ELSEVIER
Earth and Planetary Science Letters 180 (2000) 259–270
EPSL
www.elsevier.com/locate/epsl

The role of lower continental crust and lithospheric mantle in the genesis of Plio–Pleistocene volcanic rocks from Sardinia (Italy)

Michele Lustrino *, Leone Melluso, Vincenzo Morra

Dipartimento di Scienze della Terra, Università degli Studi di Napoli 'Federico II', Via Mezzocannone 8, I-80134 Naples, Italy

Received 25 October 1999; accepted 6 June 2000

Abstract

The first comprehensive chemical and Sr–Nd–Pb isotopic data set of Plio–Pleistocene tholeiitic and alkaline volcanic rocks cropping out in Sardinia (Italy) is presented here. These rocks are alkali basalts, hawaiites, basanites, tholeiitic basalts and basaltic andesites, and were divided into two groups with distinct isotopic compositions. The vast majority of lavas have relatively high $^{87}Sr/^{86}Sr$ (0.7043–0.7051), low $^{143}Nd/^{144}Nd$ (0.5124–0.5126), and are characterised by the least radiogenic Pb isotopic composition so far recorded in Italian (and European) Neogene-to-Recent mafic volcanic rocks ($^{206}Pb/^{204}Pb = 17.55–18.01$) (unradiogenic Pb volcanic rocks, UPV); these rocks crop out in central and northern Sardinia. Lavas of more limited areal extent have chemical and Sr–Nd–Pb isotopic ratios indicative of a markedly different source ($^{87}Sr/^{86}Sr = 0.7031–0.7040$; $^{143}Nd/^{144}Nd = 0.5127–0.5129$; $^{206}Pb/^{204}Pb = 18.8–19.4$) (radiogenic Pb volcanic rocks, RPV), and crop out only in the southern part of the island. The isotopic ratios of these latter rocks match the values found in the roughly coeval anorogenic (i.e. not related to recent subduction events in space and time) mafic volcanic rocks of Italy (i.e. Mt. Etna, Hyblean Mts., Pantelleria, Linosa), and Cenozoic European volcanic rocks. The mafic rocks of the two Sardinian rock groups also show distinct trace element contents and ratios (e.g. $Ba/Nb > 14$, $Ce/Pb = 8–25$ and $Nb/U = 29–38$ for the UPV; $Ba/Nb < 9$, $Ce/Pb = 24–28$ and $Nb/U = 46–54$ for the RPV). The sources of the UPV could have been stabilised in the Precambrian after low amounts of lower crustal input (about 3%), or later, during the Hercynian Orogeny, after input of Precambrian lower crust in the source region, whereas the sources of the RPV could be related to processes that occurred in the late Palaeozoic–early Mesozoic, possibly via recycling of proto-Tethys oceanic lithosphere by subduction.

Keywords: Sr-87/Sr-86; Nd-144/Nd-143; Pb-206/Pb-204; lithosphere; lower crust; Quaternary; volcanism; Sardinia Italy

1. Introduction

The Cenozoic European Volcanic Province is characterised by rocks with relatively uniform chemical and isotopic composition, when compared to the complex geodynamic evolution of the area and to the heterogeneity of the European

* Corresponding author. Present address: Dipartimento di Scienze della Terra, Università degli Studi di Roma 'La Sapienza', P.le Aldo Moro 5, I-00185 Rome, Italy; E-mail: lustrino@uniroma1.it

0012-821X/00/$ – see front matter
PII: S0012-821X(00)00185-0

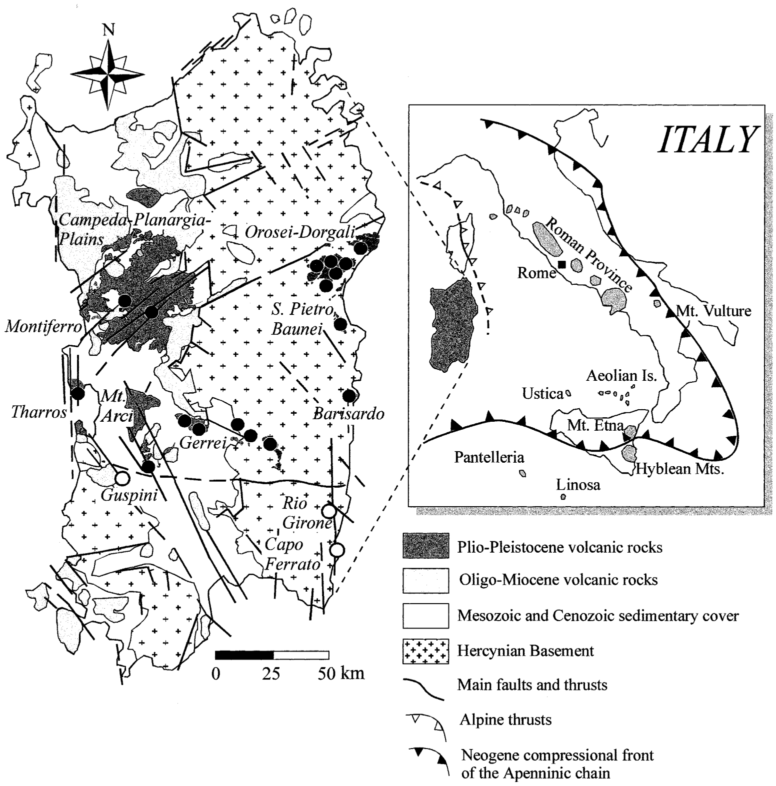

Fig. 1. Geological sketch map of Sardinia. Filled circles: UPV; open circles: RPV. The inset shows the other outcrops of Italian volcanic rocks.

lithospheric mantle, as indicated by the xenoliths hosted in the mafic alkaline volcanic rocks [1,2]. The geochemical characteristics of the Cenozoic European Volcanic rocks suggested a homogeneous asthenospheric source (i.e. the European Asthenospheric Reservoir [1,3]), only slightly modified by lithospheric or subduction-related melts. The evolution of the European mantle is much more complex in the Mediterranean, as this area suffered several compressional and extensional cycles, at least from the Panafrican Orogeny (900–450 Ma [4]).

Modifications of the European subcontinental mantle related to the Alpine orogeny can be easily detected in several regions of Europe (e.g. the 'orogenic' character of several Cenozoic volcanic provinces from Spain to Romania). These modifications are ultimately related to subduction of oceanic crust (Tethys ocean) below Europe. On the other hand, the mantle sources of the Cenozoic European volcanic rocks do not seem to record, up to now, even older modifications such as those related to the Panafrican and the Hercynian orogenies. During these events (late Proterozoic and late Palaeozoic, respectively), complex interaction among various mantle domains and crustal lithologies took place. This paper aims to search for evidence of such old modifications by studying the large, but relatively poorly known, European Cenozoic volcanic province of Sardinia (Italy). Due to the presence of plutonic and volcanic rocks dating back at least to early Palaeozoic, and to its geographic position, the island of Sardinia is an important source of information pertaining to the genesis and geodynamic implications of the European magmatism. The peculiarity of these rocks in terms of both trace elements and Sr–Nd–Pb isotopic ratios is pointed out, and a model to constrain the mantle sources of this volcanism is attempted.

