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Paradoxes of the mantle lithosphere underneath Archaean continents and models for its origin

by *Jan D. Kramers*[1]

Abstract

A brief overview is presented of the particular features of the mantle lithosphere underneath Archaean cratons (as exemplified by the South African and the Siberian case) that render it different from the "normal" mantle underneath ocean basins. These are: (1) Lithospheric character (rigidity) down to a depth of at least 150, and possibly 400 km; (2) predominantly harzburgitic petrography but with high Mg-numbers betraying a residual nature; (3) anomalous enrichment in incompatible trace elements in the harzburgites in spite of this residual nature; (4) occurrence of various types of eclogites; (5) early Proterozoic or even Archaean age indications; and (6) patchy metasomatism and veining which is relatively young.

Models for the origin of the bulk of the material, i.e. the harzburgites, are discussed in the light of known petrological and physical constraints. An explanation for the residual yet harzburgitic character of these rocks may lie in an extensive melting process at high pressure and under water-undersaturated conditions, but there are as yet no successful models which can account for *both* their major element composition and their incompatible trace element enrichment in a single process. In following up the suggestion that the pervasive incompatible element enrichment might be secondary and caused by fluids or melts moving through the lithosphere, it is shown that H_2O–CO_2 fluids could not have caused pervasive metasomatism in the mantle lithosphere because (1) they would preferably move along fractures, and (2) their buffered composition along the depth profile of the subcontinental mantle would only allow trace element enrichment within the amphibole stability field, i.e. at depths less than about 75 km. It is concluded that the pervasive incompatible trace element enrichment must be caused by melts. It can therefore have taken place only at a time when the ambient p-T conditions of the subcontinental mantle were above the solidus of such melts. This means that this enrichment must be, by definition, as old as the lithospheric mantle keel itself. Patchy metasomatism, in contrast, can be relatively young, but is limited to the vicinity of veins and probably to near the base of the lithosphere.

Keywords: Mantle lithosphere, craton, Archaean, harzburgite, trace elements, melt.

1. Introduction

The lithospheric mantle underneath Archaean cratons has geophysical, petrological and geochemical characteristics which set it apart from the suboceanic or "normal" upper mantle at comparable depths, and some of which appear paradoxical.

The seismic evidence includes surface wave dispersion (Jordan, 1981), and shear wave velocity data (Grand and Helmberger, 1984, 1985) from which a rigid lithospheric character down to a depth of at least 150 km, and possibly as deep as 400 km (Jordan, 1981), can be inferred. The petrological and geochemical evidence is obtained from xenoliths in kimberlites and similar volcanic rocks (Dawson, 1980). These include a majority of (mainly harzburgitic) peridotites, along with a variety of eclogitic rocks. There is a contradiction between the major element chemical characteristics of the peridotites, which reflect a residual nature (e.g. Boyd, 1989) and a relative enrichment in incompatible elements (e.g. Harte, 1983). The eclogites, which can be diamondiferous, can have features reminiscent of metamorphosed subducted oceanic crust on the one hand, and of a high temperature cumulate origin on the other hand (Smyth and Caporuscio, 1984; MacGregor and Manton, 1986; Neal et al., 1990).

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It is the aim of this paper to review broadly the evidence for the nature of the lithosphere between Archaean continents, to highlight the main problems and to discuss models for its formation and evolution.

2. Generalized characteristics of the mantle lithosphere underneath Archaean continental crust: the problems

2.1. GEOPHYSICAL CONSTRAINTS

On the basis of surface wave velocity dispersions, Jordan (1981) has postulated that the subcontinental mantle is more rigid than the suboceanic mantle, down to a depth of ca. 400 km, for large continental masses such as North America and Africa. Such a greater rigidity (or: greater seismic velocities) would imply a lower temperature for the subcontinental mantle than for the suboceanic mantle at the same depth. If the same chemical composition was assumed for both types of mantle, this would mean a greater density for the subcontinental than for the suboceanic mantle. Non-orogenic continental areas such as Southern Africa would then be mainly submerged. Instead, they may be up to 2000 m above sea level, an elevation which cannot be accounted for by their sialic component alone.

