Petrology, Mineral and Isotope Geochemistry of the External Liguride Peridotites (Northern Apennines, Italy)

Mantle peridotites of the External Liguride (EL) units (Northern Apennines) represent slices of subcontinental lithospheric mantle emplaced at the surface during early stages of rifting of the Jurassic Ligurian Piemontese basin. Petrological, ion probe and isotopic investigations have been used to unravel the nature of their mantle protolith and to constrain the timing and mechanisms of their evolution. EL peridotites are dominantly fertile spinel lherzolites partly recrystallized in the plagioclase lherzolite stability field. Clinopyroxenes stable in the spinel-facies assemblage have nearly flat REE patterns (Ce-Sm/Sm-N=0.6-0.8) at (10-16)×Cl and high Na, Sr, Ti and Zr contents. Kaersutitic-Ti-pargasitic amphiboles also occur in the spinel-facies assemblage. Their LREE-depleted REE spectra and very low Sr, Zr and Ba contents indicate that they crystallized from hydrous fluids with low concentrations of incompatible elements. Thermometric estimates on the spinel-facies parageneses yield lithospheric equilibrium temperatures in the range 1000-1100°C, in agreement with the stability of amphibole, which implies T<1100°C. Sr and Nd isotopic compositions, determined on carefully handpicked clinopyroxene separates, plot within the depleted end of the MORB field (∆Sr/Sr=0.70222-0.70263; 143Nd/144Nd=0.513047-0.51308) similar to many subcontinental orogenic spinel lherzolites from the western Mediterranean area (e.g. Ivera Zone and Lanzo N). The interpretation of the EL lherzolites as subcontinental lithospheric mantle is reinforced by the occurrence of one extremely depleted isotopic composition (∆Sr/Sr=0.70136; 143Nd/144Nd=0.513543). Sr and Nd model ages, calculated assuming both CHUR and DM mantle sources, range between 2.4 Ga and 780 Ma. In particular, the 1.2-Ga Sr age and the 780-Ma Nd age can be regarded as minimum ages of differentiation. The transition from spinel- to plagioclase-facies assemblage, accompanied by progressive deformation (from granular to tectonite-mylonite textures), indicate that the EL lherzolites experienced a later, subsolidus decompressional evolution, starting from subcontinental lithospheric levels. Sm/Nd isochrons on plagioclase-clinopyroxene pairs furnish ages of ~165 Ma. This early Jurassic subsolidus decompressional history is consistent with uplift by means of denudation in response to passive and asymmetric lithospheric extension. This is considered to be the most suitable geodynamic mechanism to account for the exposure of huge bodies of subcontinental lithospheric mantle during early stages of opening of an oceanic basin.

INTRODUCTION

Mantle peridotites from collisional belts have been frequently considered as products of Phanerozoic, pre-orogenic, mid-ocean ridge processes (involving asthenosphere upwelling, partial melting and sea-floor emplacement). However, petrological and geochemical studies have revealed that mantle rocks from Proterozoic to Phanerozoic subcontinental lithosphere are also represented within these associations (Menzies & Dupuy, 1991). Peridotites and ophiolites from the peri-Mediterranean Alpine chains are related to the formation of the Jurassic Tethyan basin. Their different lithological associations are thought to be related to two different geotectonic settings: (1) a Red Sea type of rifting and ocean formation for the Western Mediterranean suites and (2) a subduction-related origin for most of the Eastern Mediterranean suites (Beccaluva et al., 1980, and references therein).

Mantle peridotites and ophiolites of the Western
Alpine–Northern Apennine system are generally thought to have formed in the Jurassic Ligurian–Piemontese basin (JLPB), which separated the palaeo-European and Adriatic continental blocks. In the Northern Apennine orogenic belt, mantle peridotites and basaltic rocks occur in different structural positions, with respect to both the original palaeogeographic settings and the primary relationships with the associated sedimentary sequences (Decandia & Elter, 1969; Abbate et al., 1970; Pagani et al., 1972) (Fig. 1).

The Internal Liguride (IL) units are thought to have originated in an intra-oceanic setting. They contain depleted mantle peridotites, and basaltic intrusive and volcanic rocks, which represent the stratigraphic basement of the Upper Jurassic–Paleocene sedimentary sequences.

In the External Liguride (EL) units, fertile lherzolites and basalts occur mainly as huge olistoliths within the Cretaceous–Eocene flysch sequences, where they are associated with continental detritus (Hercynian felsic intrusives and volcanics, gneisses, micaschists and mafic granulites), which occurs in stratigraphic horizons directly overlying the ophiolites and belonging to the olistoliths proper (Eberhardt et al., 1962; Pagani et al., 1972).

Fertile spinel lherzolites, partly recrystallized in a plagioclase-facies assemblage, are characteristic of the EL units. Two alternative mechanisms are currently invoked for the exposure of fertile pl lherzolites: (1) active rifting, i.e. the adiabatic and diapiric upwelling of asthenospheric mantle (Nicolas, 1984, 1986); and (2) passive rifting, i.e. passive extension with the interpretation of mafic layers in orogenic subcontinental-type peridotites and pyroxenite xenoliths in alkaline basalts (Bodinier et al., 1987; Menzies & Hawkesworth, 1987; Fabriès et al., 1991).

Both peridotites and associated pyroxenites show effects of subsequent tectonic and metamorphic evolution, i.e. widespread partial recrystallization to plagioclase-bearing assemblages, and progressive deformation leading to porphyroelastic to granoblastic textures and tectonite to mylonite fabrics, occurred in kilometre-wide shear zones (Beccaluva et al., 1984; Piccardo et al., 1990; Rampone et al., 1993, and references therein). As a result, pyroxenite layers are strongly deformed and completely parallel to the peridotite foliation.

The EL peridotite–pyroxenite association was later intruded by gabbroic and late chilled basaltic dykes with MORB affinities (particularly abundant within the Mt Aiona massif) which crosscut most of the tectonite and mylonite fabrics (Rampone, 1987; Piccardo et al., 1990; Vannucci et al., 1993a).

**SAMPLES**

The investigated samples are lherzolites with significant modal amounts of clinopyroxene (up to 10–15% by volume). They were mainly collected at Suvero, Mt Nero and Mt Ragola, where the least serpentinized outcrops are found. Small-scale chemical heterogeneities in peridotites, possibly related to the pyroxenite intrusion, will not be addressed in this study. Therefore, we sampled only a few metre-scale homogeneous and massive rocks. They are spinel lherzolites showing different degrees of recrystallization under plagioclase-facies conditions and textures which range from granular to tectonite types. A
A summary of the main petrographic and textural features of the investigated samples is given in Table 1.