2. Geological background

The crystalline basement of Sardinia is made up of late Precambrian to Palaeozoic rocks, mostly metamorphosed during the Hercynian Orogeny (Carboniferous–early Permian). During this time, subduction of the proto-Tethys Ocean between the Gondwanan and Laurussian plates [5] was followed by a continent–continent collision, and by intrusion of calc-alkaline plutonic rocks [6]. The tectonic collapse of the Hercynian belt at the end of the Palaeozoic resulted in the fragmentation of the just accreted Pangaea, and evolved to the late Triassic formation of the North Atlantic–Tethys Ocean rift system [7]. A new compressive cycle began after the Jurassic–Cretaceous opening of the South Atlantic Ocean. The birth of this ocean caused an east-northeastward motion of Africa, and closure of the Western Tethys Ocean (Ligurian–Piedmontese Ocean), with the formation of the Alpine–Betic front [8]. During the Oligo–Miocene, a new rift system, extending from the shore of the North Sea to the Ligurian–Provençal basin, developed across Europe [7]. The development of the Ligurian–Provençal and Alboran basins (30–18 Ma) is related to the west-northwest-directed subduction of Tethyan lithosphere, which induced counterclockwise rotation and back-arc basin formation west of the Sardinia–Corsica microplate, in former continuity with the French basement [8]. Arc-tholeiitic to high-K calc-alkaline volcanics (32–15 Ma) were erupted in response to this subduction system [9–11]. About 7–10 Ma after the end of this 'orogenic' magmatic activity, a later volcanic phase took place in the island during Plio–Pleistocene, in relation to stresses induced by the eastward opening of the Tyrrhenian Sea. The formation of this basin (17–0 Ma) is linked to the same subduction regime responsible for the opening of the Ligurian–Provençal back-arc basin [12]. This subduction system also induced a coeval eastward extensional (Tyrrhenian Sea opening) and compressive waves (Apennine orogenic belt formation) in the Mediterranean area after the middle Miocene. The volcanic products of this tectonic phase are the Plio–Quaternary Italian orogenic rocks (on the eastern side of the Tyrrhenian Sea; Fig. 1), those forming the Tyrrhenian abyssal plain (made up mainly by enriched MORB and calc-alkaline basalts), 'transitional' products (Ustica island and Mt. Vulture) and sparse anorogenic volcanic rocks on the southern (Mt. Etna, Hyblean Mts., Pantelleria, Linosa) and western sides (Sardinia) of the Tyrrhenian Sea.

The age of the most recent volcanism in Sardinia, pertinent to this study, ranges from ~5.3–5.0 Ma (Capo Ferrato, southeast Sardinia) to 0.9–0.1 Ma (Logudoro, northern Sardinia [13]) and is roughly contemporaneous with the orogenic magmatic activity in central and southeastern Tyrrhenian area. The Plio–Pleistocene Sardinian Volcanic rocks (hereafter PSV) crop out in the form of large volcanic complexes (Montiferro and Mt. Arci), basaltic plateaux (Gerrei, Orosei–Dorgali, Campeda–Planargia–Abbasanta–Paulilatino Plains), lava flows from small or monogenetic volcanoes (S. Pietro Baunei, Barisardo, Tharros, Capo Ferrato) and necks (Guspini and Rio Girone; Fig. 1). The rocks range from mafic to intermediate types, and belong to tholeiitic, transitional and alkaline series. The PSV are hawaiite, mugearite, tholeiitic basaltic andesite and alkali basalt (in order of abundance), with rare basanite and evolved rocks (rhyolite, trachyte and phonolite). The affinity of the alkaline rocks ranges from sodic to slightly potassic (e.g. [14]). These rocks are erupted above a normally thick crust (~30 km) and an inferred lithospheric thickness of about 90 km [15].

3. Results

A subset of 10 samples out of a collection of about 200 [14,16] was chosen for Sr, Nd and Pb isotopic systematics, performed at S.O.E.S.T., University of Hawaii at Manoa (Table 1), in order to represent the isotopic composition of tholeiitic, alkaline and strongly alkaline volcanic rocks cropping out over the island (Fig. 1), with particular interest to mafic volcanics. The occurrence of ultramafic xenoliths in the mafic alkaline samples [17] indicates a rapid rise of the host magma during ascent, and suggests that magmas were not significantly contaminated by crust. The

Table 1
XRF and ICP–MS whole rock chemical analyses, $^{87}Sr/^{86}Sr$, $^{143}Nd/^{144}Nd$ and Pb isotopic ratios for the Plio–Pleistocene volcanic rocks of Sardinia