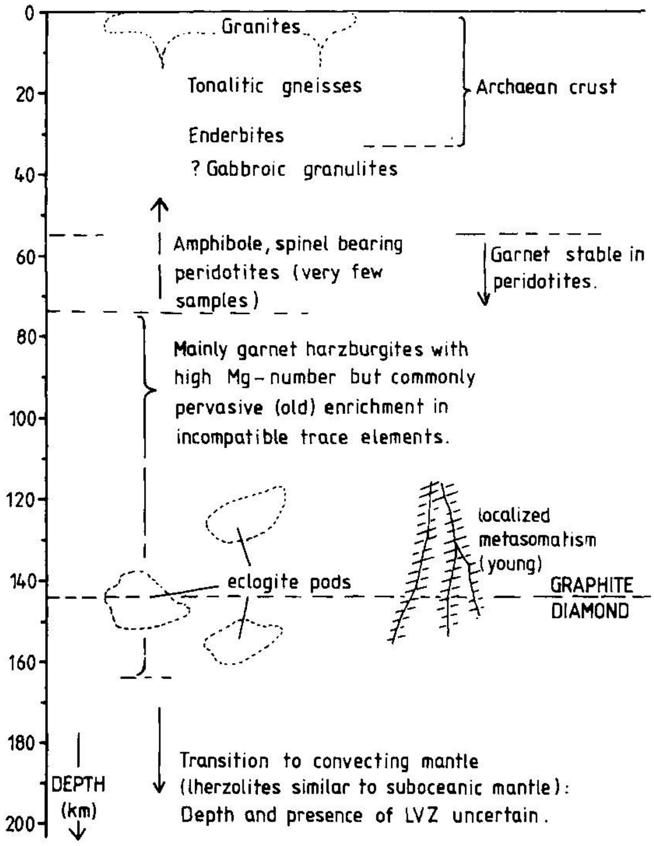

Fig. 1 Schematic cross section through Archaean continental crust and subcontinental mantle lithosphere, to its confirmed thickness. Position of graphite-diamond boundary after intersection of continental geotherm of Clark and Ringwood (1964) and inversion line of Kennedy and Kennedy (1976; see Fig. 2).

Therefore Jordan (1981) has suggested that there must be a chemical difference between subcontinental and suboceanic mantle, whereby the former would have a lower density than the latter if temperatures were the same, so that this paradox can be explained. Such a mantle lithosphere which is chemically distinct from the suboceanic mantle is regarded by Jordan (1981) as intrinsic to large continental masses, and he has proposed the term "continental tectosphere" for the mechanically coupled crust-plus-subcontinental mantle.

More direct geophysical evidence for the presence of this type of lithospheric keel is found in upper mantle shear velocities higher than those for suboceanic mantle lithosphere, down to a depth of 150–200 km under the Canadian, the Russian and the West Siberian shields (Grand and Helmberger, 1984; 1985). These authors distinguish a low velocity zone underneath this lithosphere, while Lerner-Lam and Jordan (1987) claim there is no such zone at this depth. The thickness of subcontinental mantle lithosphere is therefore still a matter of controversy.

2.2. PETROLOGICAL FEATURES OF XENOLITHS FROM THE SUBCONTINENTAL MANTLE LITHOSPHERE

Data on xenoliths from the mantle lithosphere at depth underneath Archaean crust are clearly greatest in abundance in areas where kimberlites have been extensively mined, e.g. in southern Africa and Yakutia, Siberia (e.g. Dawson, 1980; Sobolev, 1974). A schematic cross section of the continental tectosphere (Archaean crust plus underlying mantle lithosphere keel) as derived from such data is presented in Fig. 1; there is a paucity of material from depths between the base of the crust and about 70 km, and the present discussion refers entirely to xenoliths originating from below that depth.

The majority of mantle derived xenoliths from kimberlites in South Africa are coarse grained peridotitic rocks, mainly harzburgites. A summary of their petrographic and geochemical characteristics is presented by Harte (1983). Some salient major element trends and differences between oceanic mantle lithosphere and peridotitic mantle underlying Archaean cratons have been highlighted for instance by Boyd (1989). Whereas oceanic lithosphere shows a correlation between olivine Mg number and olivine content which is consistent with an origin as residues of variable shallow-level partial melting, samples from the lithosphere below the Archaean Kaapvaal craton have high olivine Mg numbers (around 92–93) for rocks that are actually harzburgites, and no trend is evident (this observed chemical difference corresponds to that postulated by JORDAN [1981] as described in section 2.1. above). Xenoliths from kimberlites outside the Archaean craton show a trend similar to the oceanic one. Some peridotitic xenoliths which appear to have come from the greatest depth, in excess of 200 km, are likewise not dissimilar to suboceanic mantle in both major and trace element geochemistry (NIXON and BOYD, 1973; SHIMIZU, 1974), and these will also not be discussed further in this paper.

A feature of some peridotitic mantle xenoliths which can be seen in both kimberlite and alkali basalt hosted associations is the occurrence of hydrous minerals such as phlogopite and amphiboles including the high pressure amphibole potassic richterite (ERLANK, 1970), as well as obvious veins and pervasive impregnations of mica, amphibole, rutile, ilmenite and diopside (the "MARID" suite, DAWSON and SMITH, 1977; KRAMERS et al., 1983). These represent clear examples of metasomatic activity in which minerals of the original rock can be distinguished from those formed by an infiltrating melt or fluid.

Eclogites, which are somewhat patchily distributed, are of a type different from that found in orogenic belts (e.g. HARTE and GURNEY, 1975; SOBOLEV et al., 1968). They range from bimineralic eclogites to grospydites, which consist of approximately equal amounts of grossular, pyroxene and kyanite. Eclogites are much more frequently diamondiferous than peridotites.

The evidence concerning their origin is clearly paradoxical. On the one hand, exsolution textures and concomitant chemistry indicate original near-liquidus clinopyroxene which unmixed as it cooled down; there is also a series of compositions which can be explained by progressive fractionation of the melt phase. These features are not consistent with an origin by prograde metamorphism of oceanic crust in a subduction zone: A high pressure-high temperature cumulate origin appears more likely (HATTON and GURNEY, 1977; SMYTH and CAPORUSCIO, 1984). On the other hand, the large range of $\delta^{18}O$ values from +2.2 to +8.5 found in eclogite xenoliths is in stark contrast to the homogeneous $\delta^{18}O$ values around 5.6 displayed by the peridotites (MCGREGOR and MANTON, 1986; GARLICK et al., 1971; NEAL et al., 1990). The values appear to indicate that at least some low temperature or near-surface process must have played a part in the prehistory of the eclogites. This is supported by the large range in $\delta^{13}C$ values for diamonds associated with these eclogites (–30 to 0, KIRKLEY and GURNEY, 1989). The eclogite $\delta^{18}O$ range does not extend to values as high as those found in ocean crust as exemplified by ophiolites (GREGORY et al., 1981), but their upper bound is strikingly similar to the very uniform $\delta^{18}O$ value found for Archaean granite-gneiss terrains (SHIEH and SCHWARCZ, 1978; WILSON, 1981; BIBIKOVA, 1984).

