high $\mathrm{Al_2O_3/CaO}$ ratio, is not considered as an important phase in komatiite genesis (see Arndt 1976).

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We now examine a small mantle segment from southern Africa. Rocks of the komatiitic series from the classic Onverwacht Group in the Barberton Mountainland and the adjacent Rhodesian Craton are shown in a plot of MgO against C values (lower abscissa) or CaO/Al_2O_3 ratio (upper abscissa). This plot is merely a transformed diagram of $CaO–MgO–Al_2O_3$ originally used by Viljoen & Viljoen (1969), but it has the advantage of emphasizing possible distinct evolutionary trends and hence mantle heterogeneity. It is rather clear that the Onverwacht data and the data of Rhodesia form two distinct and nearly parallel evolutionary trends, with the Onverwacht data having considerably higher C values (0.55–0.7), except the basaltic komatiites of Barberton type. We may add that our Finland data, as shown in figure 7*a*, and those of the Munro (Arndt *et al.* 1977) and western Australia (Naldrett & Turner 1977; Sun & Nesbitt 1978), follow very closely the trend of the Rhodesian samples. What, then, are the possible interpretations for these two markedly different trends?

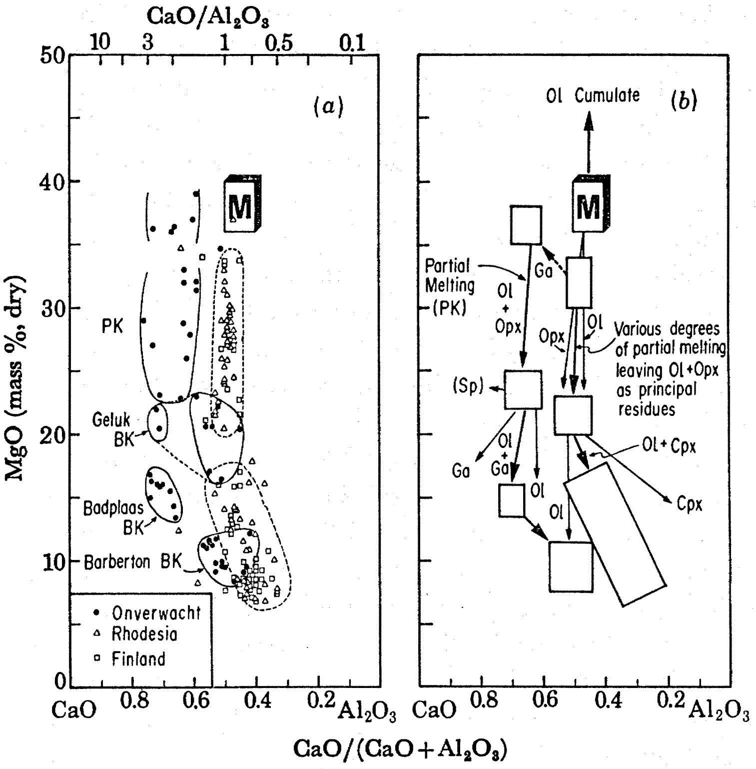

FIGURE 7. (*a*) MgO plotted against $CaO/(CaO+Al_2O_3)$ ($= C$ value) for the komatiite data from the Onverwacht Group of South Africa, Rhodesia and Finland. Data sources: Viljoen & Viljoen (1969); Bickle *et al.* (1975); Nisbet *et al.* (1977); Sun & Nesbitt (1978); Jahn *et al.* (1979). (*b*) A model interpretation for the data sets shown in (*a*).

Our interpretations are as follows and refer to figure 7*b*.

1. The two trends represent two *independent* mantle sources characterized by different CaO/Al_2O_3 ratios. This carries an implication for a long-term heterogeneity in the Archaean upper mantle. Partial melting of the high CaO/Al_2O_3 source would produce the Onverwacht peridotitic and basaltic komatiite rocks of Badplaas and low-Al Geluk types. Later clinopyroxene fractionation from the Badplaas basaltic komatiites may eventually form the basaltic komatiites of Barberton type. On the other hand, melting of the lower CaO/Al_2O_3 source,

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which is considered to be the 'normal' source owing to its wider occurrence and closer proximity to the pyrolite or other reasonable model mantle composition represented by the box M of figure 7*b*, would produce the komatiites of Rhodesia, Finland and the Munro areas. An intricate issue of whether most basaltic komatiites have evolved from peridotitic komatiites by fractional crystallization or whether they represent direct primary liquids from partial melting of mantle sources is still under debate (see Arndt *et al.* 1977; Sun & Nesbitt 1978; Nisbet *et al.* 1977). However, for Finland we have proposed not only an interplay of direct partial melting and fractional crystallization, but also that heterogeneous sources are required to account for the complexity of r.e.e. distributions in the komatiitic rocks (Jahn *et al.* 1980).

