# Chemical and isotopic heterogeneities in orogenic and ophiolitic peridotites

## M. A. Menzies

SUMMARY: Rare-earth element (REE) and isotopic analysis of clinopyroxenes from orogenic and ophiolitic peridotites reveal a complex multi-stage origin resulting in chemical and isotopic heterogeneity in the upper mantle. The bulk of the pyroxenes plot within the 'mantle array' in a region characterized by mantle depleted in light REE for much of Earth's history. Interestingly the majority of the pyroxenes separated from orogenic and ophiolitic peridotites have $\varepsilon_{Nd}$ and $\varepsilon_{Sr}$ = MORB. As a consequence of such isotopic (and REE) characteristics the mantle sections of ophiolites are not believed to be related to the overlying plutonic and volcanic rocks by a simple parent–daughter melting relationship.

Ultramafic rocks found in orogenic belts are fragments of the upper mantle emplaced by obduction processes or diapirism. Trace-element and isotopic analyses of these orogenic and ophiolitic peridotites provide an invaluable insight into mantle heterogeneity and mantle evolutionary processes. In this context neodymium (Nd) isotopes are particularly useful in conjunction with rare-earth elements (REE) since these elements appear to be unaffected by serpentinization. Furthermore, the unique isotopic signature of magmas erupted at mid-ocean ridges may assist in determining the relationship between ophiolites and underlying metamorphic peridotites and the origin of orogenic and ophiolitic peridotites.

This preliminary study of diopsides in orogenic and ophiolitic peridotites was undertaken (i) to ascertain whether variations in REE and Sr-Nd isotopic composition exist in diopside separates; (ii) to shed some light on the origin of these variations, and (iii) to stimulate further research into the origin of orogenic and ophiolitic lherzolites and harzburgites. Clinopyroxenes were separated from six ophiolitic and orogenic peridotites and analysed for Sm, Nd, $^{143}Nd/^{144}Nd$ and $^{87}Sr/^{86}Sr$ isotopes by methods to be described elsewhere. These data are listed in Table 1. In the case of the Trinity peridotites from N. Carolina more complete REE data are also available (Menzies *et al.* 1977c).

## Rare-earth geochemistry

Previous investigations of orogenic and ophiolitic peridotites established their light rare-earth element (LREE)-depleted character (Haskin *et al.* 1966; Frey 1969). Moreover these authors proposed that loss of an alkali earth and LREE-enriched fluid from a mantle with an initial heavy REE content of 1.5–3.0 × chondrite produced this depletion. Since plagioclase and spinel lherzolites have a relatively undepleted major- and minor-

element content (Boudier 1972, 1978), it was further suggested that these peridotites may be a potential source for MORB-type liquids (Menzies & Murthy 1978). Some of the heterogeneity in REE is shown in Fig. 1. Clinopyroxenes from orogenic and ophiolitic peridotites exhibit a considerable range in $(Ce/Yb)_N$ ratio and the lowest $(Ce/Yb)_N$ ratio was found in pyroxenes from the Lizard peridotite (Frey 1969).

New REE analyses of clinopyroxenes from orogenic and ophiolitic peridotites display the same range in Sm/Nd as was previously reported (Fig. 1). These variations point to a mantle extremely heterogeneous for the LREE where $(La/Yb) < 1$. This characteristic depletion in LREE differs markedly from that observed in fragments of mantle disrupted by alkaline and kimberlitic magmas.

Diopsides occurring in lherzolites from Beni Bouchera, Morocco and Trinity, California are considerably richer in $\Sigma$ REE than equivalent phases in the harzburgites underlying the Troodos and Othris ophiolites. The lherzolites are known to be significantly richer in clinopyroxenes (5–15%) relative to the harzburgites (<1%). Although complete REE analyses are not available, it can be seen that harzburgites below the Othris and Troodos ophiolitic complexes have an extreme depletion in LREE somewhat akin to the Lizard Complex (Frey 1969) (Fig. 1). One can infer from the modal analyses and clinopyroxene REE analyses that the whole-rock harzburgites would contain $(Sm)_N \sim 0.009$ and $(Nd)_N \sim 0.0039$ whereas the lherzolites would have a rare-earth content of 1–2 × chondrite. The extremely low relative abundance of LREE in the Troodos and Othris harzburgites is compatible with the chemically refractory nature of these rocks (Menzies & Allen 1974) and their petrographic and mineralogic continuity over several tens of kilometres. Moreover, major- and trace-element analyses have shown the plagioclase and spinel lherzolites to be CaO- and $Al_2O_3$-rich and relatively undepleted in major- and minor-

*M. A. Menzies*

TABLE 1. *Nd and Sr isotopic composition in orogenic and ophiolitic peridotites.*

| | | $^{147}Sm/^{144}Nd$ | $^{143}Nd/^{144}Nd$ | $^{87}Sr/^{86}Sr$ |
|---|---|---|---|---|
| **Orogenic lherzolites** | | | | |
| Beni Bouchera, Morocco | CPX(lz) | 0.205 | 0.51270 (3) | 0.70515 |
| | WR(lz)* | — | 0.51275 (4) | 0.70370 |
| | WR(1z)* | 0.293 | 0.51333 (6) | 0.70216 |
| Lanzo, Italy | CPX(lz) | 0.174 | 0.51284 (3) | 0.70333 |
| | CPX(v) | 0.289 | 0.51320 (2) | 0.70277 |
| | WR(lz)* | 0.321 | 0.51310 (7) | 0.70279 |
| | CPX(lz)* | 0.253 | 0.51323 (4) | 0.70263 |
| Ronda, Spain | CPX(lz) | 0.207 | 0.51322 (5) | 0.70295 |
| | CPX(lz) | — | — | 0.70274 |
| | CPX(lz) | — | 0.51292 (2) | 0.70414 |
| Baldissero | WR(1z)* | 0.385 | 0.51359 (10) | 0.70254 |
| **Ophiolites** | | | | |
| Troodos, Cyprus | CPX(hz) | 0.456 | 0.51300 (9) | 0.70283 |
| | WR(gb)* | — | 0.51308 (4) | 0.70353 |
| | WR(gb)* | — | 0.51283 (8) | 0.70652 |
| | WR(gb)† | — | — | 0.70344 |
| Othris, Greece | CPX(hz) | 0.468 | 0.51338 (8) | 0.70371 |
| Trinity, California | CPX(lz) | 0.254 | 0.51340 (6) | 0.70238 |
| | CPX(lz) | 0.291 | 0.51388 (6) | — |