2.3. TRACE ELEMENT CHARACTERISTICS OF SUBCONTINENTAL MANTLE LITHOSPHERE, AND AGE INFERENCES

The trace element and radiogenic isotope geochemistry of the subcontinental mantle peridotites is most interesting, and has been reviewed by HARTE (1983) and MENZIES et al. (1987). These rocks appear to have mostly relatively light-enriched rare earth element patterns (e.g. SHIMIZU, 1974), and elevated contents of large ion lithophile elements relative to the primitive mantle. This is particularly so for peridotites which show metasomatism or infiltration, but strong relative light rare earth element (LREE) enrichment (albeit with heavy REE depletion relative to chondrites) is also found in samples without such features. DAWSON (1982) distinguishes "patent" and "cryptic" mantle metasomatism. In the former, incompatible trace element enrichment accompanies visible modal metasomatism, whereas the latter constitutes trace element enrichment on its own, unaccompanied by hydrous phases or any other visible evidence of metasomatism – in this case the metasomatic origin of the trace element patterns is merely inferred.

Trace element trends of eclogites are very different from those of the peridotites; they do not show relative light rare element enrichment, and their rare earth patterns are not dissimilar to those found in mid-ocean ridge basalts and gabbro complexes in ophiolites (CAPORUSCIO and SMYTH, 1986).

Because samples from the subcontinental mantle are generally brought up to the surface by volatile-rich magmas, and their geochemistry might have been altered along grain boundaries during ascent, age determinations are problematical. Peridotites which show patent metasomatism and a corresponding overall enrichment in incompatible elements have given a suite age not significantly older than the host kimberlite (90 Ma, KRAMERS et al., 1983), and the Nd isotope signatures of veins and their patently metasomatised host rocks are those of primitive or slightly depleted mantle, again similar to those of the host kimberlites (ε_{Nd} values between 0 and +7, KRAMERS et al., 1983; RICHARDSON et al., 1985; MENZIES et al., 1987). This indicates that at least some of these metasomatic events might be a manifestation of the kimberlite magmatism itself, at depth.

However, clinopyroxene ε_{Nd} values of peridotites from the southern African subcratonic lithosphere showing only "cryptic metasomatism" or very minor hydrous phases range from 0 to –15 (MENZIES and MURTHY, 1980), and there are also some surprisingly high $^{87}Sr/^{86}Sr$ ratios (e.g. 0.713 in a clinopyroxene, unsupported by Rb/Sr, KRAMERS, 1977). Such mineral isotope ratios presum-