In the sp lherzolites (Type A samples), the development of plagioclase is only incipient and mainly confined to (1) fine-grained granoblastic aggregates of pl + ol + px, and (2) thin plagioclase rims around brown Al-rich spinels. Moreover, they contain trace amounts of kaersutitic amphibole which is typically stable in the spinel facies. In the plagioclase-bearing assemblage, these amphiboles recrystallize to a Ti-pargasitic composition. In the plagioclase lherzolites (Type B samples), which constitute the predominant rock type, the plagioclase-facies recrystallization is well developed and manifested by (1) opx + pl exsolution within sp-facies clinopyroxene, (2) well-developed plagioclase rims around Cr-rich spinels, and (3) granoblastic ol + pl + px domains. Relics of the previous spinel-facies minerals are frequently preserved as porphyroclasts within the plagioclase-bearing granoblastic matrix. Ti-pargasitic amphibole only occurs in the pl assemblage.

In the EL lherzolites, the metamorphic transition
Table 1: Petrographic and textural features of investigated samples

| Rock type                | Texture                                                                 | Sample | Locality |
|-------------------------|
| Type A                  | High strain tectonite. Highly deformed and elongated clino- and ortopyroxenes. Tabular olivines with kink-banding. Elongated spinels surrounded by thin plagioclase rims. Kaersutitic amphibole in equilibrium re-equilibration texture with the sp-facies assemblage. Narrow mylonitic bands recrystallized to pl + ol + Ti-parg + px | ER-R3/3 | Mt Ragola |
| Sp Iherzolite with      | Tabular olivines with kink-banding. Elongated spinels                                                               | ER-S2/2 | Suvero   |
| incipient plagioclase   |                                                                                                                      | ER-S2/1 | Suvero   |
| re-equilibration        |                                                                                                                      | ER-S2/4 | Suvero   |
| Type B                  | (1) Ex-protogranular-porphyroclastic. Exsolved clino- and ortopyroxene relicts. Exsolutions (opx + pl) better developed in cpx. Ameoboid dark spinel porphyroclasts with well-developed plagioclase rims. Rare triple-point equilibrium texture between opx and ol. Granoblastic domains constituted by pl + ol + px. Small grains of Ti-pargasitic amphibole in interstitial position between the porphyroclastic minerals, often showing equilibrium texture with granoblastic ol and pl | ER-N1/1 | Mt Nero   |
| Sp Iherzolite strongly  |                                                                                                                      | ER-N1/2 | Mt Nero   |
| re-equilibrated in      |                                                                                                                      | ER-N1/3 | Mt Nero   |
| pl-facies conditions    |                                                                                                                      | ER-N1/4 | Mt Nero   |
|                        |                                                                                                                      | ER-N2/14 | Mt Nero   |
|                        |                                                                                                                      | ER-N2/16 | Mt Nero   |
|                        |                                                                                                                      | ER-N2/17 | Mt Nero   |
|                        |                                                                                                                      | ER-N2/18 | Mt Nero   |
|                        |                                                                                                                      | ER-R5/1 | Mt Ragola |
|                        |                                                                                                                      | ER-R4/1 | Mt Ragola |
|                        |                                                                                                                      | ER-R4/4 | Mt Ragola |
| (2) High-strain tectonite with recrystallized mylonitic bands. Tabular olivine. Kink bands in olivine and pyroxenes. Elongated spinels with pl rims, often dismembered along the tectonite foliation. Mylonitic bands recrystallized to pl + ol + px + Ti-parg | ER-N2/7 | Mt Nero   |
|                        |                                                                                                                      | ER-N2/1 | Mt Nero   |
|                        |                                                                                                                      | ER-N2/2 | Mt Nero   |
|                        |                                                                                                                      | SP4     | Suvero    |

from spinel- to plagioclase-facies stability field is recorded by significant within-mineral redistribution of both major and trace elements. This topic has already been discussed in previous papers (Beccaluva et al., 1984) and in a recent study (Rampone et al., 1993), which documents the trace element partitioning between mantle minerals during the spinel-to plagioclase-facies reaction, based on electron and ion probe data from five selected samples. In this paper, we present a more complete data set of major and trace element data (on a larger number of samples), in addition to isotopic data. The major aim of this work is to discuss the primary compositional features of the peridotites, which are related to the complete spinel-facies equilibration, to unravel the nature of their mantle protolith.

SAMPLE PREPARATION AND ANALYTICAL PROCEDURES

Whole-rock major and trace element compositions of 20 EL peridotite samples were determined by conventional X-ray fluorescence (XRF) and inductively coupled plasma–atomic emission spectrometry (ICP-AES) techniques. Eight samples, representative of different sampling localities and of different paragenetic and textural types (see previous section), were analysed for major and trace element mineral composition and for Sr and Nd isotopic ratios.

Major element mineral analyses were performed by wavelength-dispersive spectrometry (WDS) using an ARL SEMQ electron microprobe operating at
15 kV (sample current 15 nA). Natural silicates and oxides were employed as standards. Data reduction was done according to the procedure of Bence & Albee (1968). These analyses were carried out at the Centro di Studi per la Stratigrafia e la Petrografia delle Alpi Centrali, Dipartimento di Scienze della Terra, Milano.

In situ trace element mineral analyses (mainly on clinopyroxene and amphibole) were carried out with a Cameca IMS 4f ion-microprobe at CSCC Pavia. The ‘energy filtering’ technique (Shimizu et al., 1978) was employed to remove molecular interferences. The experimental set-up and quantification are similar to those described by Bottazzi et al. (1991). A detailed description of the specific analytical procedures has been reported by Rampone et al. (1993). Trace element abundances have been normalized to the average CI chondrite composition (Anders & Ebihara, 1982).

The isotope analyses were done at the Max-Planck-Institut für Chemie in Mainz. Most of the analyses were done on clinopyroxene separates. Typical sample weights were 20-30 mg. In samples ER-N2/16 and ER-N1/3, plagioclase has also been analysed, to test isotopic equilibrium between peridotite minerals and to obtain pl-cpx isochrons. For most samples, both ‘dirty’ (direct output of the electromagnetic separator) and ‘clean’ (carefully handpicked) separates have been analysed. Both clean and dirty separates were subjected to several steps of leaching: (1) washing in 6 N double-distilled HCl in an ultrasonic bath for 20 min; (2) leaching in 6 N HC1 for 2-3 h; (3) leaching in warm diluted HNO3; and (4) repeated washing with ultrapure water. As shown below in Table 11, no appreciable differences exist between isotopic ratios in dirty and clean separates. Consequently, they can be considered good duplicates.

The leached separates were spiked with a mixed 150Nd$^{149}$Sm tracer, subsequently dissolved in HF-HNO3, and taken up in 6 N HCl. Sr was separated on a 5-ml AGV50W-X12 resin column. The separation of Sm and Nd from the REE was performed on 2-ml columns filled with HDEHP-coated Teflon powder, following the White & Patchett (1984) procedure.