Type:	UPV	UPV	UPV	UPV	UPV	UPV	UPV	RPV	RPV	RPV
Sample:	MGV245	MGV51	MGL23	MGV223	MGV76	MGV236	MGV238	MGV249	ALP6	CF
Locality:	Tharros	Orosei	Gerrei	Barisardo	Orosei	M. ferro	Planargia	Guspini	R. Girone	C. Ferrato
Classification	BA	BA	AOB	AOB	AOB	HAW	AOB	HAW	BSN	T
SiO_2	54.76	52.93	49.28	50.82	47.60	48.05	49.71	48.59	45.55	60.34
TiO_2	1.48	1.60	1.91	2.09	1.75	3.05	2.38	3.09	3.11	0.87
Al_2O_3	14.85	16.78	14.12	14.79	15.53	14.01	14.07	15.24	15.13	16.51
Fe_2O_3	11.49	10.05	10.14	11.25	10.14	11.25	11.40	11.33	11.63	6.56
MnO	0.13	0.13	0.16	0.15	0.14	0.17	0.15	0.14	0.16	0.12
MgO	5.28	6.08	9.97	6.54	10.82	4.65	7.54	5.71	7.51	1.24
CaO	6.84	7.88	7.75	8.29	8.42	8.13	8.07	8.55	10.32	3.35
Na_2O	3.50	3.41	4.28	2.80	3.58	4.14	3.80	3.68	3.47	4.72
K_2O	0.80	0.72	1.41	1.76	0.81	1.55	1.52	2.44	2.22	4.15
P_2O_5	0.19	0.19	0.98	0.50	0.38	1.06	0.60	0.76	0.39	0.36
LOI	1.23	0.87		1.76	1.59	4.58	1.54	1.23		2.18
Sc	17		17	22		19	20	18	20	
V	151	157	178	175	193	238	199	243	254	29
Cr	276	251	386	268	386	54	301	152	212	1
Ni	121	140	219	177	295	52	196	87	126	6
Rb	15	13	45	19	26	67	60	57	49	99
Sr	494	513	859	643	681	1561	821	902	970	321
Y	12	16	19	20	18	37	24	24	28	46
Zr	103	107	223	207	156	681	201	260	223	508
Nb	11	14	47	39	31	107	56	72	70	74
Ba	385	346	887	760	689	1537	1115	620	528	1022
La	11.3	14.2	42.1	37.9	26.5	110.4	42.3	49.3	47.1	67.4
Ce	23.0	29.8	78.0	72.2	52.0	215.6	85.6	95.5	96.4	130.4
Pr	2.85	4.02	8.78	8.08	6.16	23.3	9.78	10.8	12.1	15.0
Nd	14.2	18.0	33.6	32.6	25.2	86.2	37.5	47.9	46.6	58
Sm	3.50	4.28	6.40	6.22	4.87	13.6	6.95	9.37	8.27	10.74
Eu	1.39	1.57	2.18	2.06	1.63	4.11	2.38	2.85	2.53	2.81
Gd	3.29	4.15	5.27	5.19	4.36	10.1	5.90	7.39	7.02	9.30
Tb	0.48	0.54	0.71	0.73	0.60	1.35	0.81	0.94	1.01	1.43
Dy	2.56	3.01	3.97	4.09	3.22	6.89	4.09	5.16	5.66	7.84
Ho	0.48	0.55	0.62	0.76	0.55	1.20	0.69	0.89	0.88	1.48
Er	1.11	1.37	1.75	1.75	1.47	3.09	1.71	2.38	2.26	4.00
Tm	0.14	0.19	0.23	0.24	0.20	0.40	0.23	0.30	0.35	0.64
Yb	0.94	1.05	1.61	1.50	1.16	2.51	1.39	1.89	1.97	3.97
Lu	0.15	0.16	0.20	0.21	0.18	0.36	0.20	0.27	0.30	0.61
Hf	2.57	2.74	4.81	5.04	3.46	15.5	5.68	6.04	4.80	11.8
Pb	3.4	3.4	6.6	4.8	2.0	18.1	5.2	3.4	4.0	11.8
Th	1.7	2.2	7.3	4.8	3.4	13.5	5.8	6.3	5.9	14.9
U	0.3	0.4	1.6	1.0	0.8	3.4	1.4	1.6	1.3	3.5
Zr/Nb	9.4	7.5	4.8	5.3	5.0	6.4	3.6	3.6	3.2	6.8
Ba/Nb	35	24	19	20	22	14	20	9	8	14
La/Nb	1.02	1.00	0.90	0.98	0.85	1.03	0.76	0.69	0.67	0.91
Nb/U	32	33	29	38	40	32	40	46	54	21
Ce/Pb	6.8	8.8	11.9	15.1	25.4	11.9	16.5	28.4	24.1	11.1
U/Pb	0.10	0.13	0.25	0.22	0.38	0.19	0.27	0.47	0.33	0.30
$(La/Yb)_n$	8.1	9.1	17.6	17.0	15.4	29.7	20.5	17.6	16.1	11.5
$^{87}Sr/^{86}Sr$	0.70512	0.70465	0.70433	0.70437	0.70442	0.70435	0.70434	0.70315	0.70401	0.70487
$^{143}Nd/^{144}Nd$	0.51235	0.51247	0.51258	0.51257	0.51257	0.51256	0.51257	0.51289	0.51285	0.51271

Table 1 (*continued*)

Type:	UPV	UPV	UPV	UPV	UPV	UPV	UPV	RPV	RPV	RPV
Sample:	MGV245	MGV51	MGL23	MGV223	MGV76	MGV236	MGV238	MGV249	ALP6	CF
Locality:	Tharros	Orosei	Gerrei	Barisardo	Orosei	M. ferro	Planargia	Guspini	R. Girone	C. Ferrato
Classification	BA	BA	AOB	AOB	AOB	HAW	AOB	HAW	BSN	T
$^{206}Pb/^{204}Pb$	17.554	17.826	17.872	17.844	17.860	17.638	18.010	19.422	19.227	18.840
$^{207}Pb/^{204}Pb$	15.595	15.594	15.578	15.552	15.596	15.569	15.609	15.665	15.640	15.657
$^{208}Pb/^{204}Pb$	37.722	38.016	37.836	37.920	37.942	37.923	38.151	39.135	39.105	38.977
TNd_{DM} (Ga)	1.24	0.99	0.63	0.65	0.65	0.57	0.63	0.26	0.29	

The isotopic data were obtained at the laboratory of S.O.E.S.T. (University of Hawaii at Manoa) with techniques described in [48]. Blanks, uncertainties in the measurements and values of reference standards pertinent to this study are fully described in [49]. Major elements, Sc, V, Cr, Ni, Rb, Sr, Y, Zr, Nb and Ba were obtained with X-ray fluorescence spectrometry at Naples, according to procedures and analytical uncertainties described in [50], and Florence. LOI was analysed with standard gravimetric methods at Naples. Lanthanides, Hf, Pb, Th and U were obtained with inductively coupled plasma–mass spectrometry at CRPG, Nancy, France [51]. UPV: unradiogenic Pb volcanic rocks; RPV: radiogenic Pb volcanic rocks. The model ages were calculated using $^{147}Sm/^{144}Nd_{DM} = 0.24$ and $^{143}Nd/^{144}Nd_{DM} = 0.5131$. BA = tholeiitic basaltic andesite; AOB = alkali olivine basalt; HAW = hawaiite; BSN = basanite; T = trachyte.

mafic rocks have variable SiO_2 (45.6–55.0 wt%), MgO (10.8–4.7 wt%), relatively high Cr and Ni (max. 390 and 300 ppm, respectively), and decreasing incompatible element contents from alkaline to tholeiitic rocks (e.g. Nb from 107 to 11 ppm; Zr from 680 to 100 ppm; TiO_2 from 3.1 to 1.5 wt%). Increasing Zr/Nb and $(La/Yb)_n$ (the subscript 'n' means chondrite-normalised) from alkaline to tholeiitic rocks (Zr/Nb from 3.2 to 9.4, and $(La/Yb)_n$ from 8.1 to 30) can be related to their variable degrees of partial melting (estimated to be 4–5% vs. 10–12%, respectively [16]) from a substantially similar mantle source. The $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ ratios of the mafic PSV range from 0.70315 to 0.70512 and from 0.51235 to 0.51289 ($\varepsilon_{Nd} = -5.6$ to +4.9), respectively. The tholeiitic rocks tend to have only slightly higher $^{87}Sr/^{86}Sr$ and lower $^{143}Nd/^{144}Nd$ than the mafic alkaline rocks (Table 1). In the $^{87}Sr/^{86}Sr$ vs. $^{143}Nd/^{144}Nd$ diagram, the data plot close to the model Bulk Earth composition, but with lower ε_{Nd}, except for the hawaiite of Guspini and the basanite of Rio Girone, which show more 'depleted' isotopic compositions (Fig. 2). The rocks with the lowest $^{143}Nd/^{144}Nd$ do not overlap the composition of the Neogene to Quaternary mafic orogenic volcanic rocks from the eastern side of the Tyrrhenian Sea (i.e. Roman Province and the Aeolian Islands), and show a distinct trend to less radiogenic $^{87}Sr/^{86}Sr$ (Fig. 2). With the exception of the southernmost samples (Guspini, Rio Girone and Capo Ferrato), strong Sr–Nd isotopic differences exist also between the Sardinian Plio–Pleistocene and the Italian mafic anorogenic (Na-alkaline and tholeiitic) rocks (i.e. Mt. Etna, Hyblean Mts., Pantelleria and Linosa). These latter cluster in the 'depleted' quadrant ($\varepsilon_{Sr} < 0$ and $\varepsilon_{Nd} > 0$). Two distinct rock groups are also identified in the $^{208}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ (not shown) and $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ diagrams (Fig. 2). $^{206}Pb/^{204}Pb$ ranges from 17.55 to 19.42, with most samples having values < 18; $^{207}Pb/^{204}Pb$ ranges from 15.55 to 15.66 and $^{208}Pb/^{204}Pb$ from 37.72 to 39.13, with most samples < 38.3. The lowest $^{206}Pb/^{204}Pb$ ratios are by far the least radiogenic values reported both for the Italian volcanic rocks and for the Cenozoic European volcanic province in general. Again, the Guspini, Rio Girone and Capo Ferrato samples are isotopically distinct from the other samples, with the highest $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ (Table 1, Fig. 2).