2. The two trends are *related* by garnet fractionation during diapiric uprise (Green 1975). As shown in the model diagram of figure 7*b* significant garnet separation from a peridotitic komatiite melt derived from the normal mantle source (C value = 0.4–0.5) would not only displace the liquid towards a higher C value but also somewhat increase its MgO content (the change of path is shown by a broken arrow in figure 7*b*). Subsequent partial melting of this Al-depleted diapir, or olivine fractionation from the high CaO/Al_2O_3 melt would produce the same trend as earlier described for the Onverwacht rocks. The Finland–Rhodesia trend shows that rocks of MgO > 20% are slightly higher in C values than that of model mantle M. This is evidently due to the slight effect from the residual orthopyroxene, which generally has low CaO and Al_2O_3 contents (less than 1%) but a high Al_2O_3/CaO ratio (*ca.* 3–7). The skew of the trend towards low C values for rocks of MgO < 16% is consistent with clinopyroxene having joined olivine as an important liquidus phase and being fractionated from the basaltic komatiite liquids. It is important to note that no garnet role is involved in the evolution of the Finland–Rhodesia–Munro trend. The argument from r.e.e. data is equally convincing, for the rocks from Finland, Rhodesia and Munro generally have unfractionated heavy r.e.e. (figure 6), whereas the Onverwacht rocks show fractionated heavy r.e.e., suggesting a result of garnet separation (Sun & Nesbitt 1978).

Sun & Nesbitt (1978) mentioned that rocks of high CaO/Al_2O_3 and fractionated heavy r.e.e., $(Gd/Yb)_N > 1.0$, are only found in 3.5 Ga or older terrains. An example other than the Onverwacht samples is found in an Isua amphibolite (Sun & Nesbitt 1978). Rocks of 2.7 Ga or younger do not possess such a heavy r.e.e. depletion. At present we are not sure whether this represents an evolutionary change in upper mantle composition. Since we understand that garnet is unlikely to be a liquidus phase in peridotitic komatiite liquids, direct precipitation of garnet from 'normal' komatiite liquids to produce a komatite liquid of the Onverwacht type may encounter a problem. Nevertheless, the source heterogeneity, possibly created by garnet separation if the initial mantle is of the normal type, seems to exist in the Archaean, regardless of long-term or short-term effects.

An alternative interpretation of the high CaO/Al_2O_3 ratios in some komatiites was given by Cawthorn & Strong (1974), who envisaged a chemically layered mantle with a garnet rich zone (low CaO/Al_2O_3 ratio) underlying a clinopyroxene rich zone (high CaO/Al_2O_3 ratio). Partial melting at shallow depth (cpx.-rich zone) produces the Onverwacht type komatiites and melting at progressively deeper zones yields komatiites of lower CaO/Al_2O_3 ratios. This model is attractive in this respect. However, komatiites with lower CaO/Al_2O_3 ratios are inferred to be derived from the garnet-rich zone. They are expected to have r.e.e. patterns with heavy r.e.e. enrichment, i.e. $(Gd/Yb)_N < 1.0$. The fact that many Onverwacht rocks are depleted in heavy r.e.e. but have higher CaO/Al_2O_3 ratios, and many Finnish/Munro rocks have flat heavy r.e.e.

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but lower CaO/Al_2O_3 ratios, is not very consistent with the model of Cawthorn & Strong (1974). A problem thus still remains to be answered. If their model is finally proved to be correct, this would argue for a very large-scale and long-existing mantle heterogeneity.