*Data taken from Richard & Allegre (1980).
†Data from Spooner *et al.* (1977).
N.B. Abbreviations: hz = harzburgite host, gb = gabbro, lz = lherzolite host, v = vein in peridotite, WR = whole rock, CPX = clinopyroxene separate.

element chemistry. This is compatible with the relative abundance of REE in these rocks.

To further understand the origin of orogenic and ophiolitic peridotites, several authors have utilized plots of $(Ce)_N$ *v.* $(Yb)_N$ (Figs. 2 & 3). Loubet *et al.* (1975) believed that the REE heterogeneity observed in orogenic lherzolites defines two genetically significant vectors on element–element plots. Melting at high pressure in the garnet stability field produces a series of residua exhibiting considerable variation in $(Ce)_N$ at almost constant $(Yb)_N$ (trend A on Fig. 3). This results from the retention of heavy REE (*viz.* Yb) by garnet during partial melting. Although the trend on Fig. 3 has been partially verified by experimental (Mysen 1976) and theoretical considerations (Frey 1983), it seems ironic that natural garnet peridotites (Garmann *et al.* 1975) do not coincide with this high-pressure residual trend (Figs 2 & 3). In fact, natural garnet-bearing peridotites plot closer to the garnet-free trend defined by Mysen (1976) or the low-pressure residual trend of Loubet *et al.* (1975). Loubet *et al.* also defined a second trend on the same figure (trend B on Fig. 3) characterized by a greater variation in $(Ce)_N$ and $(Yb)_N$ than the high-pressure residual trend. They interpreted this vector as an artifact of low-pressure melting

defined by residua generated in the spinel plagioclase stability field. Unfortunately, a further complication arises since this trend is identical to that defined by cumulus peridotites (Potts & Condie 1971; Menzies *et al.* 1977a; Whitford & Arndt 1977). Since both low-pressure anatexis and fractionation processes (Fig. 3) produce comparable variations in REE content, one cannot unequivocally state that Ce-Yb variations in metamorphic peridotites result in all cases from partial melting events. Orogenic and ophiolitic tectonite peridotite massifs are known to contain low-pressure cumulates that may have subsided from the overlying cumulates and subsequently been incorporated in the metamorphic peridotite basement (Dickey 1975). Alternatively, olivine cumulates may be plated onto conduit walls during passage of magma through the harzburgitic upper mantle (Menzies & Allen 1974). One can infer from this that in oceanic regions the upper-mantle metamorphic suite contains significant amounts of cumulus peridotite. Oceanic peridotites dredged from the Atlantic, Pacific and Indian oceans (Shih 1972) exhibit a considerable range in $(Ce)_N$ *v.* $(Yb)_N$ that approximately parallels the low-pressure fractionation-trend (Figs 2 & 3). Although a cumulus origin is an attractive hypothesis for shallow-
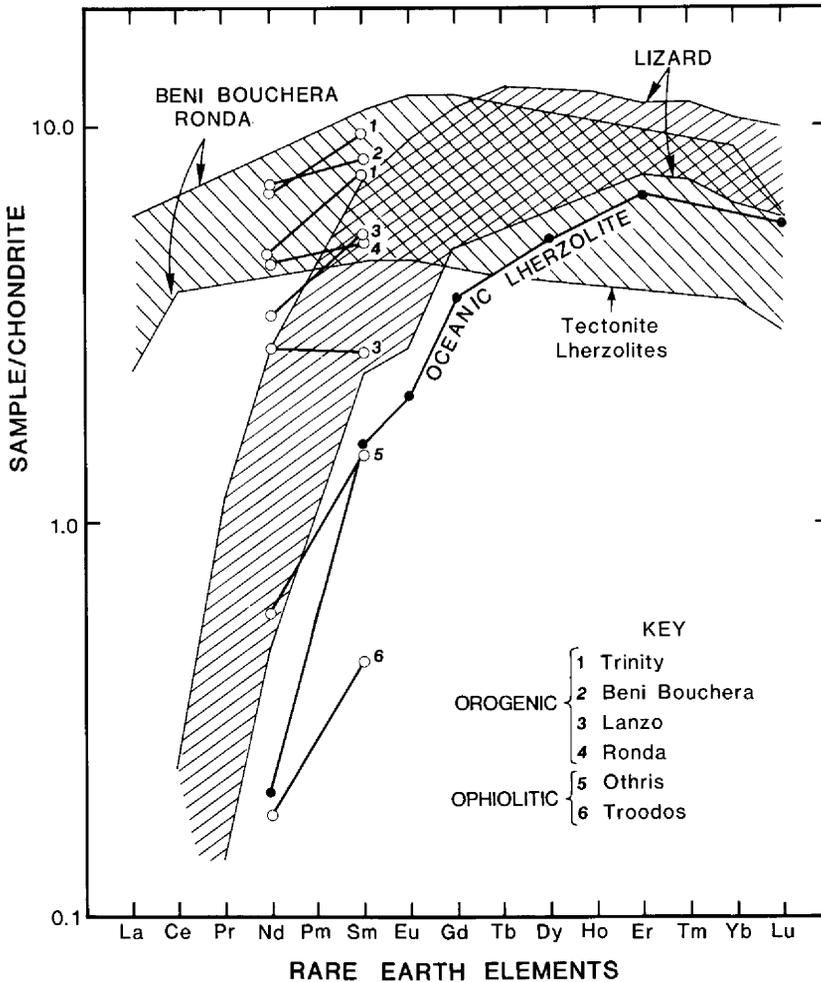
FIG. 1. Rare-earth-element abundances in clinopyroxenes separated from orogenic, ophiolitic and oceanic peridotites. Beni Bouchera, and Ronda data after Menzies (1976) and Menzies *et al* (1977b). Lizard from Frey (1969); oceanic lherzolite from Shimizu & Hart (1974). The other data are taken from Table 1.