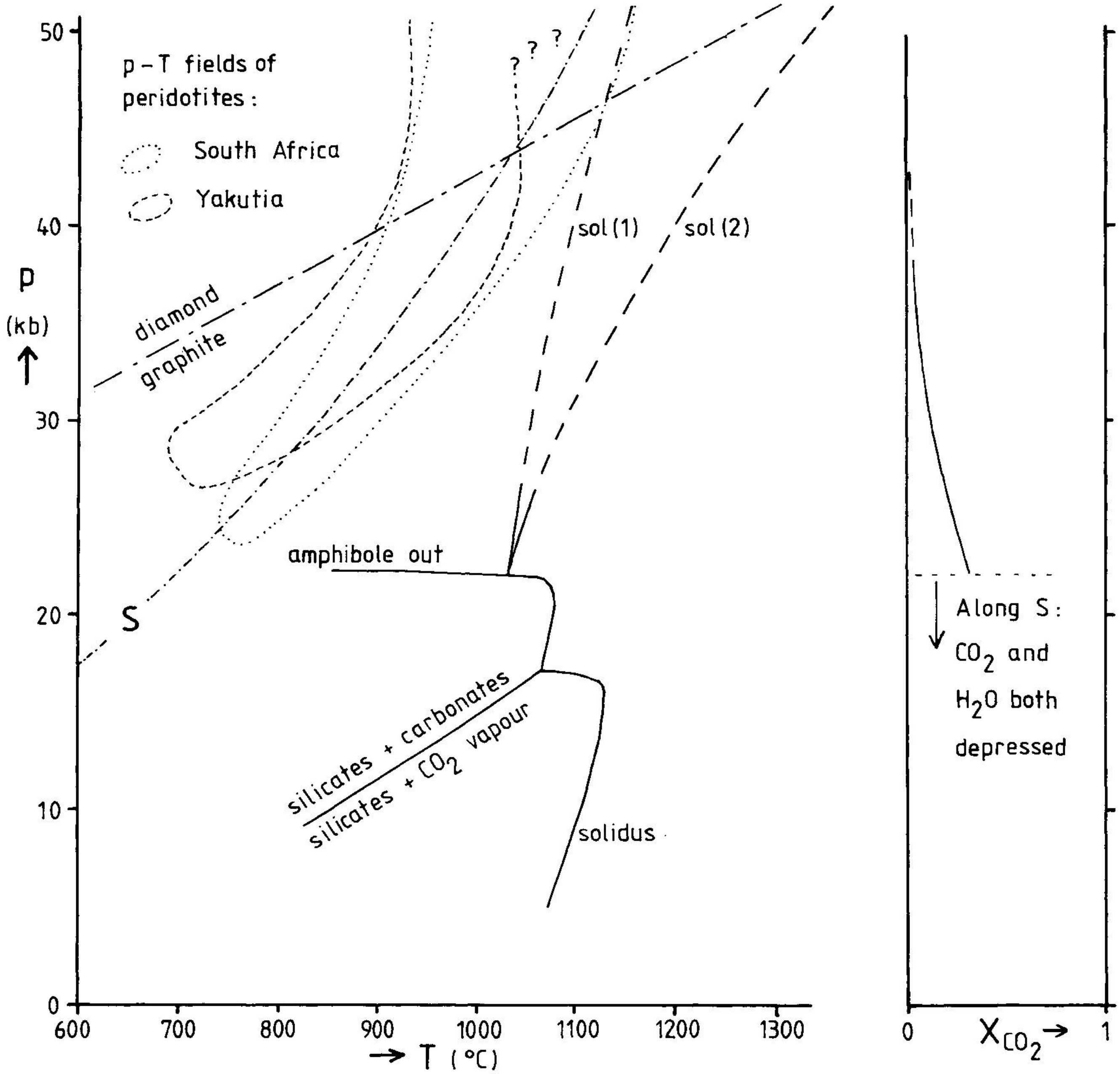

Fig. 2 p-T diagram showing peridotite + CO_2 + H_2O (undersaturated) solidi, carbonation curve and amphibole breakdown curve after OLAFSSON and EGGER (1983), Precambrian Shield geotherm (S) after CLARK and RINGWOOD (1964), diamond-graphite boundary after KENNEDY and KENNEDY (1976), and p-T fields for South African and Yakutian peridotite xenoliths in kimberlite after NICKEL and GREEN (1985). Question marks at high pressure-high temperature end of the Yakutian field indicate temperature upturn which is still the subject of debate. At high pressure two solidus curves are drawn: Sol (1) applies to dehydration melting of peridotite with minor phlogopite, sol (2) is for melting in the presence of approximately 1% H_2O+CO_2 vapour. Also shown at right is the CO_2 mole fraction of approximately 1% H_2O+CO_2 vapour in equilibrium with the peridotite along the shield geotherm, in part after SCHNEIDER and EGGLER (1986).

ably reflect those of the whole rock before it was sampled by the kimberlite magma. They cannot be evaluated quantitatively in terms of ages, because whole rock element ratios have been impossible to obtain. A minimum T_{DM} age around 2 Ga is suggested by the lowest ε_{Nd} values of MENZIES and MURTHY (1980). In this subcratonic lithosphere the anomalous trace element characteristics therefore appear to be ancient, certainly Proterozoic and possibly Archaean. It should be stressed that this is not necessarily the case for geochemically anomalous lithosphere underlying other, younger terrains (MENZIES et al., 1987).

Because of their greater geochemical diversity and generally higher trace element concentrations, eclogite xenoliths have provided better opportunities for dating than peridotites. Clinopyroxene lead isotope data (KRAMERS, 1979) gave an age of roughly 2.5 Ga for the eclogite suite from Roberts Victor, South Africa, whereas Sm–Nd data have given early Proterozoic (NEAL et al., 1990) and early Archaean (JAGOUTZ et al., 1984; MCCULLOUGH, 1986) T_{CHUR} model ages. Although the data were obtained on mineral separates, these ages have the significance of whole rock ages and may be interpreted as dating the petrogenesis of the eclogites, either as magmatic cumulates or by metamorphism of a subducted ocean floor basalt. Because of the small size of individual samples, no zircon studies similar to those done on alpine-type eclogites (GEBAUER, 1990) have yet been possible.

Very interesting and spectacular data have also been obtained on inclusions in diamonds. Pb isotope results on composite samples of sulphide inclusions in diamonds from the Finsch Mine and the cluster of diamond pipes in Kimberley, South Africa (KRAMERS, 1979) indicated model ages certainly in excess of 2000 Ma, although the kimberlite pipes in question are only around 90 Ma old. Nd and Sr isotope data obtained on composite samples of garnet and clinopyroxene inclusions from the same locality, which were of peridotitic mineralogy (RICHARDSON et al., 1984) indicated an age of origin of 3300 Ma for these diamonds. Although details of the interpretation have been the subject of a controversy, there can be no doubt that the diamonds are of Archaean age, and formed in a rock province which already had a relative light rare earth enrichment.

2.4. INFERENCES FROM THE GEOCHEMISTRY OF VOLCANIC ROCKS

From anomalous trace element and particularly radiogenic isotope characteristics of young mafic volcanic and dyke rocks in Archaean cratonic areas, the presence of mantle lithosphere which was enriched in incompatible trace elements has been inferred in a number of cases. The type locality is again southern Africa, where the evidence from volcanics and mantle-derived xenoliths can be tied together (HAWKESWORTH et al., 1983), and some tests for crustal contamination have been carried out (e.g. ELLAM and COX, 1989). There are many examples of anomalous trace element and isotope characteristics in mafic intrusives and volcanics, such as from the Columbia River basalts (CARLSON, 1984; VOLLMER et al., 1984) and Scotland (WATERS et al., 1990) in which the existence of ancient anomalous subcontinental mantle material is implied, but in most cases the question of crustal contamination is not entirely resolved. In any case such "enriched-source" volcanics are not confined to Archaean cratons (e.g. ROGERS et al., 1985, for central Italy; KRAMERS et al., 1981, for the West Eiffel) and they even occur in oceanic provinces such as the Azores (HAWKESWORTH et al., 1979).