Isotopic analyses were carried out by a Finnigan-MAT 261 multicollector mass spectrometer. Sr was loaded with TaF5 onto W filaments. Nd was run in its metal form on Re filaments. During the period of analyses, several measurements of NBS 987 Sr and La Jolla Nd standards gave the following values: $^{143}$Nd/$^{144}$Nd = 0.511833 ± 21 ($n = 18$); $^{87}$Sr/$^{86}$Sr = 0.710238 ± 18 ($n = 23$). Mass fractionations were corrected to 146Nd/144Nd = 0.7219 and 86Sr/88Sr = 0.1194. Total procedural blanks were of the order 0.3-0.5 ng Sr, 0.01-0.1 ng Nd, 0.04 ng Sr. The Sm-Nd isochrons were calculated using a value of $l_{Sm} = 6.54 \times 10^{-12}$/year. Model ages were calculated using: $(^{143}$Nd/$^{144}$Nd)$_{CHUR} = 0.512638$; $(^{87}$Sr/$^{86}$Sr)$_{BSE} = 0.7047$; $(^{147}$Sm/$^{144}$Nd)$_{CHUR} = 0.1967$; $(^{87}$Rb/$^{86}$Sr)$_{BSE} = 0.0868$; $^{144}$Nd/$^{144}$Nd$_{DM1} = 0.513114$; $(^{147}$Sm/$^{144}$Nd)$_{DM1} = 0.222$; $^{144}$Nd/$^{144}$Nd$_{DM2} = 0.51335$ (highest Nd isotopic ratio for present MORB; White, 1985); $(^{147}$Sm/$^{144}$Nd$_{DM2} = 0.2148$ (Hart & Zindler, 1986); $^{87}$Sr/$^{86}$Sr$_{DM} = 0.7026$ (averaged value for MORB; Ito et al., 1987); $^{87}$Rb/$^{86}$Sr$_{DM} = 0.0509$; $l_{Rb} = 1.42 \times 10^{-11}$/year (where CHUR, BSE and DM stand for chondritic uniform reservoir, bulk silicate Earth, and depleted mantle, respectively).

## WHOLE-ROCK CHEMISTRY

Major and trace element concentrations of 20 EL peridotites are listed in Table 2. In spite of widespread serpentinization, most of the analysed samples have LOI values <5%. Their fertile character (10-15 vol % modal clinopyroxene) is confirmed by the relatively high Al2O3 (2.86-4.00 wt %) and CaO (2.33-3.39 wt %) contents. Samples with the lowest MgO concentrations approach primitive mantle values (Ringwood, 1975; Jagoutz et al., 1979) (Figs 2 and 3). Such compositions are consistent with published whole-rock abundances for the EL peridotites (Ernst & Piccardo, 1979; Ottonello et al., 1984).

The volatile-free compositions exhibit some well-defined chemical trends (Figs 2 and 3), consistent with compositional ranges for mantle peridotites described in the literature (Bonatti & Michael, 1989; and references therein). Abundances of CaO, Al2O3, TiO2, V and Sc are inversely correlated with MgO contents. Similar trends have been reported for peridotites from Ronda, Horoman (see the compositional fields in Figs 2 and 3; Frey et al., 1985, 1991) and the Yugoslavian Dinaric Belt (Lugovic et al., 1991). However, the EL peridotites show a more limited range of variation of MgO abundances (38-41 wt %). These trends are generally consistent with the primitive mantle estimates of Ringwood (1975) and Jagoutz et al. (1979). In fact, the primitive mantle values plot at the low-MgO end of most correlations.

Na2O, Ni and Cr concentrations of the analysed samples are not well correlated with MgO abundance. Na2O contents can be affected by significant analytical errors, owing to the very low concentrations. Na and, to a minor extent, Cr contents are
Table 2: Whole-rock major and trace element compositions for the External Liguride (EL) peridotites

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Trace elements in p.p.m.
*Data from Rampone et al. (1991, 1993).

rather low with respect to normal abundances in peridotites with similar MgO contents. Again, this may be due to analytical errors. Moreover, the random Cr variation may reflect a non-uniform distribution of spinel, the main host phase for this element. The same possibility has been considered for the Horoman peridotites (Frey et al., 1991).

In Figs 2 and 3, the compositional field for the Internal Liguride (IL, Northern Apennine) peridotites is also reported for comparison (Ottonello et al., 1984; Rampone, 1992). Relative to the EL samples, they show distinctly lower CaO, Al$_2$O$_3$, Na$_2$O and TiO$_2$ contents. These depleted compositions are comparable with abyssal peridotites (Dick, 1989). Moreover, the Northern Apennine (EL and IL) peridotite compositional trends correspond well to the trends of the Ronda and Horoman ultramafics (Figs 2 and 3). This is especially clear for CaO, Al$_2$O$_3$ and TiO$_2$.

The element correlations defined by the EL peridotites are important in view of the recent controversy concerning the Ca/Al ratio of the upper mantle. Palme & Nickel (1985) found that the average Ca/Al ratio of many spherelherzolite mantle xenoliths is ~15% higher than in chondrites. Hart & Zindler (1986) have subsequently suggested that the non-chondritic Ca/Al ratio may be due to a sampling bias favouring cpx-rich rocks. More recently, Lugovic et al. (1991) have found a positive correlation of MgO vs Ca/Al in peridotites from the Yugoslavian Dinaric Belt. They observed, however, that samples at the low-MgO end have chondritic Ca/Al ratios. The EL peridotite data reinforce the observations of Lugovic et al. (1991) (Fig. 4). Most of the samples have lower Ca/Al ratios than the xenoliths reviewed by Palme & Nickel, and many have approximately chondritic ratios. Moreover, when the EL suite is combined with the peridotites of the Internal Ligurides (Rampone, 1992), a significant positive correlation becomes apparent, and the trend of this correlation line is consistent with a chondritic Ca/Al ratio at primitive-mantle MgO values.
### MAJOR AND TRACE ELEMENT MINERAL CHEMISTRY

Representative major and trace element compositions of EL peridotite minerals are listed in Tables 3-10. Analyses have been performed on both sp-facies (porphyroclastic) and pl-facies (granoblastic) minerals. As a whole, the major element mineral compositions presented here are consistent with previously published data (Beccaluva et al., 1984; Ernst & Piccardo, 1979).

#### Olivine

Olivines have forsteritic compositions (Fo89-91). Their restricted compositional range is independent of grain-size or occurrence as porphyroclasts or granoblasts.