We therefore divide the PSV into two groups: (1) the unradiogenic Pb volcanic rocks (hereafter UPV; Gerrei, Orosei–Dorgali, Planargia, S. Pietro Baunei, Barisardo, Montiferro and Tharros; northern to central–southern Sardinia) with $^{206}Pb/^{204}Pb < 18$, relatively low $^{143}Nd/^{144}Nd$

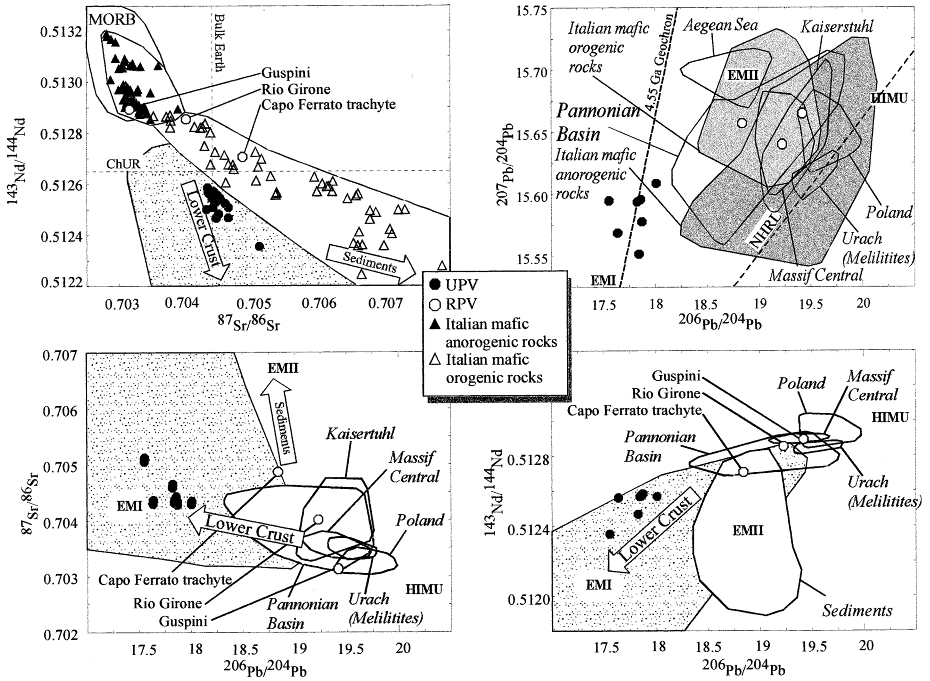

Fig. 2. Sr–Nd–Pb isotopic diagrams for the Plio–Pleistocene volcanic rocks of Sardinia. Filled circles: UPV; open circles: RPV. Field of modern sediments from [32,33]. The range of the lower crust is from [41,52]. See text for references on the Italian volcanics. Cenozoic European volcanic rocks from [1,3,16,26] and references therein.

(<0.51246) and high $^{87}Sr/^{86}Sr$ (~ 0.7044), and (2) the radiogenic Pb volcanic rocks (hereafter RPV; Guspini, Rio Girone and Capo Ferrato; southern Sardinia), with $^{206}Pb/^{204}Pb > 18$, relatively high $^{143}Nd/^{144}Nd$ (>0.51246) and low $^{87}Sr/^{86}Sr$ (<0.7040, in the mafic samples). The UPV have $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ higher than the Northern Hemisphere Reference Line of oceanic basalts (NHRL [18]). They plot very close to, or at, the geochron, approximately along a single-stage lead growth curve with μ ($^{238}U/^{204}Pb$) $= 8.35$. The RPV plot fairly close to, but above the NHRL, overlapping the field of the Italian and European Na-rich anorogenic volcanic rocks (Fig. 2).

In mantle-normalised diagrams (Fig. 3) the alkali basalts and the tholeiites of the UPV have the highest peaks at Ba, whereas the mafic RPV have their highest peak at Nb. The UPV have relatively high Ba/Nb (>14) and La/Nb (>0.8) when compared to the Ba/Nb (8–9) and La/Nb (<0.7) of the mafic RPV (Table 1, Fig. 4). The Nb/U ratios of the RPV (average 49.6) fall within the MORB–HIMU range of 47 ± 10 [19], whereas the UPV show lower values of this ratio (average 33.9) overlapping with the EMI field (Nb/U ≈ 35 [19]). Also the range of Ce/Pb of the RPV (24–28) overlaps with the MORB–HIMU (HIgh MU = $\mu = {}^{238}U/^{204}Pb$) field (Ce/Pb ≈ 26 [20,21]), and is higher than that of the UPV (Ce/Pb = 7–15).

The slightly higher $^{87}Sr/^{86}Sr$ (0.70487), Ba/Nb (14), La/Nb (0.88) and lower $^{143}Nd/^{144}Nd$ (0.51271) of the Capo Ferrato trachyte, with respect to the other RPV rocks, is likely to be related to small amounts of crustal contamination during the ascent, contemporaneous with extensive crystal fractionation from an alkali basaltic parental magma. Similarly, the sample from Tharros, which is a tholeiitic basaltic andesite and does not carry mantle-derived xenoliths, has the highest $^{87}Sr/^{86}Sr$ (0.70512) and Ba/Nb (35), and the lowest $^{143}Nd/^{144}Nd$ (0.51235) among the UPV, again compatible with coupled minor crustal contamination and fractional crystallisation.