An interesting case is presented by PEGRAM (1990) for the Mesozoic Appalachian tholeiites, which yielded whole rock Rb–Sr, Sm–Nd and Pb/Pb ages of c. 1000 Ma on a regional basis, while it could be shown that this was not an effect of crustal contamination. This age is quite similar to the date of an important crust forming event in the area. This suggests that subcontinental mantle keels may be similar in age to the overlying crust in areas other than Archaean cratons.

2.5. GEOTHERMAL GRADIENTS AT THE TIME OF KIMBERLITE ERUPTION... AND IN THE PAST

NICKEL and GREEN (1985) have presented a critical review of the geothermometric and geobarometric work done on peridotite xenoliths in kimberlites and similar rocks (Fig. 2). In the two areas of large-scale diamond mining, South Africa and Yakutia, Siberia, similar p-T conditions prevailed which, down to a depth of 150 km, lie within 100 °C of the shield geotherm proposed by CLARK and RINGWOOD (1964), reaching temperatures between 900 and 1100 °C at 150 km – conditions about 150–200 °C below the H_2O–CO_2 undersaturated peridotite solidus at that depth (OLAFSSON and EGGLER, 1983). At depths greater than 150 km, suites from both areas display a not entirely confirmed, but much discussed, isobaric shift to several 100° higher temperatures (up to 1400 °C for South Africa). If this is a true shift, it can only reflect a transient situation which might be connected to kimberlite formation or ascent, and it is not discussed further in this paper.

Further p-T evidence comes from the diamonds themselves, not merely from their origin within the diamond stability field, but also from p-T work on diamond bearing peridotite (Shee et al., 1982) and from the paragenesis of mineral inclusions in diamonds (Hervig et al., 1980), which indicate the same range of temperatures, 900–1000 °C.

Along with a lithospheric origin of many diamonds, a lithospheric thickness of at least 150 km, as postulated by the geophysical work mentioned above, has thus been fairly well established. The lithosphere may be somewhat thicker, as kimberlite formation may have been preceded by some thermal erosion at the base of the lithosphere, but there is so far no petrological evidence of a 400 km thickness as suggested by Jordan (1981).

From the relative positions of the recorded Cretaceous (South Africa) or Palaeozoic (Siberia) pressure-temperature conditions and the likely H_2O–CO_2 undersaturated peridotite solidus (see Fig. 2), the amount of cooling that can have occurred at 100–150 km depth in the subcontinental mantle since the late Archaean or early Proterozoic is only 100–200 °C. The diamond-graphite boundary is also probably subsolidus throughout the depth range under consideration, and diamonds in peridotites may have formed from graphite in a subsolidus, retrograde reaction, as the geotherm gradually changed: They should thus be younger than the host rocks, and this is in agreement with the observation of Richardson et al. (1984) that the mantle environment in which the apparently 3.3 Ga old diamonds formed already had an incompatible trace element enriched signature.

From considerations of heat flow decrease it would be possible for continental lithosphere of a thickness well over 100 km to be stabilized purely from differences in thermal regimes since about 2.5 Ga ago (Richter, 1988). Prior to that, a compositional difference was necessary: Only more refractory material could have survived as thick mantle lithosphere. Following this logic, the anomalous major element chemistry and old age of the lithospheric mantle underneath Archaean cratons are intrinsically coupled: If it did not have the refractory major element composition, it simply could not be so ancient.

3. Theories for the formation and evolution of subcontinental mantle lithosphere

3.1. GENERAL CHARACTERISTICS WHICH HAVE TO BE ACCOUNTED FOR

Any hypothesis addressing the origin and evolution of the mantle lithosphere underneath Archaean cratons has to address all of the following features observed in this lithosphere:

- "Depleted" major element chemistry of the peridotitic rocks.
- "Enriched" incompatible trace element chemistry of the peridotitic rocks, coupled with
- absence of hydrous minerals in many of the rocks.
- Frequently observed old ages of the incompatible trace element enrichment: possibly comparable to ages of Archaean continent formation or at least stabilization.
- Presence of ancient eclogite pods embedded in the subcontinental mantle, which do not display the incompatible trace element enrichment.
- Coincidence of the above features with Archaean cratons.

While there is clear evidence for infiltration or metasomatism on a limited spatial scale, as discussed above, the origin of "cryptic" incompatible trace element enrichment by a metasomatic process is merely inferred. Since the ultimate origin of the mantle lithosphere underneath Archaean cratons must be magmatic, it is tempting to seek magmatic models which explain its major *and* trace element characteristics in a single process. Such efforts are briefly reviewed in the following section.

3.2. SUGGESTIONS FOR MAGMATIC MODELS

Herzberg et al. (1988) have put forward the interesting suggestion that the South African peridotitic mantle samples represent the result of fractionation of majorite (a high pressure mineral with a garnet-type structure and a composition close to that of enstatite) from an original melt of chondritic composition. This implies that a very ancient process, related to a hypothetical, approximately 500 km deep magma ocean which might have existed following the accretion of the Earth, could still be witnessed by these samples.