#### Spinel

Spinels vary considerably in both Mg/(Mg + Fe\(^{2+}\)) (0.52-0.81) and Cr/(Cr + Al) (0.07-0.47) (Fig. 5). Rampone et al. (1993) demonstrated that this is a metamorphic compositional trend from Al-rich spinels in sp lherzolites (samples ER-R3/3 and ER-S2/2) to Cr-rich spinel relics in pl lherzolites (Type B samples). In fact, in spite of the large compositional variations recorded by the spinels, the bulk compositions of these samples are very similar. Al-rich spinels of sp lherzolite samples display compositions comparable with those of some subcontinental fertile sp lherzolites from the Alpine chain (e.g. Erro-Tobbio, Voltri Massif; Piccardo et al., 1990; Hoogerduijn Strating et al., 1993; Baldissero, Ivrea Zone; Rivalenti et al., 1979) (Fig. 5). They are, moreover, significantly different from spinel compositions of IL depleted peridotites (Beccaluva et al., 1984; Rampone, 1992), showing high Cr and low Al contents.

#### Clinopyroxene

Spinel-facies clinopyroxenes are Cr-bearing diopsides and are characterized by higher Al, Na and lower Cr contents (Fig. 6) relative to plagioclase-facies clinopyroxenes. This is particularly evident in

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| 1982    | 2278     | 2027     | 1898     | 1965    | 1756    | 1858    | 1876    | 1981     | 1848    | 1747    |
| 106     | 124      | 107      | 84       | 86      | 78      | 99      | 88      | 104      | 101     | 118     |
| 16      | 16       | 17       | 16       | 16      | 15      | 15      | 13      | 14       | 14      | 17      |
| 78      | 78       | 83       | 71       | 67      | 68      | 73      | 63      | 64       | 68      | 78      |
| 2879    | 2724     | 2659     | 2549     | 2416    | 2511    | 2540    | 2868    | 2405     | 2673    | 2360    |
| 28      | 25       | 35       | 28       | 28      | 24      | 30      | 21      | 23       | 28      | 32      |
| 52      | 48       | 47       | 56       | 50      | 47      | 52      | 48      | 48       | 48      | 47      |
| 9       | 11       | 13       | 11       | 10      | 10      | 11      | 9       | 8        | 10      | 13      |
Fig. 2. Whole-rock abundances of CaO, Al₂O₃, Na₂O and TiO₂ vs MgO (all data on anhydrous basis in wt %). Symbols in the legend inside are as follows: EL, External Liguride peridotites; A, B, C, primordial mantle estimates from Hofmann (1988), Jagoutz et al. (1979) and Ringwood (1975) (pyrolite III), respectively; D, representative oceanic abyssal peridotite composition (Dick, 1989). The fields refer to the composition of peridotites from: (1) Internal Liguride (IL) Units, Northern Apennines (data from Ramponi, 1992), (2) Ronda (data from Frey et al., 1985) and Horoman (data from Frey et al., 1991).

Fig. 3. Whole-rock abundances (p.p.m.) of Sc, V, Ni and Cr vs MgO (wt %). Symbols and fields as in Fig. 2.
Type B samples, which record systematic compositional variations (Al and Na decrease, and Cr increase) from core to rim in porphyroclasts, to granoblastic grains recrystallized in the plagioclase-facies assemblage (Rampone et al., 1993). The highest contents of Al (Al₂O₃ up to 7-8 wt%) and Na (Na₂O up to 2 wt%) are recorded by clinopyroxene porphyroclasts in ER-R3/3 (the least plagioclase-re-equilibrated sample). This feature is probably related to the lack of plagioclase exsolution in ER-R3/3 clinopyroxene porphyroclasts, which are different in this respect from all the other samples studied.

Similarly to spinels, spinel-facies clinopyroxenes have compositions comparable with those of the Erro-Tobbio and Baldirosso undepleted lherzolites (see Fig. 6). By contrast, clinopyroxenes from the IL peridotites (see compositional field in Fig. 6) exhibit significantly lower Na and higher Cr contents. Clinopyroxenes have nearly flat REE patterns, at \( \sim (10-25)\times \text{Cl} \) (Ce₁/ₘg/Yb₁/ₙ = 0.58-1.24; Sm₁/ₘg/Nd₁/ₙ = 0.96-1.25) (Fig. 7). Similar REE patterns are common in undepleted subcontinental-type spinel lherzolites from various environments such as xenoliths, orogenic massifs and pre-oceanic rifts (Frey & Prinz, 1978; Stosch, 1982; Bodinier et al., 1988; Fabries et al., 1989; Vannucci et al., 1991). The slight LREE depletion in clinopyroxenes is also consistent with published REE bulk-rock compositions of EL peridotites (Ottonello et al., 1984) (Fig. 7a). The lowest absolute REE concentrations [(10-12)\times \text{Cl}] are shown by clinopyroxene porphyroclasts in Type A samples. It is noticeable that the cores of sp-facies porphyroclasts in samples ER-R3/3 and ER-S2/2 display slightly positive Eu₁/ₙ anomalies (Fig. 7), which disappear towards the rims. By contrast, most of the clinopyroxenes in Type B samples are characterized by higher REE abundances [up to (25-30)\times \text{Cl}] in samples ER-N1/2 and ER-N1/5, and the development of slightly negative Eu₁/ₙ anomalies. Moreover, clinopyroxene porphyroclasts in sample ER-R3/3 exhibit high Sr contents (55-65 p.p.m.), in agreement with their high Na and Al concentrations. Their Zr, Y, Ti, V and Sc contents are 45-52 p.p.m., 25-30 p.p.m., 3500-3700 p.p.m., 210-220 p.p.m. and 52-62 p.p.m., respectively (Fig. 8a). These compositional ranges are typical for clinopyroxenes from undepleted sp lherzolites (O’Reilly et al., 1991; Vannucci et al., 1991). On the other hand,
### Table 4: Major element compositions of spinels

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c, core; r, rim. *Data from Rampone et al. (1993).

### Table 5: Major element compositions of clinopyroxenes

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p, porphyroclast; g, granoblast; c, core; r, rim. *Data from Rampone et al. (1993).
### Table 6: Major element compositions of orthopyroxenes

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*Data from Rampone et al. (1993).

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*Data from Rampone et al. (1993).

### Table 7: Major element compositions of plagioclases

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*Data from Rampone et al. (1993).
Table 8: Major element compositions of amphiboles

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p, porphyroclast; g, granoblast; c, core; r, rim.
*Data from Rampone et al. (1993).

Table 9: Trace element mineral compositions of Type A samples

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p, porphyroclast; g, granoblast; c, core; r, rim.
*Data from Rampone et al. (1993).
Table 10: Trace element mineral compositions of Type B samples

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Data from Rampone et al. (1993).
Orthopyroxene

Orthopyroxenes are enstatites, with Mg/(Mg+Fe) in the range of 0.892–0.907. Spinel-facies orthopyroxenes exhibit high Al2O3 (up to 6 wt%) and CaO (up to 1 wt%) contents. Both Al and Ca decrease in plagioclase-facies granoblastic orthopyroxenes.