The southernmost outcrops of Guspini, Rio Girone and, possibly, the mafic, uncontaminated,

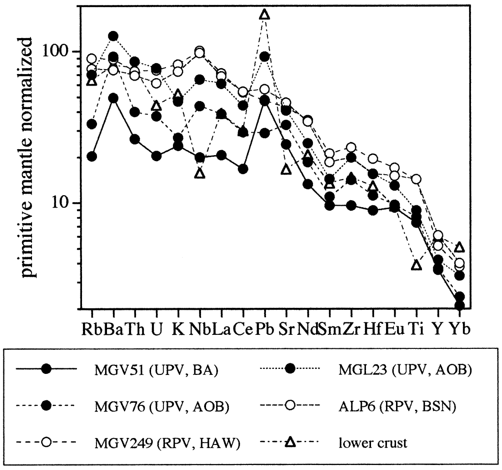

Fig. 3. Primitive mantle-normalised patterns of mafic UPV (filled symbols) and RPV (open symbols) rocks. Mantle normalisation values after [53]. The average lower crustal estimate is from Wedepohl [39]. BA = tholeiitic basaltic andesite; AOB = alkali olivine basalt; HAW = hawaiite; BSN = basanite.

parental magmas of the Capo Ferrato trachyte are the only Sardinian Plio–Pleistocene volcanic rocks with the RPV chemical and isotopic signature.

4. Discussion

There are two main observations drawn from the chemical and isotopic characteristics of the Sardinian Plio–Pleistocene volcanic province: (1) the identification of unusually unradiogenic Pb compositions in most of the mafic rocks of the PSV, and (2) the presence of a chemical and isotopic discontinuity in the mantle sources of the southern part of the PSV.

The UPV are chemically distinct from all other volcanic rocks cropping out in the circum-Mediterranean area. For example, the Italian anorogenic volcanic rocks, which belong to sodic alkaline and tholeiitic series, have indications of being derived from variable contribution of HIMU-like and MORB sources (e.g. low large ion lithophile to high field strength element ratios (LILE/HFSE) and $^{87}Sr/^{86}Sr$ ($\sim$0.7028–0.7036), and high $^{143}Nd/^{144}Nd$ ($\sim$0.5128–0.5131) and $^{206}Pb/^{204}Pb$ ($\sim$19–20)) [22–25]. These characteristics are present also in the other Cenozoic anorogenic mafic volcanic rocks throughout Europe (e.g. [1,3,26]) (Fig. 2). In contrast, the well-known high LILE/HFSE ratios and $^{87}Sr/^{86}Sr$ (0.705–0.717) and low $^{143}Nd/^{144}Nd$ (0.5121–0.5127) of the Italian mafic orogenic rocks (calc-alkaline to ultrapotassic series) have been linked to a subduction-modified upper mantle [27–31]. The geochemical characteristics of the Italian mafic orogenic rocks are currently best interpreted as the result of the recent introduction of pelitic ± calcareous meta-sediments, upper crustal rocks, or melts derived from these, into variably residual-to-fertile peridotites [27,31]. Meta-sedimentary materials, as well as upper crustal rocks, are indeed characterised by high LILE/HFSE ratios (e.g. Ba/Nb > 80), high $^{87}Sr/^{86}Sr$ (0.707–0.726), and high $^{206}Pb/^{204}Pb$ (18.5–19.8) [32,33].

The unradiogenic Pb and Nd isotopic composition and the mildly radiogenic $^{87}Sr/^{86}Sr$, the highest peaks at Ba and Pb in the mantle-normalised patterns, and the relatively high Ba/Nb and La/Nb of the UPV make them akin to the EMI oceanic basalts [20,34–36], and to some continental flood basalts (e.g. the high-Ti continental flood basalts of the Paranà basin, [37]), for which recycling of lower crustal material or melting of old

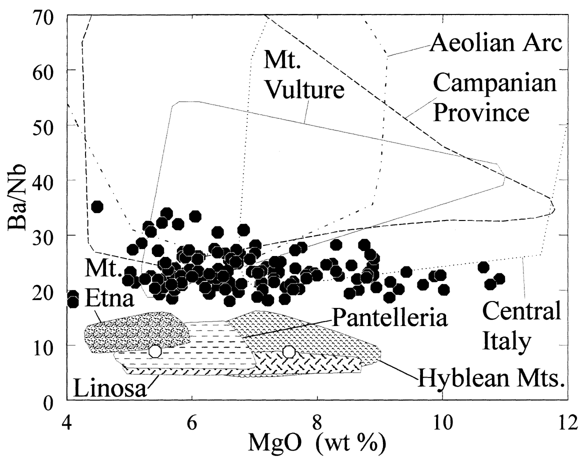

Fig. 4. Ba/Nb vs. MgO for the PSV (data from this study and [16]). Filled circles: Plio–Pleistocene rocks from central and northern Sardinia; open circles = Guspini and Rio Girone. The trachyte sample of Capo Ferrato (MgO = 1.24 wt%; Ba/Nb = 14) has been omitted. See text for the references on the Italian volcanics.

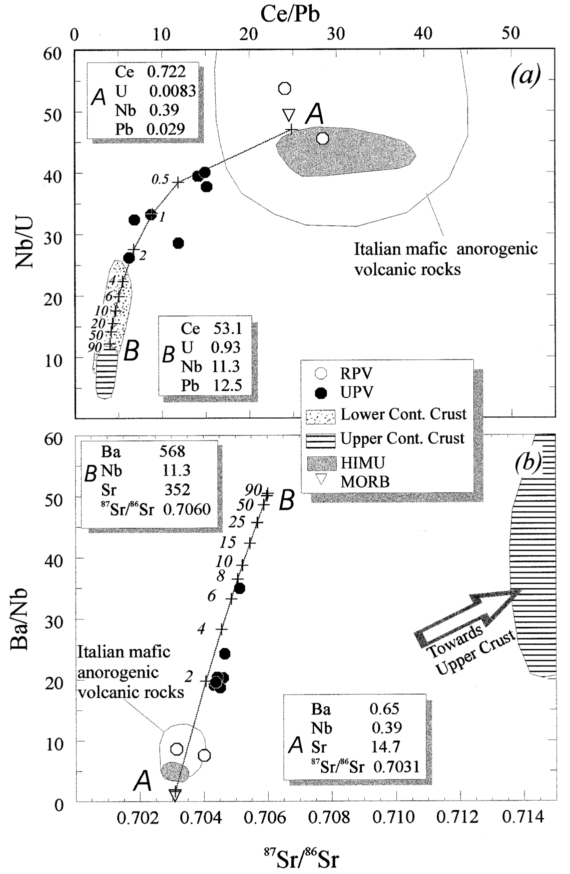

Fig. 5. Nb/U vs. Ce/Pb (a) and $^{87}Sr/^{86}Sr$ vs. Ba/Nb (b) diagrams for the PSV. Also shown is the mixing lines between depleted MORB mantle (component A) and the estimate of the European lower continental crust [39] (component B). Ce and Nb of DMM are from [54], while U and Pb have been calculated on the basis of values of Ce/Pb (25) and Nb/U (47) [19]. The fields of upper and lower continental crust from [55], HIMU field (Rurutu island, Austral chain) from [21]. The RPV plot close to the HIMU range, whereas the UPV lie on the mixing line between component A and component B. Numbers in italics represent the percent of lower continental crust component added to a MORB mantle.