oceanic peridotites, these rocks are variably serpentinized (Shih 1972) and consequently their origin cannot be unequivocally resolved.

Clearly, the precise origin of individual orogenic and ophiolitic peridotites is problematical and requires consideration of all aspects of their petrology and geochemistry. Such an approach has confirmed that the Beni Bouchera, Ronda and Lanzo lherzolites are residual products generated in the spinel lherzolite stability field (e.g. Loubet *et al.* 1975; Boudier 1978). Similarly the REE and isotopic data for the Trinity massif, California can be interpreted in the light of detailed petrological information. Petrographically, the Trinity lherzolites appear to have retrograded from a spinel-bearing facies to a plagioclase-bearing facies. This is believed to

indicate that the massif equilibrated in the spinel lherzolite field prior to inception of melting and intercepted the solidus at <30 km (Quick 1981, 1982). Therefore, the melt fractions preserved in these peridotites are the result of a melting event at <20 kb in a garnet-free environment. This is compatible with the clinopyroxene REE data that define a shallow trend on a $(Ce)_N$ *v.* $(Yb)_N$ plot (Figs 2 & 3). This trend is believed to have resulted from partial melting of mantle peridotites in the absence of garnet.

A somewhat more complex history may apply to the metamorphic harzburgites flooring the Troodos and Othris ophiolites. These peridotites are associated with plagioclase ± spinel-bearing lherzolites (Menzies & Allen 1974) but the marked depletion in the LREE inferred from the
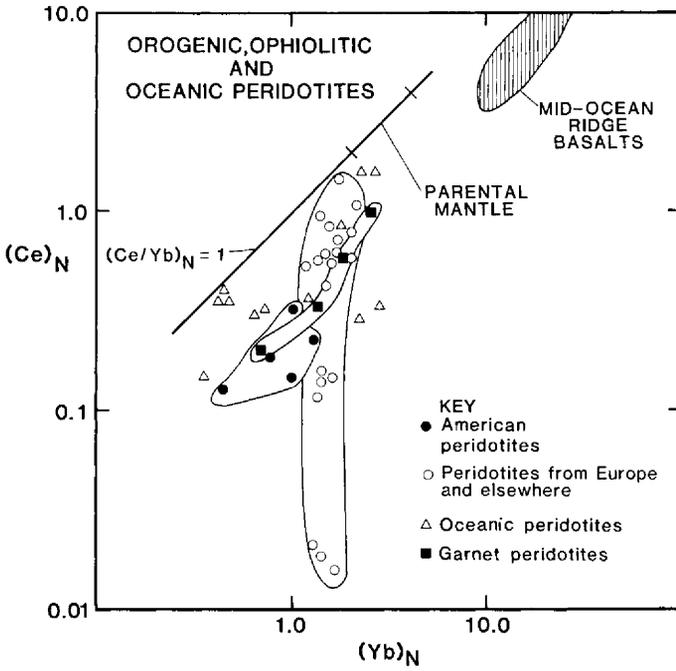
FIG. 2. $(Ce)_N$ $v$. $(Yb)_N$ variations in orogenic and oceanic peridotites. Other data taken from Shih (1972), Loubet *et al* (1975), Helmke (1983, unpubl. data) and Frey (1983).
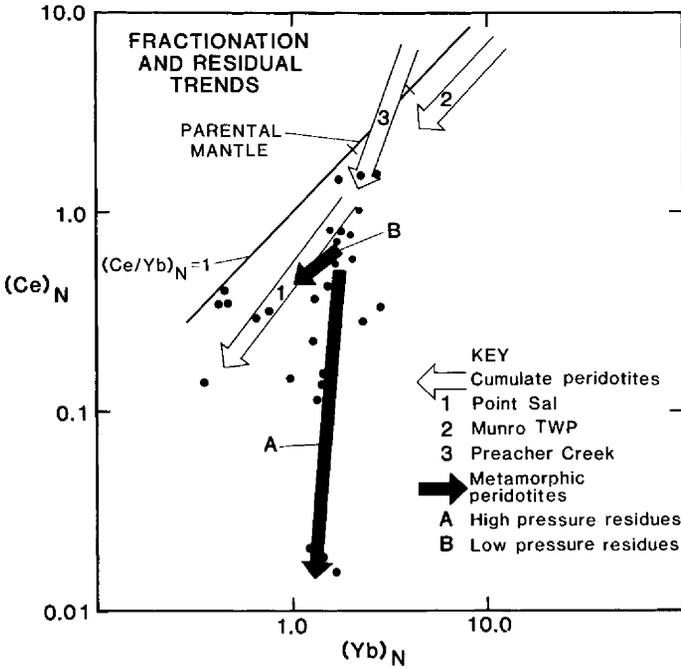
FIG. 3. $(Ce)_N$ $v$. $(Yb)_N$ variations in orogenic and oceanic peridotites (data points and references as in Fig. 2). Trend A and B represent the interpretations of Loubet *et al* (1975). Fractional crystallization trends are based on natural cumulus peridotites and pyroxenites (Potts & Condie 1971; Menzies *et al*. 1977a; Whitford & Arndt 1977). Note that natural garnet peridotites do not coincide with trend A, as would be expected, and that the low-pressure residual trend B parallels low-pressure fractionation trends.