Orthopyroxenes are characterized by very low REE absolute concentrations (<1-2) and strongly fractionated patterns (Ce/Yb = 0.02–0.07) (Fig. 7b). This is in agreement with literature data for REE compositions of orthopyroxene in mantle peridotites (Stosch & Seck, 1980; Stosch, 1982; Rampone et al., 1991). The orthopyroxenes also show very low contents of Sr (<1 p.p.m.) and Y (1.5–3 p.p.m.), whereas Zr ranges from 2 to 7 p.p.m. (Fig. 8b). These values are consistent with recently published trace element data for orthopyroxene in sp lherzolite xenoliths (O’Reilly et al., 1991). Larger variations for Ti (700–1400 p.p.m.), V (88–1400 p.p.m.) and Sc (12–25 p.p.m.) (Fig. 8b) can be related to subsolidus effects (Rampone et al., 1993).

Amphibole

According to Leake’s (1978) classification, amphiboles range from kaersutite to Ti-rich pargasite (and subordinate Mg-hastingsite) types. Kaersutitic amphiboles (TiO2 up to 5–6 wt%) occur only in sample ER-R3/3; they are part of the spinel-facies paragenesis. Amphiboles recrystallized in equilibrium with plagioclase have Ti-pargasitic compositions (TiO2 in the range 2.96–3.97 wt%). A similar compositional trend, namely a Ti decrease from sp-to pl-facies amphiboles, has also been documented for the pre-oceanic rift lherzolites of Zabargad (Red Sea) (Piccardo et al., 1988). All the analysed amphiboles have very low K contents (K2O < 0.4 wt%).

The REE spectra are similar to those of coexisting clinopyroxenes (Fig. 7b). They show slight LREE depletion (Ce/Yb = 0.5–0.6) and almost flat M- to HREE patterns at 16–22 times CI. Comparable REE abundances have been documented for kaersutite-Ti-pargasite amphiboles in the Zabargad sp lherzolites (Piccardo et al., 1993) and in sp lherzolite xenoliths from the Baikal rift zone (Ionov et al., 1992) and the Arabian side of the Red Sea (Henjes-Kunst et al., 1990). REE patterns of sp-facies amphiboles are generally similar to the respective clinopyroxene patterns. They display lower REE absolute concentrations [(10–18)×CI] and positive Eu anomalies (Fig. 7b), whereas pl-facies amphiboles have higher REE abundances [up to (30–40)×CI] and develop negative Eu anomalies (Eu/Eu* = 0.76–0.82).
EL amphiboles show, moreover, low Sr (<200 p.p.m.) and Ba (<20 p.p.m.) contents relative to amphiboles crystallized from alkaline melts (Basu, 1978; Menzies & Hawksworth, 1987). Similarly low Sr and Ba concentrations have been documented in sp-facies amphiboles from the Zabargad lherzolites (Piccardo et al., 1993) (Fig. 9).

**Plagioclase**

Plagioclase in the EL lherzolites is a subsolidus reaction product. Its compositional features and the relationships with the chemical variations in coexisting minerals have been discussed by Beccaluva et al. (1984) and Rampone et al. (1993).

Plagioclases range from An43 to An70. No correlations have been observed between An composition and textural occurrence (granoblastic grains, exsolution in cpx, rims around spinel). Their REE patterns are strongly fractionated, with significant LREE enrichment (CeN/YbN = 26–62, SmN/NdN = 0.11–0.44), well-developed positive Eu anomalies [EuN = (4.3–5.7)×Cl] and very low HREE abundances (<0.2×Cl), as usually reported in the literature (Grauch, 1989; McKay, 1989). EL plagioclases exhibit, moreover, high Sr contents...
sphere (Stosch et al., 1986; Stosch & Lugmair, 1986; Galer & O’Nions, 1989).

Tite xenoliths from Proterozoic continental lithologies are extremely depleted isotopic compositions and are similar to values of peridote xenoliths from mobile belt regions (Polve & Allegre, 1980) and in spinel lherzolite xenoliths from mobile belt orogenic peridotites (Polve & Allegre, 1980) characterized by MORB-type isotopic signatures. It must be emphasized that MORB isotopic compositions are typical for many subcontinental peridotites, both in orogenic peridotites (Polvé & Allegre, 1980) and in spinel lherzolite xenoliths from mobile belt orogenic peridotites (Polve & Allegre, 1980).

Noticeably, one sample from the Lanzo peridotite (Northern body) shows very depleted Sr isotopic composition, similar to ER-S2/2, and has been considered to represent subcontinental lithospheric mantle (Bodinier et al., 1991).

In Fig. 10, the compositional field for the IL peridotite is also reported. As for Lanzo and Baldissero peridotites, Nd isotopic ratios have been calculated assuming ages of emplacement of 270 Ma for Baldissero and 200 Ma for Lanzo (Voshage et al., 1988; Bodinier et al., 1991). The zero-age isotopic ratios, without the effect of depletion events, have been computed using $^{147}\text{Sm}/^{144}\text{Nd} = 0.222$, characteristic of MORB mantle (Faure, 1986). The calculated $^{143}\text{Nd}/^{144}\text{Nd}$ ratios, coupled with measured $^{87}\text{Sr}/^{86}\text{Sr}$ compositions, have been used to draw the compositional fields reported in Fig. 10. It is noteworthy that the EL, Lanzo and Baldissero peridotites display very similar isotopic composition and are invariably characterized by MORB-type isotopic signatures. It must be emphasized that MORB isotopic compositions are typical for many subcontinental peridotites, both in orogenic peridotites (Polvé & Allegre, 1980) and in spinel lherzolite xenoliths from mobile belt orogenic peridotites (Polve & Allegre, 1980).

Sr AND Nd ISOTOPE GEOCHEMISTRY

Sr and Nd isotopic compositions of separated clinopyroxenes are listed in Table 11. Most samples display a limited range ($^{87}\text{Sr}/^{86}\text{Sr} = 0.702216-0.702634$; $^{143}\text{Nd}/^{144}\text{Nd} = 0.513047-0.513205$), and plot within the depleted end of the MORB field (White, 1985) (Fig. 10). Sm and Nd concentrations obtained by isotopic dilution (Table 11) are consistent with the ion probe data. $^{147}\text{Sm}/^{144}\text{Nd}$ ratios range from 0.232 to 0.253, and are consistent with $^{147}\text{Sm}/^{144}\text{Nd}$ ratio (0.222-0.25) characteristic of the depleted mantle reservoir (Faure, 1986; Zindler & Hart, 1986).