The effect of majorite fractionation on REE patterns is likely to be some HREE depletion and LREE enrichment in the remaining melt (Kato et al., 1989), just as seen in the South African xenoliths, so that this hypothesis has considerable appeal. However, it raises two questions: First, why is this type of peridotite, which is relatively refractory and has a relatively low density, preserved under Archaean continental crust and nowhere else? Second, why is there no trace element geochemical evidence of a global magma ocean in the mantle outside continental masses (KATO et al., 1989; DRAKE et al., 1988)? In addition to this it must be noted that, while subcontinental peridotites are undoubtedly old, no evidence of dates close to the accretion age of the Earth has been found in them.

An attempt to connect the formation of the subcontinental mantle to Archaean continent formation is the MARCY model (KRAMERS, 1987) in which it is suggested that (a) Archaean continents formed by open-system near steady-state fractionation of near-surface very extensive magma layers, and (b) the subcontinental mantle is derived from cumulates of this steady state process as the ultimate endproduct of repeated remelting accompanied by garnet settling. Trace element geochemical features of the subcontinental peridotites are thereby approximated. This treatment of the subcontinental peridotitic lithosphere suffers from two weaknesses: First, there is no evidence of a gradual change from the postulated cumulate (which has a tholeiitic composition) to peridotite, and second, the process suggested to achieve this transition in the original MARCY model is somewhat contrived. However, the "crustal" part of this model may explain many geochemical and other features of the earliest continental crust (KRAMERS, 1988; RIDLEY and KRAMERS, 1990), and it offers a possible alternative explanation for the occurrence of eclogite pods in the subcontinental mantle: They might be derived from remolten cumulate underneath a steady state tonalitic magma layer. This could to some extent accommodate the above observations on their $\delta^{18}O$ values and the $\delta^{13}C$ values of eclogitic diamonds.

Models such as those discussed above, which seek to explain the characteristics of the major *and* trace element characteristics of peridotitic lithosphere underneath Archaean continents in one single magmatic process, appear to have been unsuccessful so far. However, before the possible metasomatic processes are examined, models intended to account for the *major* element composition of peridotitic subcratonic mantle lithosphere are now reviewed.

The refractory (depleted) chemical characteristics of the peridotites has tempted geophysicists (e.g. JORDAN, 1981) to view them as residues more or less directly connected to the formation of Archaean continental crust. The nature of the depletion is not addressed, but a partial melting process is implied, which would in the first instance give rise to a mafic to intermediate melt. In an examination of this type of residue hypothesis using CMAS projections, HERZBERG et al. (1988) conclude that the peridotites are both too Si-rich and too close to the S-M join of the tetrahedron to be residues of such melts.

This objection against a residue nature does not hold if the melt removed is ultrabasic. BOYD (1989) has conducted successful mass balance calculations modelling the removal of large fractions (up to 50%) of komatiite-type magmas from model primitive mantle compositions. He emphasises the abovementioned coincidence of Mg numbers of 92–93 for olivines with relatively low modal olivine content (50–80%) for the peridotites. This would indicate extensive melting, but the melt extracted would have to be ultramafic, with (Mg+Fe)/Si ratios intermediate between opx and olivine, otherwise a dunitic residue would be left. Unless melting occurs under completely dry conditions, the production of an ultramafic melt leaving a harzburgitic residue is only possible at pressures above about 22–25 kb, the stability limit of most amphiboles, and under water-undersaturated conditions (GREEN, 1973; MYSEN and BOETTCHER, 1975, 1976). The observed depth of origin of the subcontinental peridotite xenoliths fits these constraints.

The question arises where a very large volume of ultramafic melt might have gone to, since komatiites form only a very small proportion of Archaean volcanic rocks. BOYD (1989) advances the possibility that the residue might have had a slightly smaller density than the melt at these high pressures, and might have floated up against the continent in statu nascendi, while the ultramafic melt itself could have been incorporated back into the convecting mantle. Melting experiments of SCARFE and TAKAHASHI (1986) indicate that the residue might float at pressures above ca. 60 kb in the anhydrous case where solidus and liquidus temperatures are 1700 and 1900 °C, respectively. The hydrous system has not yet been investigated at these pressures. There are many possibilities for the physical behaviour of a partially molten system of the extent and nature considered here, and a detailed discussion would require thermal modelling, which is outside the scope of this paper. However, it is certain that melt would be separated upwards for the upper 150 km or so of subcontinental mantle. According to RIDLEY and KRAMERS (1990), the nascent continent would in many instances not be solid enough to allow eruption of magmas through cracks. They might have formed a separate magma layer that moved horizontally.

Even without detailed modelling, it is probably valid to state that the difference in major element chemistry between peridotites in the lithosphere underneath Archaean cratons and those in suboceanic lithosphere is a result of their different locations of origin. The former resulted as a residue from extensive melting under high pressure, blanketed underneath less dense sialic material. Oceanic crust was generated in a convective regime (whether one-layer or two-layer) without a compositional boundary below earth surface, where even in the Archaean the bulk of solidus overshoot of upwelling hot mantle material would occur at depths less than 60 km (see e.g. CAMPBELL and JARVIS, 1984), leading to mafic rather than ultramafic melts and correspondingly more olivine-rich residues.