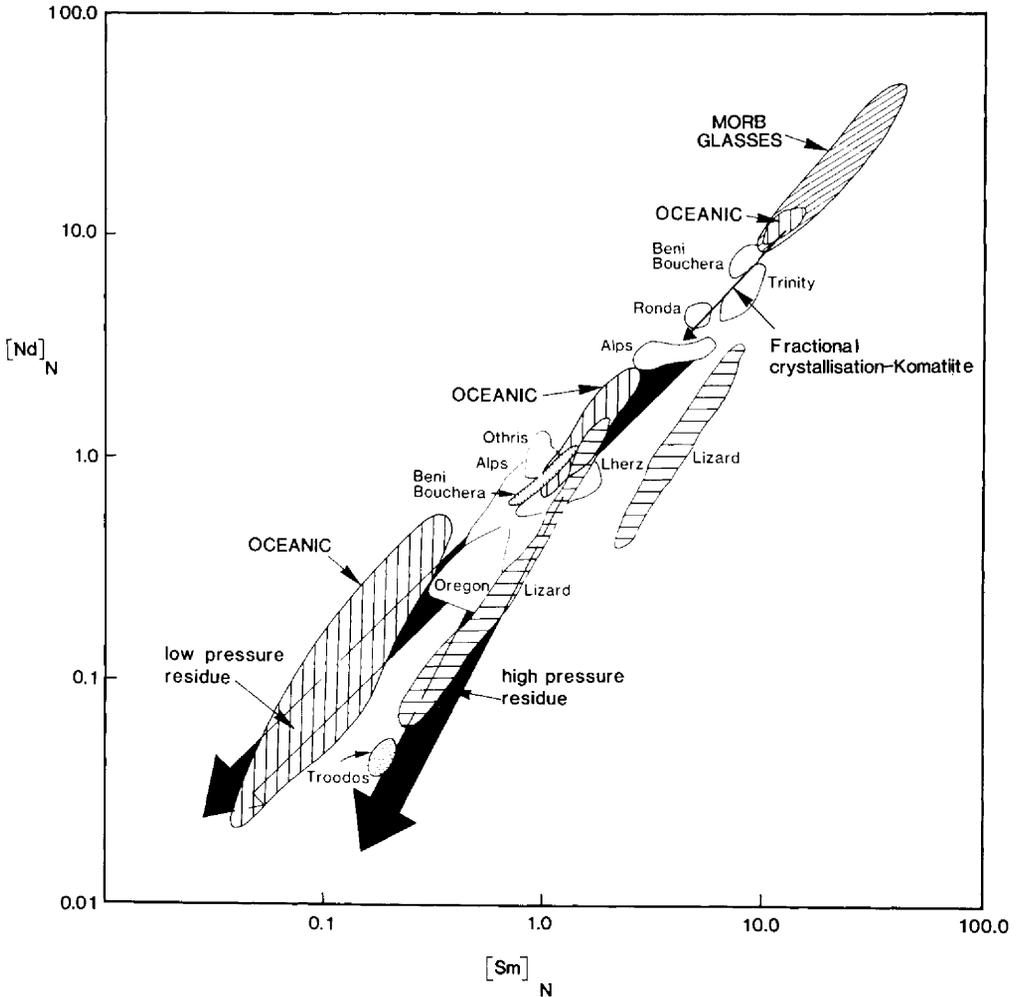
FIG. 4. $(Nd)_N$ v. $(Sm)_N$ variations in orogenic, ophiolitic and oceanic peridotites. Data sources are MORB glass (Cohen *et al.* 1980) and peridotite data (Loubet *et al.* 1975 and references therein; Ottonello *et al.* 1979). Note that the Lizard and Troodos tectonite peridotites appear to lie along a steeper trend perhaps indicative of formation in the garnet stability field.

Sm/Nd data are not compatible with melting in either the plagioclase or spinel stability field (Fig. 4). The extreme depletion in LREE relative to heavy REE (HREE) implies that the HREE were retained during melting in the garnet stability field. Frey (1969) proposed a similar origin for the Lizard metamorphic peridotites on the basis of a detailed REE study. These data have important implications since the Troodos, Othris and perhaps the Lizard metamorphic peridotites form a relatively inert basement for the accumulation of plutonic and volcanic members of an ophiolite suite. In the past it has been generally accepted that this metamorphic basement of refractory harzburgite is a residue left after removal of tholeiitic melts similar to those found in the volcanic and plutonic suite. If indeed the harzburgites formed during melting events in the garnet stability field, it is highly unlikely that the melt extracted was tholeiitic in composition. Consequently, a simple genetic relationship between mantle harzburgites and the overlying plutonic-volcanic suite seems unwarranted. This is compatible with the limited isotopic data to be presented next. Several other authors have suggested that the metamorphic peridotites underlying ophiolites and the volcanic-plutonic rocks are unrelated (Suen *et al.* 1979; Pallister & Knight 1981). For the mantle peridotite to represent a residue produced by removal of ophiolitic lavas, they must be characterized by a LREE-depleted profile. This is not the case in the

Semail, Bay of Islands and several other ophiolites where the refractory peridotites have a U-shaped profile. This LREE-enrichment suggests, albeit tentatively, that the crust and mantle suites are not co-genetic through a simple parent–daughter melting relationship (Pallister & Knight 1981).