Sample ER-S2/2 is peculiar, being characterized by extremely depleted isotopic composition ($^{87}\text{Sr}/^{86}\text{Sr} = 0.701736$; $^{143}\text{Nd}/^{144}\text{Nd} = 0.513543$), so that it plots on an extension of the oceanic mantle array towards residual mantle composition (Fig. 10). Such a low value is unusual for Phanerozoic mantle rocks, and bears some similarity to values of peridotite xenoliths from Proterozoic continental lithosphere (Stosch et al., 1986; Stosch & Lugmair, 1986; Menzies & Hawkesworth, 1987; Roden et al., 1988; Galer & O’Nions, 1989).

In Fig. 10, Sr and Nd isotopic compositions of EL peridotites are compared with those of orogenic peridotites from the Western Alps (e.g. Lanzo and Baldissero; data from Bodinier et al. (1991) and Obermiller et al. (unpub. data, 1994)). For the EL samples, the measured values have been plotted directly. No age correction is needed for the Sr isotopic ratios, because of the very low Rb/Sr ratios in mantle clinopyroxene. The $^{143}\text{Nd}/^{144}\text{Nd}$ ratios were also left uncorrected, because the $^{147}\text{Sm}/^{144}\text{Nd}$ ratios of analysed samples are very similar to unmodified MORB mantle. However, a decay correction was necessary for the Lanzo and Baldissero peridotites, to make the data comparable with present-day MORB mantle and with the EL lherzolites, because most of $^{147}\text{Sm}/^{144}\text{Nd}$ ratios in Lanzo and Baldissero peridotites are higher than 0.25. The initial Nd isotopic compositions have been calculated assuming ages of emplacement of 270 Ma for Baldissero and 200 Ma for Lanzo (Voshage et al., 1988; Bodinier et al., 1991). The zero-age isotopic ratios, without the effect of depletion events, have been computed using $^{147}\text{Sm}/^{144}\text{Nd} = 0.222$, characteristic of MORB mantle (Faure, 1986). The calculated $^{143}\text{Nd}/^{144}\text{Nd}$ ratios, coupled with measured $^{87}\text{Sr}/^{86}\text{Sr}$ compositions, have been used to draw the compositional fields reported in Fig. 10. It is noteworthy that the EL, Lanzo and Baldissero peridotites display very similar isotopic composition and are invariably characterized by MORB-type isotopic signatures. It must be emphasized that MORB isotopic compositions are typical for many subcontinental peridotites, both in orogenic peridotites (Polvé & Allegre, 1980) and in spinel lherzolite xenoliths from mobile belt orogenic peridotites (Polve & Allegre, 1980). This is not surprising: presumably this similarity simply reflects a long-term depletion history of the upper mantle before the accretion of the continental lithosphere.

In Fig. 10, the compositional field for the IL peridotite is also reported. As for Lanzo and Baldissero peridotites, Nd isotopic ratios have been calculated assuming $^{147}\text{Sm}/^{144}\text{Nd} = 0.222$. In spite of the age correction, the IL peridotites exhibit extremely depleted Nd isotopic compositions, different from those of the EL peridotites.
Table 11: Sr, Nd isotopic compositions and Sm, Nd concentrations for the EL peridotites

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<th>(^{143})Nd/(^{144})Nd</th>
<th>Sm (p.p.m.)</th>
<th>Nd (p.p.m.)</th>
<th>(^{147})Sm/(^{144})Nd</th>
<th>(E_{Nd})</th>
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<td>5.218</td>
<td>0.254</td>
<td>+17.6</td>
<td>+15.8</td>
<td>2.36 Ga</td>
<td>2.45 Ga</td>
<td></td>
</tr>
<tr>
<td></td>
<td>cpx c</td>
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<td>0.513544±21</td>
<td>2.348</td>
<td>5.622</td>
<td>0.253</td>
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<td>+16.5</td>
<td>1.2 Ga</td>
<td>779 Ma</td>
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<td>cpx d2</td>
<td>0.701722±22</td>
<td>0.513546±15</td>
<td>1.838</td>
<td>4.42</td>
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<tr>
<td>ER-R3/3</td>
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<tr>
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<td>0.513151±21</td>
<td>1.967</td>
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<td>0.513147±16</td>
<td>2.668</td>
<td>6.611</td>
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<td>+7.7</td>
<td>1.75 Ga</td>
<td>1.70 Ga</td>
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Sr model age: *DM model age (\(^{87}\)Sr/\(^{86}\)Sr\(_{DM}\) = 0.7026; \(^{87}\)Rb/\(^{86}\)Sr\(_{DM}\) = 0.0509). Nd model age: \(t_{DM1}\) model age (\(^{143}\)Nd/\(^{144}\)Nd\(_{DM1}\) = 0.513114; \(^{147}\)Sm/\(^{144}\)Nd\(_{DM1}\) = 0.222), \(t_{DM2}\) model age (\(^{143}\)Nd/\(^{144}\)Nd\(_{DM2}\) = 0.51335; \(^{147}\)Sm/\(^{144}\)Nd\(_{DM2}\) = 0.2148). c, 'clean' (carefully handpicked) separates; d, 'dirty' (direct output of the electromagnetic separator) separates (see the section on analytical procedures for explanation). \(d_1\) and \(d_2\) refer to different analysed aliquots in a single sample.

**P–T ESTIMATES**

The \(P\) and \(T\) nominal values for equilibration of EL peridotites under spinel- and plagioclase-facies conditions have been obtained using the composition of cpx–opx pairs pertaining to both assemblages. Temperatures have been calculated according to different geothermometers. The thermometers of Wood & Banno (1973) and Wells (1977) are based on the exchange of Ca and Mg between clinopyroxene and orthopyroxene. The thermometer of Sachtleben & Seck (1981) is based on the partitioning of Al between orthopyroxene and spinel, and is not applicable to plagioclase-bearing assemblages. Therefore, this method has been only applied to obtain the \(T\) equilibration of the spinel-bearing assemblage. Mercier's (1980) thermobarometry furnishes \(P\) and \(T\) estimates from the composition of a single-pyroxene phase (in equilibrium with a second pyroxene and an Al-rich mineral).

Application of the above methods yields mean equilibration temperatures which, for each assemblage, fall within a range of 100°C. The thermometers of Wood & Banno (1973) and Wells (1977) mark, respectively, the upper and lower bounds of this \(T\) range.

The spinel-facies assemblage, based on compositions of homogeneous to slightly exsolved porphyroclast cores, yields mean temperatures between 1000 and 1050°C, at ~15–20 kbar confining pressure. The plagioclase-facies re-equilibration, which has been determined using the compositions of recrystallized grains, yields lower-temperature conditions, mainly within the range 900–950°C, and lower pressures (8–10 kbar). The temperature estimates are consistent with equilibration conditions obtained by Beccaluva et al. (1984).