5. CONCLUSIONS

The available chemical and isotopic data for Archaean igneous rocks seem to indicate a significant heterogeneity in their mantle sources. However, some caution must be used in the interpretation of these data. The I_{Sr} values show a wide range, regardless of acidic or mafic rocks, but a long-term large degree heterogeneity in Rb/Sr ratios in their sources cannot be determined with certainty. The I_{Pb} data show a narrower range in comparison with the I_{Sr} data. Their inferred μ_1 values range from 7.6 to 8.6. However, the implication of heterogeneous U/Pb or Th/Pb ratios in the mantle sources is very model-dependent. So far, most I_{Nd} values fall on or very near to the chondritic uniform reservoir evolution line. This observation reinforces a common assumption that the terrestrial mantle is characterized by having r.e.e. distributions of chondritic type. Nevertheless, a few igneous complexes, such as Stillwater and Duluth (DePaolo 1978; DePaolo & Wasserburg 1976*a*; McCulloch & Wasserburg 1978) do show a small but significant degree of deviation from the chondritic evolution line, i.e., $\epsilon_{ch.u.r.}(T) \neq 0$. Perhaps more similar examples will be found in the near future.

Regarding the trace element distributions, komatiites very depleted in light r.e.e. occur in Finland, Canada, western Australia and Rhodesia. These materials should provide good candidates to test by the Sm/Nd isotopic technique if long-term heterogeneity in Sm/Nd ratios and thus r.e.e. distributions did exist in the Archaean. The I_{Nd} data of Zindler *et al.* (1978) on some light r.e.e. depleted komatiites from the Munro area show a slight positive $\epsilon_{ch.u.r.}(T)$ value, but its error bars do intersect the chondritic reservoir evolution line.

If the light r.e.e. depletion in komatiites is found to be very widespread, the causal effect and the complementary part might be related to (1) the production of some contemporaneous light r.e.e. enriched tholeiitic magmas, (2) enrichment of light r.e.e. and other incompatible elements by adding some undersaturated melt to higher level mantle source regions which later produced light r.e.e. enriched rocks, or (3) contemporaneous or previous continental-forming events.

In terms of the major element data, we emphasize the significant differences in CaO/Al_2O_3 ratios in the source regions of komatiitic rocks. However, we are not very sure if this represents an evolutionary feature of the upper mantle or a true compositional heterogeneity at a given time in the Archaean. Taking the possible evolutionary change and heterogeneity into consideration, the common practice of identifying tectonic settings for Archaean greenstone belts by comparing their trace element abundances (even the refractory r.e.e.) with those of modern volcanic rocks should be subject to a more cautious examination.

REFERENCES (Jahn *et al.*)

Arndt, N. T. 1976 *Carnegie Instn Wash. Yb.* **75**, 555–562.
Arndt, N. T., Naldrett, A. J. & Pyke, D. R. 1977 *J. Petr.* **18**, 319–369.
Arth, J. G., Arndt, N. T & Naldrett, A. J. 1977 *Geology*, **5**, 590–594.
Baadsgaard, H. 1976 *Earth planet. Sci. Lett.* **33**, 261–267.
Bickle, M. J., Martin, A. & Nisbet, E. G. 1975 *Earth planet. Sci. Lett.* **27**, 155–162.

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Blais, S., Auvray, B., Bertrand, J. M., Capdevila, R., Hameurt, J. & Vidal, P. 1977*a* *Bull. Soc. géol. Fr.* **5**, 1033–1039.

Blais, S., Auvray, B., Capdevila, R. & Hameurt, J. 1977*b* *Bull. Soc. géol. Fr.* **5**, 965–970.

Blais, S., Auvray, B., Capdevila, R., Jahn, B. M., Bertrand, J. M. & Hameurt, J. 1978 In *Archean geochemistry* (ed. B. F. Windley & S. M. Naqvi), pp. 87–107. Amsterdam: Elsevier.

Carlson, R. W., Macdougall, J. D. & Lugmair, G. W. 1978 *U.S. geol. Surv. Open-File Rep.* no. 78–701. (Snowmass Meeting), pp. 58–60.

Cawthorn, R. G. & Strong, D. F. 1974 *Earth planet. Sci. Lett.* **23**, 369–375.

Cooper, J. A., Nesbitt, R. W., Platt, J. P. & Mortimer, G. E. 1978 *Precambrian Res.* **7**, 31–59.

DePaolo, D. J. 1978 (Abst.) *Geol. Soc. Am. Annual Meeting*, Toronto, vol. 10, p. 388.

DePaolo, D. J. & Wasserburg, G. J. 1976*a* *Geophys. Res. Lett.* **3**, 249–252.

DePaolo, D. J. & Wasserburg, G. J. 1976*b* *Geophys. Res. Lett.* **3**, 743–746.