lithospheric mantle has been proposed [37,38]. Indeed, average lower crustal rocks could have high LILE/HFSE ratios (e.g. Ba/Nb = 50–60; see also Fig. 3), and relatively low Rb/Sr, Sm/Nd and U/Pb, due to the extraction of anatectic melts, decreasing Rb and U relative to Sr and Pb, and/or to addition of mafic components from the mantle (e.g. the European lower crustal estimate of Wedepohl [39]). Given the estimated parent/daughter isotope ratios of typical lower continental crust, it could evolve to relatively low time-integrated $^{87}Sr/^{86}Sr$, $^{143}Nd/^{144}Nd$ and $^{206}Pb/^{204}Pb$ [40,41]. Taking into account the repeated orogenic cycles recorded in the island, it is possible that slices of mafic lower crust could have been subducted and mixed to depleted mantle beneath Sardinia, to form the lithospheric source of the later UPV mafic volcanic rocks. A mixture of depleted and lithospheric mantle, or lithospheric mantle without the direct involvement of lower crust or depleted mantle, could be suitable sources of the mafic UPV. Nevertheless, the ultimate chemical features of the lithospheric mantle involved should still be caused by mixing of older depleted mantle with subducted lower crust. It is extremely difficult to distinguish between lithospheric mantle that was metasomatised by mafic lower crust-derived melts, and the products of a mixture between depleted mantle and mafic lower crustal lithologies. Moreover, models without involvement of crust, though certainly possible, appear only to shift back the ultimate cause of the low Nb/U, Ce/Pb, $^{206}Pb/^{204}Pb$ and the other chemical and isotopic characteristics of the UPV, which are typical features due to interaction with lower crust. We therefore propose involvement of lower crust in the UPV genesis both on geological and geochemical grounds. Crustal materials show widely variable lead isotopic composition, but only lower crustal rocks actually reach the least radiogenic Pb compositions (e.g. $^{206}Pb/^{204}Pb$ down to 13; e.g. [41]).

To test this hypothesis, mixing calculations between the compositions of incompatible element-depleted mantle and European bulk lower crust were performed (Fig. 5). The chosen ratios between similarly incompatible elements (and the isotope ratios) have the advantage of being relatively insensitive to the extent of partial melting, and therefore, could be taken as representative of the source ratios. The results show that the UPV sources could be accounted for by 1–3% of a lower crustal component added to depleted mantle.

The Nd_{DM} model ages (DM = referred to depleted mantle isotopic composition; cf. Table 1) range from 1.2 to 1 Ga for the tholeiitic basaltic andesites, from 0.69 to 0.65 Ga for the mafic, mantle xenolith-bearing, alkali basalts of the UPV, and from 0.29 to 0.26 Ga for the mafic RPV (Table 1). It is unclear whether these ages could represent the age of the source stabilisation; however, the model ages of the alkaline UPV are similar to those of the Panafrican tectonic event (900–450 Ma [3]), whereas the Nd model ages of the RPV are close to the late stages of the Hercynian Orogeny. The model ages of the UPV should be considered both as a minimum age of the mixing event and the age of the crust added, in view of the mixing model between depleted mantle and lower crust we proposed above, but, anyway, as the contribution of the lower crust dominates the Nd and Sm budget, the ages indicate that mixing with Precambrian crust could be a viable petrogenetic model.

We suggest that the chemical and isotopic characteristics of the UPV were inherited from melting lithospheric mantle (as proposed for EMI end-member by [20]) modified by introduction of low-Rb/Sr, -Sm/Nd and -U/Pb, lower crust, during the Panafrican event, or, in more recent times (Hercynian?), by subduction of late Precambrian crust. Lower crust recycling could have taken place via: (1) pre- to syn-tectonic subduction (before or during the Hercynian continent–continent collision) with décollement between upper crust (thrusted and napped) and lower crust (subducted), as observed during Pyrenean Orogeny and now in the Alpine belt, or (2) sinking of a dense thickened lithospheric root after continent–continent collision during late Carboniferous–early Permian (extensional stages of the Hercynian Orogeny).

Evidence of old (Panafrican/Hercynian) subduction-related modifications in the European subcontinental mantle comes mainly from late Palaeozoic lamprophyres (e.g. Massif Central, France [42] and Black Forest, Germany [2]), and from mantle (e.g. [43]) and crustal xenoliths (e.g. [44]) hosted in Cenozoic European alkaline volcanic rocks. Such evidence seems to lack or be diluted in the Cenozoic European volcanic record [3]. Therefore, the chemical and isotopic composition of the UPV would represent the lacking lithospheric source end-member among the Cenozoic European Volcanic Province [16].

The chemical and isotopic characteristics of the RPV (and, by analogy, of the Neogene-to-Recent sodic anorogenic mafic volcanic rocks of Italy), could have been inherited by source modifications that occurred during or shortly after the Hercynian Orogeny, among several of which, recycling of dehydrated, subducted, oceanic lithosphere of the Palaeozoic proto-Tethys Ocean, could be a possibility. The relatively uniform HIMU-like composition of the Cenozoic European volcanic rocks and of the RPV could be viewed, in the light of recent experimental works (e.g. [45–47]), as the product of asthenospheric mantle modified by small amounts of subducted, residual, oceanic lithosphere, characterised by low Rb, Pb and Rb/Sr, and high Nb/U, Ce/Pb and U/Pb ratios.

The reason why lithospheric melts derived from sources with old subduction-related modifications appear to be relatively rare in the Cenozoic European Volcanic Province is yet an intriguing and unresolved aspect, as is the geodynamic significance of coexisting EMI-like and HIMU-like geochemical characteristics in the Sardinian Plio–Quaternary volcanic region. Strong heterogeneity and sharp boundaries between different mantle domains have been proposed also for other regions in the Mediterranean area (e.g. [27]).

5. Conclusions

The Plio–Pleistocene mafic volcanic rocks of Sardinia are divided into two groups: (1) the abundant unradiogenic Pb volcanic rocks (UPV), characterised by high Ba/Nb (>14), $^{143}Nd/^{144}Nd$ ($\varepsilon_{Nd}<0$), intermediate $^{86}Sr/^{87}Sr$ (~ 0.7044) and low $^{206}Pb/^{204}Pb$ (<18), and (2) rare radiogenic Pb volcanic rocks (RPV), with lower Ba/Nb (8–9) and $^{86}Sr/^{87}Sr$ (<0.7040), and higher $^{143}Nd/^{144}Nd$ ($\varepsilon_{Nd}>0$) and $^{206}Pb/^{204}Pb$ (>19). These rocks occur in distinct areas, and define a significant geochemical discontinuity, roughly running east–west, across the southern part of the island. The isotopic signature of the UPV is not observed in any of the Italian orogenic and anorogenic rocks, nor in the Cenozoic European Volcanic Province. The unradiogenic Pb and Nd isotopic signature of the UPV closely resembles the EMI mantle end-member composition, and is believed to be a source signature related to mixing of depleted mantle with low-$^{87}Sr/^{86}Sr$, low-$^{143}Nd/^{144}Nd$ and Precambrian, low-μ, lower crustal materials. Alternatively, this could be a signature of an anciently metasomatised lithospheric mantle. The rarer mafic volcanic rocks cropping out in the southern sector of the island (Guspini and Rio Girone, and the evolved trachyte of Capo Ferrato, i.e. the RPV) match the HIMU-like geochemical characteristics found in other Italian sodic anorogenic volcanic rocks, and could be related to recycling of the palaeo-subducted proto-Tethys oceanic slab.