## Sr and Nd isotopes

Orogenic peridotites have been the focus of isotopic research since the classic work of Steuber (1965). After a trace-element and isotopic study of several orogenic massifs, Stueber (1965) commented on the depleted nature of orogenic and ophiolitic peridotites and concluded that they originated in the upper mantle and represented residua of an earlier continent-forming episode. A residual origin is consistent with the already-described LREE-depleted geochemistry. More recent research (Menzies & Murthy 1978; Polve & Allegre 1980) demonstrated that a significant proportion of the Sr in orogenic harzburgites has a radiogenic composition and is therefore secondary. Detailed Sr-isotopic analyses of fresh clinopyroxenes (Menzies & Murthy 1978) confirmed the depleted nature of orogenic lherzolites and revealed a hitherto undiscovered genetic link to MORB-type liquids and ophiolites. This was later verified by Polve & Allegre (1980) in a study of several Mediterranean peridotite massifs (Fig. 5a) and by Zindler et al. (1983) after an in depth study of a layer in the Ronda massif, Spain.

In general, clinopyroxenes separated from orogenic lherzolites, pyroxenites and ophiolitic peridotites have a considerable range in $^{87}Sr/^{86}Sr \sim 0.702–0.705$ (Fig. 5a) that differs markedly from that observed in whole-rock peridotites. This confirms the addition of radiogenic Sr to the peridotites during hydrothermal alteration or serpentinization. It is interesting to note that the bulk of the clinopyroxenes (Fig. 5a) have a Sr-isotopic signature identical to MORB-type liquids (average $^{87}Sr/^{86}Sr \sim 0.7029$). Nd-isotopic analyses of orogenic and ophiolitic peridotites record ancient depletion events (Richard & Allegre 1980) and a complex multi-stage history. New Sr- and Nd-isotopic data $(T \sim 0)$ reveal local and regional isotopic heterogeneities within orogenic and ophiolitic lherzolite massifs (Table 1, Fig. 6). Furthermore, the majority of clinopyroxenes from peridotites have a $^{143}Nd/^{144}Nd$ ratio identical to MORB-type liquids. If such isotopic characteristics are to be meaningful with reference to the mantle it is important to ascertain whether or not contamination has affected the isotopic systems. Clinopy-

roxenes have a considerable range in Nd content (0.1–4.3 ppm) that does not relate systematically to the Nd-isotopic composition as would be expected had contamination occurred. This implies that the measured $^{143}Nd/^{144}Nd$ ratio reflects a primary mantle isotopic feature. The validity of using the Sr- and Nd-isotopic composition of clinopyroxenes as a valuable petrogenetic indicator has already been established (Menzies & Murthy 1978; Zindler et al. 1983). Although the Sr isotopic compositon of whole-rock peridotites may be adversely affected by post-emplacement hydrothermal alteration, it is unlikely that this will affect the Nd-isotopic composition (Cohen et al. 1980). This arises because hydrothermal waters tend to be rich in Sr relative to the LREE. Consequently, Nd isotopic analyses of whole rocks should compare favourably with the range in $^{143}Nd/^{144}Nd$ shown by constituent diopsides, whereas the Sr-isotopic composition should be somewhat different. Isotopic data from orogenic and ophiolitic peridotites verify that whole-rock peridotites have an extreme range in $^{87}Sr/^{86}Sr = 0.702–0.720$ (Fig. 5) (Stueber 1965) while clinopyroxenes have $^{87}Sr/^{86}Sr \leqslant 0.7055$. Moreover the range in Nd isotopic geochemistry is identical for both clinopyroxenes $^{143}Nd/^{144}Nd \sim 0.5138–0.5127$) and whole-rock peridotites ($^{143}Nd/^{144}Nd \sim 0.5136–0.5127$).

More specifically the following aspects of the Sr and Nd isotopic data are perhaps worth noting:

(i) Gabbroic dykelets within the Lanzo massif have been interpreted as gabbroic melts extracted from the lherzolitic wall rock (Boudier 1972, 1978). This event depletes the conduit wall rock of diopside and plagioclase. Clinopyroxenes from the wall-rock peridotites have a considerable range in $^{143}Nd/^{144}Nd \sim 0.51284–0.51323$ and $^{87}Sr/^{86}Sr \sim 0.70263–0.70333$ (Table 1). Furthermore, a clinopyroxene separated from a gabbro dykelet has a Sr and Nd-isotopic composition identical to clinopyroxene from one of these wall rocks. These data imply that isotopic equilibration between melt, now crystallized as gabbroic veins, and wall-rock peridotite is possible within the Lanzo massif. It is also worth mentioning that both the wall-rock peridotite and the melt have the isotopic characteristics of N-MORB. Clearly at Lanzo there has been interaction between MORB-type depleted mantle and a mid-ocean ridge tholeiitic melt (gabbro). The narrow isotopic variation shown by the Lanzo lherzolites is similar to that found in LREE-depleted lherzolite inclusions