DePaolo, D. J. & Wasserburg, G. J. 1977 *Geophys. Res. Lett.* **4**, 465–468.

Gancarz, A. J. & Wasserburg, G. J. 1977 *Geochim. cosmochim. Acta* **41**, 1283–1301.

Green, D. H. 1975 *Geology* **3**, 15–18.

Hamilton, P. J., O'Nions, R. K., Evensen, N. M., Bridwater, D. & Allaart, J. H. 1978 *Nature, Lond.* **272**, 41–43.

Hart, S. R. & Davis, G. L. 1969 *Bull. geol. Soc. Am.* **80**, 595–614.

Hart, S. R. & Brooks, C. 1977 *Contr. Miner. Petr.* **61**, 109–128.

Hawkesworth, C. J., Moorbath, S., O'Nions, R. K. & Wilson, J. F. 1975 *Earth planet. Sci. Lett.* **25**, 251–262.

Hawkesworth, C. J. & O'Nions, R. K. 1977 *J. Petr.* **18**, 487–520.

Herrmann, A. G., Blanchard, D. P., Haskin, L. A., Jacobs, J. W., Knake, D., Korotev, R. L. & Brannon, J. C. 1976 *Contr. Miner. Petr.* **59**, 1–12.

Hurst, R. W. 1978 *Geochim. cosmochim. Acta* **42**, 1585–1586.

Jacobsen, S. B. & Wasserburg, G. J. 1978 *Earth planet Sci. Lett.* **41**, 245–253.

Jahn, B. M. & Murthy, V. R. 1975 *Geochim. cosmochim. Acta* **39**, 1679–1689.

Jahn, B. M. & Nyquist, L. E. 1976 In *The early history of the Earth* (ed. B. F. Windley), pp. 55–76. London: John Wiley & Sons.

Jahn, B. M., Auvray, B., Blais, S., Capdevila, R., Cornichet, J., Vidal, F. & Hameurt, J. 1980 *J. Petr.* (In the press.)

Kyser, T. K. & O'Neil, J. R. 1978 *U.S. geol. Surv. Open-File Rep.* 78–701 (Snowmass Meeting), pp. 237–240.

McCulloch, M. T. & Wasserburg, G. J. 1978 *Science, N.Y.* **200**, 1003–1011.

Moorbath, S. 1978 *Geochim. cosmochim. Acta* **42**, 1583–1584.

Naldrett, A. J. & Turner, A. R. 1977 *Precambrian Res.* **5**, 43–103.

Nesbitt, R. W. 1972 *Geol. Soc. Aust. spec. Pub.* **3**, 331–347.

Nesbitt, R. W. & Sun, S. S. 1976 *Earth planet. Sci. Lett.* **31**, 433–453.

Nisbet, E. G., Bickle, M. J. & Martin, A. 1977 *J. Petr.* **18**, 521–566.

O'Nions, R. K., Hamilton, P. J. & Evensen, N. M. 1977 *Earth planet. Sci. Lett.* **34**, 13–22.

Oversby, V. M. 1975 *Geochim. cosmochim. Acta* **39**, 1107–1125.

Oversby, V. M. 1978 *Earth planet. Sci. Lett.* **38**, 237–248.

Pyke, D. R., Naldrett, A. J. & Eckstrand, O. R. 1973 *Bull. geol. Soc. Am.* **84**, 955–978.

Rosholt, J. N., Zartman, R. E. & Nkomo, I. T. 1973 *Bull. geol. Soc. Am.* **84**, 989–1002.

Stacey, J. S. & Kramers, J. D. 1975 *Earth planet. Sci. Lett.* **26**, 207–221.

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**, 301–325.

Taylor, P. N. 1975 *Earth planet. Sci. Lett.* **27**, 35–42.

Taylor, P. N. 1978 *Earth planet. Sci. Lett.* (In the press.)

Vidal, P., Tilton, G. R., Blais, S., Jahn, B. M. & Capdevila, R. 1980 Submitted to *Geochim. cosmochim. Acta.*

Viljoen, M. J. & Viljoen, R. P. 1969 *Geol. Soc. S Africa spec. Pub.* **2**, 87–112.

Whitford, D. J. & Arndt, N. T. 1978 *Earth planet. Sci. Lett.* **41**, 188–196.

Zindler, A., Brooks, C., Arndt, N. T. & Hart, S. 1978 *U.S. geol. Surv. Open-File Rep.* no. 78–701, pp. 469–471.