The chemical and isotopic characteristics of the Sardinian Plio–Pleistocene rocks and those of the roughly coeval Roman Magmatic Province imply that completely different mantle processes occurred in the central Mediterranean area, and that no genetic or tectonic relationships between these two provinces can be allowed.

Acknowledgements

We gratefully thank John Mahoney for his hospitality at S.O.E.S.T. and useful suggestions on early versions of the manuscript, and Sandro Conticelli for analysing some samples with XRF. Thanks to P. Brotzu for his continued interest in this project and useful suggestions, and to G. Secchi for assistance in the field trips. Michael Roden and an anonymous reviewer provided very helpful official reviews. This research was granted by the Italian MURST (40%, P. Brotzu, and 60%, L.M. and V.M.).***[EB]***

References

[1] M. Wilson, H. Downes, Tertiary–Quaternary extension-related alkaline magmatism in western and central Europe, J. Petrol. 32 (1991) 811–849.

[2] E. Hegner, M. Kolbl-Ebert, J. Loeschke, Post-collisional Variscan lamprophyres (Black Forest, Germany): ^{39}Ar-^{40}Ar phlogopite dating, Nd, Pb, Sr isotope, and trace element characteristics, Lithos 45 (1998) 395–411.

[3] K.H. Wedepohl, A. Baumann, Central European Cenozoic plume volcanism with OIB characteristics and indications of a lower mantle source, Contrib. Mineral. Petrol. 136 (1999) 225–239.

[4] A. Von Quadt, U-Pb zircon and Sm-Nd geochronology of mafic and ultramafic rocks from the central part of the Tauren window (eastern Alps), Contrib. Mineral. Petrol. 110 (1992) 57–67.

[5] J.A. Tait, V. Bachtadse, W. Franke, H.C. Soffel, Geodynamic evolution of the European Variscan Fold Belt: palaeomagnetic and geological constraints, Geol. Rundsch. 86 (1997) 585–598.

[6] S. Tommasini, G. Poli, A.N. Halliday, The role of sediment subduction and crustal growth in Hercynian plutonism: isotopic and trace element evidence from the Sardinia–Corsica batholith, J. Petrol. 36 (1995) 1305–1332.

[7] P.A. Ziegler, Plate tectonics, plate moving mechanism and rifting, Tectonophysics 215 (1992) 9–34.

[8] C. Doglioni, E. Gueguen, F. Sàbat, M. Fernandez, The western Mediterranean extensional basins and the Alpine orogen, Terra Nova 9 (1997) 109–112.

[9] P. Brotzu, The calcalkaline volcanism in Sardinia, Per. Mineral. 66 (1997) 1–231.

[10] V. Morra, F.A.G. Secchi, L. Melluso, L. Franciosi, High-Mg subduction-related Tertiary basalts in Sardinia, Italy, Lithos 40 (1997) 69–91.

[11] V. Morra, F.A.G. Secchi, A. Assorgia, Petrogenetic significance of peralkaline rocks from Cenozoic calcalkaline volcanism form SW Sardinia, Italy, Chem. Geol. 118 (1994) 109–142.

[12] C. Doglioni, P. Harabaglia, S. Merlini, F. Mongelli, A. Peccerillo, C. Piromallo, Orogens and slabs vs. their direction of subduction, Earth Sci. Rev. 45 (1999) 167–208.

[13] L. Beccaluva, L. Civetta, G. Macciotta, C.A. Ricci, Geochronology in Sardinia: results and problems, Rend. Soc. It. Mineral. Petrol. 40 (1985) 57–72.

[14] M. Lustrino, L. Melluso, V. Morra, F. Secchi, Petrology of Plio–Quaternary volcanic rocks from central Sardinia, Per. Mineral. 65 (1996) 275–287.

[15] S. Scarascia, A. Lozej, R. Cassinis, Crustal structures of the Ligurian, Tyrrhenian and Ionian seas and adjecent onshore areas interpreted from wide-angle seismic profiles, Boll. Geof. Teor. Appl. 36 (1994) 5–19.

[16] M. Lustrino, Petrogenesis of Plio–Quaternary Volcanic Rocks from Sardinia: Possible Implications on the Evolution of the European Subcontinental Mantle: Ph.D. Thesis, Consorzio Università di Catania e Napoli Federico II, 1999, 188 pp.

[17] M. Lustrino, L. Melluso, V. Morra, Origin of glass and its relationships with phlogopite in mantle xenoliths from central Sardinia (Italy), Per. Mineral. 68 (1999) 13–42.

[18] S.R. Hart, A large-scale isotope anomaly in the Southern Hemisphere mantle, Nature 309 (1984) 753–757.

[19] A.W. Hofmann, K.P. Jochum, M. Seufert, W.M. White, Nb and Pb in oceanic basalts: new constraints on mantle evolution, Earth Planet. Sci. Lett. 79 (1988) 33–45.

[20] A.W. Hofmann, Mantle geochemistry: the message from oceanic volcanism, Nature 385 (1997) 219–229.