In spite of the controversial validity of barometers for spinel lherzolites, pressure estimates obtained by
DISCUSSION

Amphibole metasomatism

Disseminated Ti-pargasite is ubiquitous in all the subcontinental-mantle derived materials (Fabriès et al., 1991), and is generally considered to indicate metasomatic introduction of a volatile-rich melt or fluid phase (Menzies & Hawkesworth, 1987; and references quoted therein) also enriched in incompatible elements. This evidence is based on trace element data of amphiboles crystallized in mantle peridotites (mainly documented in xenoliths), which are predominantly characterized by strong enrichment in incompatible elements. This amphibole metasomatism is commonly thought to modify the isotopic composition (especially the Sr isotope ratio) of the bulk peridotites as well as the clinopyroxenes.

Amphiboles stable in the spinel-facies assemblage of the EL peridotites have low contents of Sr and Ba (Fig. 9), and their REE patterns are LREE depleted, similar to the coexisting clinopyroxene (Fig. 7b). Also, the clinopyroxenes in amphibole-bearing peridotites (see Table 11) show MORB-type isotopic signatures, consistent with the other analysed samples. This indicates, in contrast to the above generalization, that amphibole metasomatism did not affect the clinopyroxene chemistry.

Amphibole formation without concomitant introduction of LREE and Sr has also been documented for spinel lherzolites from the Red Sea (Zabargad peridotites; Piccardo et al., 1993; Western Arabian xenoliths; Henjes-Kunst et al., 1990) and spinel lherzolite xenoliths from the Bartoy volcanoes of Siberia (Ionov et al., 1992). According to these works, the above features suggest that the metasomatic agent from which amphibole crystallized was not enriched in incompatible elements, thus excluding alkaline melts or related high-density fluids.

The interpretation proposed for the Zabargad lherzolites (Piccardo et al., 1993) considers that the composition of H$_2$O-rich fluids in equilibrium with the upper mantle could have been originally depleted in LREE and Sr. If an originally enriched metasomatic agent is assumed, we suggest that the explanation for this apparently paradoxical behaviour lies in the chromatographic nature of infiltration metasomatism (Korzhuskii, 1968; Hofmann, 1972). This causes the metasomatic fronts of the solutes to lag behind the fluid front itself. This lag is zero only if the bulk partition coefficient between solid and fluid is zero, and it increases with increasing partition coefficients. The delay of the cation enrichment fronts is also proportional to the distance the fluid has travelled from its source. Therefore the effect is expected to be significant only if the metasomatic fluid has migrated for some distance through the mantle before precipitating the amphiboles found in the EL peridotites.

In this context, it is worth noting that the stability of amphibole in mantle peridotites implies temperatures <1100°C. If the fluid had traversed hotter mantle before encountering the peridotites sampled, it would not have precipitated amphibole, and the chromatographic effects described above would have had 'time and space' to develop. The 1100°C limit is...
The EL peridotites are characterized by rather fer-

The nature of mantle protolith

The EL peridotites are characterized by rather fer-

Also consistent with temperature estimates obtained

Also consistent with the nature of their mantle protolith, is the occurrence

Another remarkable geochemical feature of the

Also consistent with temperature estimates obtained

Also consistent with temperature estimates obtained
Eu/Eu* = 1.18—1.20). A simple explanation for this feature would be to assume an equilibration with a mineral phase able to compensate this Eu enrichment by a corresponding depletion. However, the analysed spinel-facies orthopyroxene does not show negative Eu anomalies (Fig. 7b), and olivine and spinel have REE contents too low to be considered in the mass balance. Moreover, it is evident that the positive Eu anomaly tends to disappear towards the rims of porphyroclasts and in plagioclase-facies granoblastic clinopyroxenes which, instead, show the progressive development of negative Eu anomalies, owing to the equilibrium crystallization with plagioclase (Rampone et al., 1993). This means that the Eu enrichment in spinel-facies clinopyroxene cannot be related to diffusive redistribution from plagioclase, as proposed by Henderson et al. (1976) and Seifert & Chadima (1989). The plagioclase-facies re-equilibration, in fact, causes the opposite effect. Because there is no obvious mineral present in the rock (stable in the sp assemblage) which might compensate the positive Eu anomaly of the clinopyroxene with a complementary and sufficiently large negative anomaly, it is likely that the whole rocks also have positive anomalies (although we cannot be absolutely certain of this because we do not have the appropriate whole-rock analyses to demonstrate this conclusively). How such positive anomalies might originate in the mantle is not at all clear. Compensating negative Eu anomalies are sometimes found in garnets, for example in garnet lherzolites from the Alpe Arami and Monte Duria, central Alps (unpublished ion probe data). However, to generate a positive whole-rock anomaly, the garnet would have to be removed from the rock by some sort of disequilibrium melting process. Another way to create positive whole-rock Eu anomalies in mantle rocks might be equilibration with subducted crustal materials containing cumulus plagioclase, which would become unstable at mantle pressures but leave behind Eu anomalies as their geochemical fingerprint. Such speculations gain some credibility by the observations from Hawaiian basalts, some of which have significant positive Eu anomalies, which have apparently been inherited from their mantle sources (e.g. Feigenson et al., 1983; West et al., 1992).

Alternatively, if we assume that no positive Eu anomaly exists in the bulk rock, we could speculate that the EL peridotites were formerly equilibrated in the garnet-facies stability field. During the sp-facies re-equilibration, most of clinopyroxenes preserved the Eu positive anomaly (recording their previous equilibration with garnet), whereas some clinopyroxenes (not analysed) crystallized as product of the garnet breakdown and inherited a compensating negative Eu anomaly. This garnet-facies equilibration could be related to the early (Proterozoic) upwelling of the EL peridotites from deeper asthenospheric mantle and consequent accretion to the lithosphere. More detailed ion probe investigations would be necessary to substantiate this hypothesis. We note, however, that a similar process, implying local geochemical disequilibrium during subsolidus partitioning, has been involved to explain the occurrence of unusual (garnet-like) trace element abundances in orthopyroxene and clinopyroxene from the Zabargad Al-Di pyroxenites (Vannucci et al., 1993b).

The decompressional evolution during Triassic–Jurassic extension of the Ligurian Tethys

The EL peridotites record a composite history of progressive deformation (from granular to tectonite fabrics) and recrystallization from spinel- to plagioclase-facies stability field. This subsolidus evolution is testified by major and trace element redistribution between mineral phases, such as (1) Al, Na, Sr and
Eu/Eu* decrease, and Y, V, Sc, Cr, Zr and Ti increase in clinopyroxene, and (2) Al decrease, and Cr and Ti increase in spinel. Mass balance calculations (Rampone et al., 1993) demonstrate that the element redistribution occurred in response to the metamorphic reaction governing the transition from the spinel- to the plagioclase-facies assemblages. Thermometric estimates indicate that this decompressional evolution occurred under slightly decreasing temperature conditions (from 1000–1100°C to 900°C).