a  STRONTIUM ISOTOPIC COMPOSITION
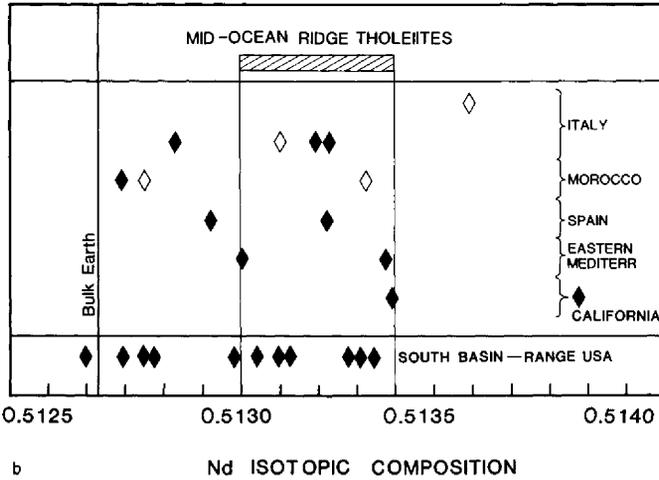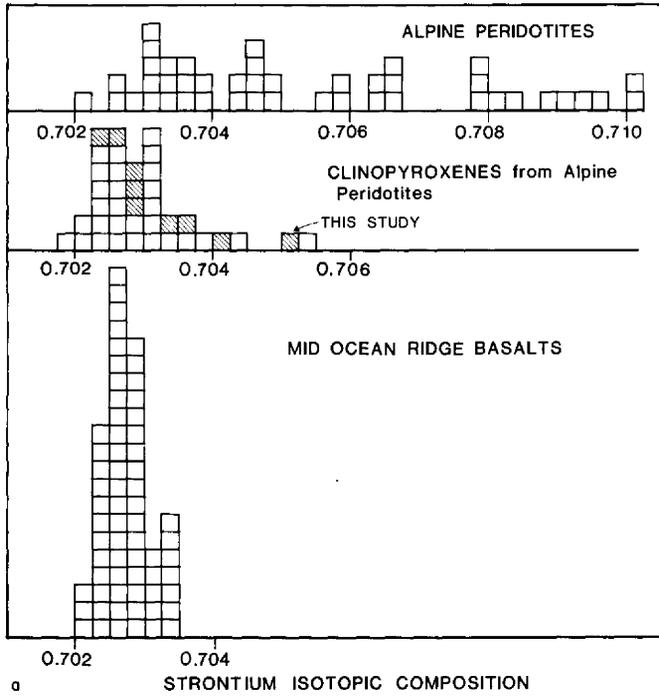


b  Nd ISOTOPIC COMPOSITION

FIG. 5. (a) Histograms of Sr isotopes in ophiolitic and orogenic lherzolites; clinopyroxenes and MORB-type liquids. (Data sources are: Polve & Allegre 1980; Stueber 1965; Richard & Allegre 1980; Menzies & Murthy 1978; Hofmann & Hart 1978.) (b) Nd isotopes in ophiolitic and orogenic lherzolites, relative to MORB-type liquids and continental lherzolite inclusions. (♦ = clinopyroxenes and ◇ = whole-rock peridotites). Data sources are: MORB (Hofmann & Hart 1978); orogenic lherzolite data. (This study and Richard & Allegre 1980, see Table 1 for exact samples); south Basin and Range Arizona–California (Menzies *et al.* 1982; Jagoutz *et al.* 1980). Note that the majority of the orogenic and ophiolitic peridotites plot *within* the range defined by mid-ocean ridge tholeiites, and that orogenic peridotites compare favourably with lherzolite xenoliths entrained in continental lavas.
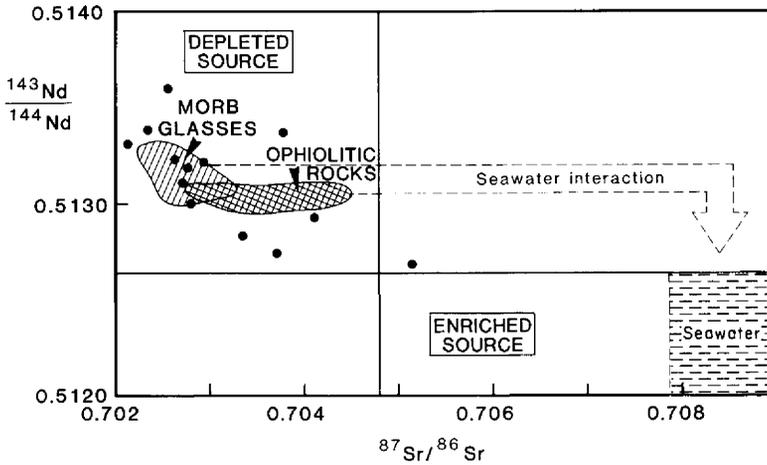
*M. A. Menzies*



FIG. 6. $^{143}Nd/^{144}Nd$ $v$. $^{87}Sr/^{86}Sr$ for orogenic and ophiolitic peridotites compared with ophiolitic volcanic and plutonic rocks (Richard & Allegre 1980) and MORB glasses (Cohen *et al.* 1980). Note the overall isotopic heterogeneity in orogenic lherzolite massifs.

brought up by alkaline basalts (Menzies *et al.* 1982). However, there appears to be a fundamental difference between orogenic-ophiolitic peridotites and xenolithic peridotites. While both groups display near identical heterogeneity in Sr and Nd isotopes the xenolithic peridotites are both LREE-enriched and -depleted. To date a marked enrichment in the LREE $(Ce/Yb)_N \geqslant 1$ has not been reported in any orogenic or ophiolitic metamorphic peridotites. Consequently, there is either a fundamental difference between mantle sampled by diapirism-obduction and volcanism, or, more likely, lherzolite inclusions are modified by the very processes transporting them to the surface.