[21] C. Chauvel, W.F. McDonough, G. Guille, R. Maury,

R.A. Duncan, Constrasting old and young volcanism in Rurutu island, Austral chain, Chem. Geol. 139 (1997) 125–143.
[22] S. Esperança, G.M. Crisci, The island of Pantelleria: a case for the development of DMM–HIMU isotopic compositions in a long-lived extensional setting, Earth Planet. Sci. Lett. 136 (1995) 167–182.
[23] M. D'Orazio, S. Tonarini, F. Innocenti, M. Pompilio, Northern Valle del Bove volcanic succession (Mt. Etna, Sicily) petrography, geochemistry and Sr-Nd isotope data, Acta Vulcanol. 9 (1997) 73–86.
[24] L. Beccaluva, F. Siena, M. Coltorti, A. DiGrande, A. LoGiudice, G. Macciotta, R. Tassinari, C. Vaccaro, Nephelinitic to tholeiitic magma generation in a transtensional tectonic setting: an integrated model for the Iblean volcanism, Sicily, J. Petrol. 39 (1998) 1547–1576.
[25] L. Civetta, M. D'Antonio, G. Orsi, G.R. Tilton, The geochemistry of volcanic rocks from Pantelleria island, Sicily channel: petrogenesis and characteristics of the mantle source region, J. Petrol. 39 (1998) 1453–1491.
[26] M. Granet, M. Wilson, U. Achauer, Imaging a mantle plume beneath the French Massif Central, Earth Planet. Sci. Lett. 136 (1995) 281–296.
[27] L. Beccaluva, P. Di Girolamo, G. Serri, Petrogenesis and tectonic setting of the Roman Magmatic Province, Italy, Lithos 26 (1991) 191–221.
[28] S. Conticelli, A. Peccerillo, Petrology and geochemistry of potassic and ultrapotassic volcanism in central Italy: petrogenesis and inferences on the evolution of the mantle sources, Lithos 28 (1992) 221–240.
[29] G. Serri, F. Innocenti, P. Manetti, Geochemical and petrological evidence of the subduction of delaminated Adriatic continental lithosphere in the genesis of the Neogene–Quaternary magmatism of central Italy, Tectonophysics 223 (1993) 117–147.
[30] M. D'Antonio, G.R. Tilton, L. Civetta, Petrogenesis of Italian alkaline lavas deduced from Pb-Sr-Nd isotope relationships: Earth processes: reading the isotopic code, AGU Geophys. Monogr. 95 (1996) 253–267.
[31] A. Peccerillo, Multiple mantle metasomatism in central-southern Italy geochemical effects, timing and geodynamic implications, Geology 27 (1999) 315–318.
[32] D. Ben Othman, W.M. White, P.J. Patchett, The geochemistry of marine sediments, island arc magma genesis, and crust–mantle recycling, Earth Planet. Sci. Lett. 94 (1989) 1–21.
[33] J. Dostal, B. Cousens, C. Dupuy, The incompatible element characteristics of an ancient subducted sedimentary component in ocean island basalts from French Polynesia, J. Petrol. 39 (1998) 937–952.
[34] A. Zindler, S.R. Hart, Chemical geodynamics, Annu. Rev. Earth Planet. Sci. 14 (1986) 493–571.
[35] B.L. Weaver, The origin of ocean island basalt end-member compositions: trace element and isotopic constraints, Earth Planet. Sci. Lett. 104 (1991) 381–397.
[36] J.J. Mahoney, W.M. White, B.G.J. Upton, C.R. Neal, R.A. Scrutton, Beyond EM-1 lavas from Afanasy–Nikitin rise and the Crozet archipelago, Indian Ocean, Geology 24 (1996) 615–618.
[37] D.W. Peate, C.J. Hawkesworth, M.S.M. Mantovani, N.W. Rogers, S.P. Turner, Petrogenesis and stratigraphy of the high-Ti Urubici magma type in the Paraná flood basalt province and implications for the nature of 'Dupal'-type mantle in the south Atlantic region, J. Petrol. 40 (1999) 451–473.
[38] S.R. Hart, A. Zindler, Constraints on the nature and development of chemical heterogeneities in the mantle, in: W.R. Pelter (Ed.), Plate Tectonics and Global Dynamics, Gordon and Breach, London, 1989, pp. 261–387.
[39] K.H. Wedepohl, The composition of the continental crust, Geochim. Cosmochim. Acta 59 (1995) 1217–1232.
[40] R.E. Zartman, S.M. Haines, The plumbotectonic model for Pb isotopic systematics among major terrestrial reservoirs – a case for bi-directional transport, Geochim. Cosmochim. Acta 52 (1988) 1327–1339.
[41] A.N. Halliday, A.P. Dickin, R.N. Hunter, G.R. Davies, T.J. Dempster, P.J. Hamilton, B.G.J. Upton, Formation and composition of the lower continental crust: evidence from Scottish xenolith suites, J. Geophys. Res. 98 (1993) 581–607.
[42] L. Turpin, D. Velde, G. Pinte, Geochemical comparison between minettes and kersantites from Western European Hercynian Orogen: trace element and Pb–Sr–Nd constraints on their origin, Earth Planet. Sci. Lett. 87 (1988) 73–86.
[43] T.J. Dunai, H. Baur, Helium, neon, and argon systematics of the European subcontinental mantle: implications for its geochemical evolution, Geochim. Cosmochim. Acta 59 (1995) 2767–2783.
[44] E.R. Neumann, J. Martì, J. Mitjavila, E. Wulff-Pedersen, Origin and implications of mafic xenoliths associated with Cenozoic extension-related volcanism in the Valencia Trough, NE Spain, Mineral. Petrol. 65 (1999) 113–139.
[45] M.J. Cordery, G.F. Davies, I.H. Campbell, Genesis of flood basalts from eclogite-bearing mantle plumes, J. Geophys. Res. 102 (1997) 20179–20197.
[46] T. Kogiso, K. Hirose, E. Takahashi, Melting experiments on homogeneous mixtures of peridotite and basalt: application to the genesis of ocean island basalts, Earth Planet. Sci. Lett. 162 (1998) 45–61.
[47] G.M. Yaxley, D.H. Green, The effects of eclogitic heterogeneities on upper mantle melting, Ofioliti 24 (1999) 190–191.
[48] J.J. Mahoney, C. Nicollet, C. Dupuy, Madagascar basalts: tracking oceanic and continental sources, Earth Planet. Sci. Lett. 104 (1991) 350–363.
[49] J.J. Mahoney, R. Frei, M.L.G. Tejada, X.X. Mo, P.T. Leat, T.F. Nägler, Tracing the Indian Ocean mantle domain through time: isotopic results from old west Indian, east Tethyan, and south Pacific seafloor, J. Petrol. 39 (1998) 1285–1306.
[50] L. Melluso, V. Morra, P. Brotzu, A. Razafiniparany, V. Ratrimo, D. Razafimahatratra, Geochemistry and petro-

genesis of the Cretaceous flood basalt sequence of northern Madagascar, J. Afr. Earth Sci. 24 (1997) 371–390.

[51] K. Govindaraju, G. Mevelle, Fully automated dissolution and separation methods for inductively coupled plasma atomic emission spectrometry rock analysis. Application to the determination of the rare earth elements, J. An. At. Spectrosc. 2 (1987) 615–621.

[52] R.S. Cohen, R.K. O'Nions, J.B. Dawson, Isotope geochemistry of xenoliths from east Africa: implications for development of mantle reservoirs and their interaction, Earth Planet. Sci. Lett. 68 (1984) 209–220.

[53] S.S. Sun, W.F. McDonough, Chemical and isotopic systematics of oceanic basalts: implications for mantle composition and process, in: A.D. Saunders, M.J. Norry (Eds.), Magmatism in the Ocean Basins, Geol. Soc. Spec. Publ. 42 (1989) 313–345.

[54] D. McKenzie, R.K. O'Nions, The source regions of ocean island basalts, J. Petrol. 36 (1995) 133–159.

[55] S. Gao, T.C. Luo, B.R. Zhang, H.F. Zhang, Y.W. Han, Y.K. Hu, Z.D. Zhao, Chemical composition of the continental crust as revealed by studies in East China, Geochim. Cosmochim. Acta 62 (1998) 1959–1975.