Plagioclase and clinopyroxene of two EL samples which were significantly recrystallized in plagioclase-facies conditions have been analysed for Sr and Nd isotopic compositions and Sm and Nd concentrations, to determine the age of this subsolidus pl-facies re-equilibration.

In both samples, clinopyroxene and plagioclase show the same $^{87}\text{Sr}/^{86}\text{Sr}$ ratio (Table 11), thus indicating Sr isotopic equilibrium. By contrast, they are characterized by different Nd isotopic compositions, in response to their different $^{143}\text{Sm}/^{144}\text{Nd}$ ratios (plagioclase: 0.095–0.098; clinopyroxene: 0.238–0.243). This is shown in Fig. 11b, where the two plagioclase–clinopyroxene pairs define essentially parallel isochrons giving ages of 163 and 165 Ma ($\pm$ 20 Ma).

Plagioclase–clinopyroxene isochrons are not used frequently because the Sm/ Nd ratios of these two minerals are not very different, and the resulting uncertainty on the calculated ages is comparatively large. Nevertheless, the good agreement between the two obtained ages and the relatively small analytical errors lend credibility to these isochrons and provide important information about the timing of EL peridotite subsolidus evolution. In particular, the 163–165 Ma cooling age indicates that this event occurred slightly before the oceanic opening stage (Bathonian–Callovian) of the JLPB (Hoogerduijn Strating, 1991).

Geodynamic implications
The predominance of rather fertile lherzolites is a peculiar feature of the upper-mantle rocks early exposed on the sea-floor within the Jurassic Ligurian Piemontese basin (JLPB) (Bezzi & Piccardo, 1971; Nicolas & Jackson, 1972). These fertile lherzolites have been interpreted as (1) asthenospheric mantle which upwelled adiabatically and intruded into the continental lithosphere to produce the newly formed Jurassic oceanic sequences (Nicolas, 1984, 1986), or (2) lithospheric subcontinental mantle which was exposed on the sea-floor during the ocean opening by passive extensional mechanisms (Decandia & Elter, 1969; Lombardo & Pognante, 1982; Lemone, 1983), following a subsolidus decompressional trajectory (Beccaluva et al., 1984). These two different interpretations have been applied to chemically and texturally similar lherzolites from different sectors of the Alpine chain.

The results of our petrologic and isotopic investigations indicate that the EL peridotites represent pre-Triassic subcontinental lithospheric mantle, presumably accreted to the lithosphere during Proterozoic time. Similar compositional features (i.e. a rather fertile character coupled with MORB isotopic signature) have been documented for other ophiolitic peridotites from the Alpine chain (e.g. Lanzo massif; Bodinier et al., 1991), as well as for peridotite bodies still linked to the Adria continental basement (e.g. Baldissero; Obermiller et al., unpub. data, 1994). It seems plausible, therefore, to interpret the above ophiolitic peridotites as representative of the pre-oceanic subcontinental mantle of this part of the Western Tethys. The same interpretation has been inferred for the Lanzo peridotite body, at least concerning its northern part (Bodinier et al., 1991).

Our petrologic and isotopic studies indicate, moreover, that the EL subcontinental mantle underwent an uplift history marked by a subsolidus evolution with slightly decreasing temperatures under decompression. This stage probably developed during the pre-oceanic rifting of the JLPB, in agreement with the 165-Ma cooling age obtained for the re-equilibration under plagioclase-facies conditions.

The above evolutionary features, and in particular the Proterozoic model age of one sample, are consistent with a passive uplift of the subcontinental lithosphere, rather than with an active and adiabatic upwelling of the asthenosphere. Thermal modelling (Ruppel et al., 1988) for rocks exhumed via passive (pure and/or simple shear) mechanisms suggests that, in the case of simple (i.e. asymmetric) shear, rocks follow almost linear $P-T$ paths with progressively decreasing temperature and pressure when they evolve as the footwall of a low-angle normal fault.

A consistent petrological and structural evolution has been documented for the ophiolitic peridotites of the Erro–Tobbio body (Ligurian Alps; Piccardo et al., 1990; Vissers et al., 1991; Hoogerduijn Strating et al., 1993). This subsolidus evolution has been related to the rifting stage in the JLPB, leading to tectonic unroofing of the lithospheric upper mantle by means of passive and asymmetric extension of the lithosphere (Wernicke, 1985). This geodynamic process is regarded as the most plausible for the pre-oceanic rift in the JLPB (Lemoine et al., 1987; Dal Piaz, 1987).
1993), and is thought to be responsible for the exposure of the lithospheric subcontinental mantle during the inception of the oceanic basin (Piccardo et al., 1992).

CONCLUSIONS

The EL peridotite evolution

The External Liguride peridotites represent a slice of subcontinental lithospheric mantle which was exposed on the sea-floor during early stages of oceanic opening of the Ligurian Piemontese basin. Chemical and isotopic compositions highlight their similarity to relatively fertile MORB-type mantle sources. The MORB isotopic signature is documented for various orogenic spinel lherzolites of subcontinental lithospheric provenance, at present outcropping in the Alpine—Apennine chain (Richard & Allegre, 1980; Bodinier et al., 1991; Obermiller et al., unpub. data, 1994). This could indicate that part of the subcontinental lithospheric mantle of the Western Tethys was derived through early (Proterozoic) accretion of MORB-type asthenospheric mantle.

The history of the External Liguride peridotites can be summarized as follows:

1. ancient (presumably Proterozoic) upwelling and depletion of an asthenospheric MORB-type mantle;
2. lithospheric accretion and annealing recrystallization (at $T<1100°C$) in the spinel-facies stability field;
3. modal metasomatism (crystallization of kersantitic—Ti-pargasitic amphibole) by means of hydrous fluids not enriched in incompatible elements. A chromatographic mechanism is considered to account for the low concentration of LREE, Sr and Ba in the reacting fluid;
4. Triassic—Jurassic decompressional evolution recorded by progressive deformation (from granular to tectonite—mylonite types) and metamorphic reequilibration from spinel- to plagioclase-facies stability field (plagioclase—clinopyroxene cooling ages of $~165$ Ma);
5. shallow emplacement, recorded by the intrusion of MORB-type chilled basaltic dykes and low-grade alteration processes (peridotite serpentinitization and basic dyke rodingitization).

Early Jurassic decompressional evolution recorded by the EL peridotites is consistent with uplift by means of denudation in response to asymmetric lithospheric extension. This is considered to be the most likely geodynamic mechanism to account for the exposure of huge bodies of subcontinental lithospheric mantle during early stages of opening of an oceanic basin.

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