3.3. METASOMATIC MODELS

Via the magmatic process outlined above, the peridotite lithosphere underneath Archaean crust with its particular major element characteristics could have become established approximately at the end of the Archaean. We next have to discuss the processes whereby trace element enrichment could come about. The possibilities are, first, metasomatism by a fluid phase, and second, metasomatism by a melt phase. For reasons of density (bar the possibility of density inversions at the base of the lithosphere) it is assumed that both types of metasomatic agent would be derived from the convecting mantle underneath the lithosphere and would move upward through it. Constraints on metasomatic models involving fluids and melts, and the major difficulties arising from these, are discussed below.

3.3.1. Composition of fluids in equilibrium with peridotite

Phase equilibria and the chemistry of fluids in equilibrium with mantle peridotites have been studied by OLAFSSON and EGGLER (1983) and SCHNEIDER and EGGLER (1986). The following discussion is based on their work and refers to Fig. 2:

The p-T conditions found for lithospheric peridotites underneath Archaean continents are subsolidus throughout, and lie in the region of p-T space where carbonation reactions buffer the CO_2 activity, keeping it generally low. In addition, hydration reactions occur in the stability field of amphibole, at pressures below around 22 kb, and if sufficient K is present to produce phlogopite, at much higher pressures as well. These buffer the activity of H_2O. The composition of the fluid phase as buffered in case of a total of 1% fluid (CO_2 + H_2O) is shown in Fig. 2 for pressures above 22 kb.

The chemical composition of solutes in fluids of this changing subsolidus composition is highly variable (SCHNEIDER and EGGLER, 1986). H_2O solutes (up to several %) are quartz-feldspar normative, whereas CO_2 bearing fluids have very much less solute, which is peralkaline in character. WENDLANDT and HARRISON (1979) studied the partitioning of rare earth elements between melt and a CO_2-rich fluid phase. They showed that REE solubility is high but decreases with decreasing pressure, and argued that such a phase might carry REE upward into the lithosphere from underlying mantle, and deposit them. However, CO_2 activity increases with decreasing pressure, so that decreasing solubility of REE alone is not enough to draw that conclusion. Further SCHNEIDER and EGGLER (1986) argue that, if a fluid moved upwards through the subcontinental mantle, leaching of minor and trace elements might occur in the lower portion (below the region of amphibole stability), whereas these would be redeposited in the upper part of the lithosphere, where the fluid would become depleted in H_2O due to the formation of amphibole. For a fluid moving along a p-T path wholly within the carbonate stability field (as represented by the shield geotherm, see Fig. 2) this effect would be extreme, since at pressures below about 22 kb the activity of both CO_2 and H_2O would be buffered to low levels.

It is however apparent that only a very small portion of the subcontinental mantle column could be enriched in incompatible elements by this type of process: The region within the amphibole stability field, whereas incompatible trace element depletion would be observed at greater depths. The characteristic "cryptic metasomatism" has, in contrast, largely been observed in amphibole free, garnet bearing rocks. The hypothesis does therefore not fit the observations, and upward moving fluid phases would appear to be unlikely metasomatic agents in the deep subcontinental lithosphere.

Because of these obvious problems, it is more likely that the observed metasomatic features would have been caused by a melt phase than by a fluid phase. Similar observations have been made by SCHNEIDER and EGGLER (1986), MENZIES et al. (1987) and MCKENZIES (1990). Before the role of fluids and melts can be assessed more fully, it is necessary to review the available constraints on their mode of occurrence and transport in the lithosphere.

3.3.2. Physical constraints on the movement of melts and fluids in the lithosphere

Studies of the physical behaviour of melts and fluids in refractory rocks have been published (WATSON, 1982; WATSON and BRENAN, 1987; BRENAN and WATSON, 1988). A melt phase is seen to infiltrate a rock (WATSON, 1982), whereas a H_2O–CO_2 fluid phase moves through rapidly healing cracks (BRENAN and WATSON, 1988). It is of interest to consider how this difference in behaviour could be determined by a difference in interface energies between vapour-crystal and melt-crystal interfaces.

Interface energy ratios in partially molten or fluid-impregnated rock systems can be assessed by measuring the dihedral angle (θ) at crystal-melt-crystal interfaces, from which

$$\frac{\gamma_{cc}}{\gamma_{cl}} = 2 \cos {}^1/_2\theta$$

where γ_{cc} is the crystal to crystal interface tension (interface energy per unit surface), and γ_{cl} that on the crystal-melt or crystal-fluid interface (e.g. WATSON, 1982). Values for θ are between about 30° and 50° for mafic rock-melt systems (WAFF and BULAU, 1979; HUNTER and MCKENZIE, 1989), whereas it is greater than 60° in fluid-rock systems (WATSON and BRENAN, 1987). Thus the interface tension of crystal-crystal interfaces is higher than that of crystal-melt interfaces, but lower than that of fluid-crystal interfaces.

The interfacial energetics of inhomogeneous and homogeneous melt distribution in an idealized model of a rock containing a melt or fluid phase have been studied by KRAMERS and BLENKINSOP (submitted). A result for a 1% volumetric melt or fluid content is shown in Fig. 3. Larger values of the distribution parameter X_{porous} indicate a more homogeneous distribution of the melt or fluid on grain edges in the system. It can be seen that a homogeneous melt distribution is favoured in the case of dihedral angles typical of mafic melts, whereas H_2O–CO_2 fluids are expected to become concentrated in cavities, or pockets. In such cases, where infiltration is not energetically preferred, the build-up of hydrostatic overpressure may lead to crack formation.