(ii) Mantle heterogeneity is apparent in all the orogenic lherzolite massifs, in particular Beni Bouchera, Morocco (Figs 5b and 6). The considerable range in Nd- and Sr-isotopic composition can result from a complex multi-stage history occurring over several billion years. Model ages calculated for all the orogenic lherzolites vary from 0.5 to 2.4 billion years supporting a long complex evolutionary history. Quick (1981) established a record of complex mantle processes in the Trinity body, California. These included melt segregation, migration and wall rock reaction. This may in part explain the isotopic heterogeneity observed in the two Trinity lherzolites that occur within 100 m of each other (Table 1). These preliminary data indicate that isotopic heterogeneities occur on a local and regional scale. However it should be

stressed that some of this isotopic variation may be the result of $^{147}Sm$ decay after emplacement.

(iii) Chemical and isotopic heterogeneity at the level observed in orogenic and ophiolitic peridotites is also apparent in terrestrial basalts and, by inference, their source regions. Flood basalts may have retained some of the isotopic variability inherent in the mantle, prior to limited crustal contamination. Furthermore, a similar degree of Nd- and Sr- isotopic heterogeneity to that of the peridotites is also apparent in the source regions of successive episodes of volcanism erupted on Oahu, Hawaii (Stille *et al.* 1982).

(iv) A diopside separated from the tectonite harzburgite that underlies the Troodos ophiolite has a Sr-isotopic composition within the range of MORB-type liquids. However, this differs from the isotopic composition of the overlying volcanic and plutonic rocks where studies of fresh feldspar separates from the plutonic suite produces a minimum $^{87}Sr/^{86}Sr \sim 0.70344$ (Spooner *et al.* 1977). If this is an accurate representation of the minimum isotopic composition of the parental magma that formed the plutonic suite and overlying volcanic suite, then it is rather unlikely that it equilibrated with the underlying harzburgite ($^{87}Sr/^{86}Sr \sim 0.70283$) (Table 1). Consideration of both the isotopic and REE data indicates that the mantle sequence attached to ophiolite complexes is not simply related to the overlying tholeiitic basalts.

## Conclusions

The upper mantle, as represented by plagioclase and spinel-facies lherzolites, is heterogeneous both for rare earths and isotopes, on a regional and local scale. Orogenic and ophiolitic peridotites exhibit a considerable variation in Sr- and Nd-isotopic composition comparable with that observed in Type-1A xenoliths (LREE-depleted lherzolites and harzburgites) and volcanic rocks erupted in oceanic and continental environments. Furthermore, the similar Sr- and Nd-isotopic composition of most orogenic peridotites and N-type MORB liquids appears to indicate that the peridotites are genetically related to the production of MORB-type magmas. The melt segregations within the Lanzo massif are derivatives of N-type MORB that have interacted with MORB-type residual wall rock.

Mantle sampled by obduction-diapirism (ophiolites-orogenic lherzolites) and that entrained in alkaline magmas (ultramafic xenoliths) differ in their degree of heterogeneity for both the REE and isotopes. Since a fundamental difference in mantle-type is highly unlikely, the inhomogeneities evident in ultramafic xenoliths *must* in part record the very magmatic process that eventually transport them to the surface. Ultramafic xenoliths are disrupted mantle conduits whose wall rocks record a varied history of fluid and magma transport, whereas orogenic lherzolites are random mantle slabs that are less chemically and isotopically heterogeneous, and thus record a less complex history.

In conclusion, one can speculate that perhaps much of the earth's upper mantle has a Sr- and Nd-isotopic signature similar to MORB, a time-integrated response to an overall depletion in LREE. The isotopic and trace-element heterogeneity observed in mantle xenoliths is possibly a time-integrated manifestation of interaction between ancient MORB-type (depleted component) and upwelling alkaline-kimberlitic magmas (enriched component).

# References

BOUDIER, F. 1972. *Relations lherzolite-gabbro-dunite dans le massif de Lanzo (Alpes piemontaisses): Example de fusion partielle.* These presentee a l'institut des Sciences de la Nature de l'Universite de Nantes, janvier 29.

—— 1978. Structure and petrology of the Lanzo peridotite (Piedmont Alps). *Bull. Geol. Soc. Am.* **89**, 1574–91.

COHEN, R. S., EVENSEN, N., HAMILTON, P. J. & O'NIONS, R. K. 1980. U-Pb, Sm-Nd and Rb-Sr systematics of mid-ocean ridge basalt glasses. *Nature, Lond.* **283**, 149–53.

DICKEY, J. S. 1975. A hypothesis of orogin for podiform chromite deposits. *Geochim. Cosmochim. Acta*, **39**, 1061–1074.

FREY, F. A. 1969. Rare earth abundances in a high temperature peridotite intrusion. *Geochim. Cosmochim. Acta* **33**, 1429–77.

—— 1983. Rare earth abundances in upper mantle rocks. *In*: HENDERSON, P. (ed.). *Rare earth element geochemistry*. Elsevier, (in press.).

GARMANN, L. B., BRUNFELT, A. O., FINSTAD, K. G. & HEIER, K. S. 1975. Rare earth element distribution in basic and ultrabasic rocks from west Norway. *Chemical Geology* **15**, 103–16.

HASKIN, L. A., FREY, F. A., SCHMITT, R. A. & SMITH, R. H. 1966. Meteoritic, solar and terrestrial rare-earth distributions. *Phys. Chem. Earth* **7**, 169–321.

HOFMANN, A. W. & HART, S. R. 1978. An assessment of local and regional isotopic equilibrium in the mantle. *Earth Planet. Sci. Lett.* **38**, 44–62.

JAGOUTZ, E., CARLSON, R. W. & LUGMAIR, G. 1980.

Equilibrated Nd-unequilibrated Sr isotopes in mantle xenoliths. *Nature, Lond.* **286**, 708–710.