The Griffith fracture criterion (GRIFFITH, 1920), which is the basis of fracture mechanics, is also an energy criterion: A fracture forms, in global energetic terms, when a critical strain energy release rate G_c is exceeded. It has been shown that

$$G_c = 2\,Q_f$$

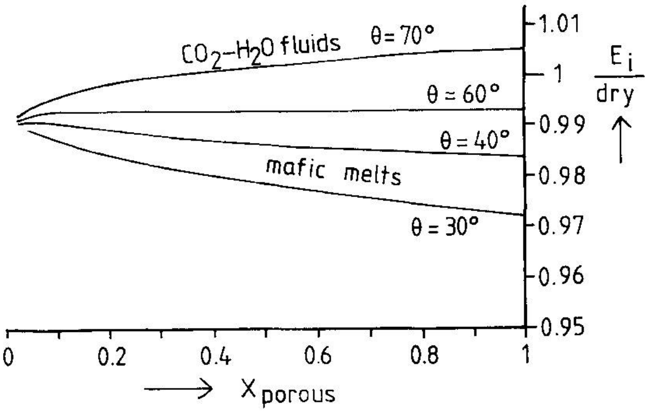

Fig. 3 The ratio of interface energy in a system with a total infiltrated melt or fluid content of 1 volume % to the interface energy of a dry system (E_i/dry) as a function of the portion of the system in which the melt or fluid is concentrated (X_{porous}), for different values of the dihedral angle θ at grain-melt-grain triple junctions. Derived from polyhedral model of KRAMERS and BLENKINSOP (submitted). Discussion in text.

(ATKINSON, 1984), where Q_f is almost equal to the tensile fracture surface energy. In other words, hydraulic fracture occurs to reduce the total energy of the fluid-rock or melt-rock system. It follows from the above that CO_2–H_2O fluids should move by crack propagation only. However, from the interface energy point of view it is also clear that a few larger cracks represent a lower total interface energy than many microcracks, and therefore fluid flow in the lithosphere is expected to be concentrated in large channels or fractures. Added to the geochemical considerations above, this is another reason why pervasive metasomatism by fluids is a highly unlikely process in the lithosphere.

The study by WATSON (1982) on melt absorption in a peridotite demonstrates the converse situation for mafic melts: Infiltration occurs even if no fluid overpressure exists. However, the process involves solution and reprecipitation of material from the rock and is therefore limited by the diffusion rate of cations in the melt. The front can therefore proceed at a rate proportional to the square root of the time elapsed since the beginning of the process:

$$x = A\sqrt{t}$$

WATSON (1982) measured a rate of approximately 2 mm/day, which corresponds to $A \approx 0.04$ if x is expressed in metres and t in years. Thus for instance, the penetration depth would be 40 m in 1 Ma, 400 m in 100 Ma and less than 3 km in the age of the Earth. This infiltration process on its own would clearly be insufficient to achieve pervasive metasomatism by melts in the subcontinental mantle, but it cannot be excluded that it could trigger off a gravity-driven upward percolating partial melt "wave" if the density difference between rock and melt was sufficient (Richter and McKenzie, 1984; Scott and Stevenson, 1986), and this type of melt percolation could in principle give rise to the trace element characteristics observed.

Pervasive infiltration of a small percentage melt over a large area of the subcontinental mantle is however only possible as long as the ambient p-T regime is at or above the solidus of the infiltrating melt and that of its modified products. Since the major element composition of any infiltrating melt will be rapidly buffered, by mineral reactions, to that of the eutectic melt of the country rock, this effectively means that the p-T conditions have to be around the solidus of subcontinental mantle peridotites themselves, for such pervasive infiltration and metasomatism to occur.

4. Conclusions: metasomatism and the thermal history of the subcontinental mantle

In the above discussion it has been argued that the pervasive "cryptic metasomatism" of the peridotites in the lithosphere underneath Archaean cratons (a) must have been caused by melts, not by CO_2–H_2O fluids, and (b) can only have occurred before these portions of the subcontinental mantle cooled down completely below their solidus. The trace element characteristics peculiar to "cryptic metasomatism" can thus not have been imposed on the subcontinental lithosphere after this became fully solid (lithospheric). The few isotope data on peridotites from the mantle underneath Archaean continents, which show this pervasive "cryptic metasomatism", indicate Archaean to early Proterozoic ages. The above conclusion is in accord with this observation.

In contrast, patent metasomatism associated with hydrous minerals in visible patchy impregnations or veins can be young, and this may be clearly understood since the present thermal regime of the subcontinental mantle lithosphere is fully subsolidus, and allows only local infiltration, limited by the ambient cooling and major element chemical buffering of infiltrating melts.

The discussion has tentatively addressed all the points raised under 3.1, except the question of why eclogite bodies within the mantle lithosphere underneath Archaean cratons appear to have escaped the metasomatic action of the postulated upward moving melts. This may be merely a result of the higher intrinsic trace element contents of the eclogites, which might render the infiltration effect less visible, or a consequence of the different chemistry or solidus relationships of eclogites at the pressures considered. While the origin of the eclogites themselves is an unsolved problem, their lack of metasomatic modification does not present a clear contradiction to the speculations of this paper.

However, the requirement that the ambient p-T conditions of the subcontinental mantle would have to be around the solidus during pervasive metasomatism is in fundamental contradiction to the characterization of this mantle region as lithospheric. Therefore, viable models for the pervasive trace element enrichment of peridotites in the mantle underlying Archaean continents will eventually have to be built into magmatic models for their origin after all, rendering these even more complex.

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