KEMPTON, P., DUNGAN, M. & MENZIES, M. A. 1982. Petrology and geochemistry of ultramafic xenoliths from the Geronimo volcanic field. *Terra Cognita* **2**, 222.

LOUBET, M., SHIMIZU, N. & ALLEGRE, C. J. 1975. Rare earth elements in alpine peridotites. *Contr. Mineral. Petrol.* **53**, 1–12.

MENZIES, M. A. 1976. Rare earth geochemistry of fused ophiolitic and alpine lherzolites. I. Othris, Lanzo and Troodos. *Geochim. Cosmochim. Acta* **40**, 645–56.

—— & ALLEN, C. A. 1974. Plagioclase lherzolite residual mantle relationships within two eastern Mediterranean ophiolites. *Contr. Mineral. Petrol.* **45**, 197–213.

—— & MURTHY, V. R. 1978. Strontium isotope geochemistry of alpine tectonite lherzolites: data compatible with a mantle origin. *Earth Planet. Sci. Lett.* **38**, 346–54.

——, BLANCHARD, D., BRANNON, J. & KOROTEV, R. 1977(a). Rare earth and trace element geochemistry of a fragment of Jurassic seafloor Point Sal, California. *Geochim. Cosmochim. Acta* **41**, 1419–30.

——, ——, —— & —— 1977(b). Rare earth Geochemistry of fused ophiolitic and alpine lherzolites. *Contr. Mineral. Petrol.* **64**, 53–74.

——, —— & JACOBS, J. 1977(c). Rare earth geochemistry of alpine tectonite lherzolites from northern California. Geol. Soc. Abs. (abs) **9**, 465.

——, KEMPTON, P. & DUNGAN, M. 1982. Nature of the

continental mantle below the Geronimo volcanic field Arizona, USA. *Terra cognita* **2**, 230.

MYSEN, B. O. 1976. Magma genesis in peridotite upper mantle in the light of experimental data on partitioning of trace elements between garnet peridotite minerals and partial melt. *An. Report Carnegie Instit. Wash.* **76**, 545–50.

—— 1978. Experimental determination of rare earth element partitioning between hydrous silicate melt, amphibole and garnet peridotite minerals at upper mantle pressures and temperatures. *Geochim. Cosmochim. Acta* **42**, 1253–63.

OTTONELLO, G., PICCARDO, G. B. & ERNST, W. G. 1979. Petrogenesis of some Ligurian peridotites. II. Rare earth element chemistry. *Geochim. Cosmochim. Acta*, **43**, 1273–84.

PALLISTER, J. A. & KNIGHT, R. J. 1981. Rare earth element geochemistry of the Samail ophiolite near Ibra, Oman. *J. Geophys. Res.* **86**, 2673–97.

POLVE, M. & ALLEGRE, C. J. 1980. Orogenic lherzolite complexes studied by $^{87}Rb$-$^{86}Sr$: A clue to understanding mantle convection processes. *Earth Planet. Sci. Lett.* **51**, 71–93.

POTTS, M. J. & CONDIE, K. C. 1971. Rare earth element distribution in a proto-stratiform ultramafic intrusion *Contrib. Mineral. Petrol.* **33**, 245.

QUICK, J. E. 1981. Petrology and petrogenesis of the Trinity peridotite, an upper mantle diapir in the eastern Klamath Mountains, northern California. *J. Geophys. Res.* **86**, 11837–63.

—— 1982. The origin and significance of large, tabular dunite bodies in the Trinity peridotite Northern California. *Contrib. Mineral. Petrol.* **78**, 413–22.

RICHARD, P. & ALLEGRE, C. J. 1980. Nd and Sr isotope study of ophiolite and orogenic lherzolite petrogenesis. *Earth Planet. Sci. Lett.* **47**, 65–74.

SHIH, C. Y. 1972. *The rare earth geochemistry of oceanic igneous rocks.* Unpubl. PhD Thesis Columbia University. August, 151 pp.

SHIMIZU, N. & HART, S. R. 1974. Rare earth element concentrations in clinopyroxene from an ocean-ridge lherzolite. *Carnegie Yearbook* **73**, 964–7.

SPOONER, E., CHAPMAN, H. & SMEWING, J. D. 1977. Strontium isotopic contamination and oxidation during ocean floor hydrothermal metamorphism of the ophiolitic rocks of the Troodos Massif, Cyprus. *Geochim. Cosmochim. Acta* **41**, 873–90.

STILLE, P., TATSUMOTO, N. & UNRUH, D. 1982. Sr, Nd, Hf and Pb isotope geochemistry on Oahu lavas. *Proceedings of the 5th Int. Conf. on Geochronology, Cosmochronology and Isotope Geology, Japan,* 354–5.

STUEBER, A. 1965. *A geochemical study of ultramafic rocks.* Unpubl. PhD Thesis, Department of Geology, University of California, San Diego, USA, 182 pp.

SUEN, C. J., FREY, F. A. & MALPAS, J. 1979. Bay of Islands ophiolite suite Newfoundland: petrologic and geochemical characteristics with emphasis on rare earth element geochemistry. *Earth Planet. Sci. Lett.* **45**, 337–48.

WHITFORD, D. & ARNDT, N. T. 1977. Rare earth abundances in a layered komatiite lava flow. *Carnegie Yearbook* **76**, 863–70.

ZINDLER, A., STAUDIGEL, H., HART, S. R., ENDRES, R. & GOLDSTEIN, S. 1983. An in-depth Nd and Sr isotopic study of a mafic layer from the Ronda ultramafic complex. *Nature Lond.* **304**, 226–30.

M. A. MENZIES, Department of Earth Sciences, Open University, Walton Hall, Milton Keynes MK7 6AA, UK.