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Debated topics of modern igneous petrology

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ABSTRACT. — Progress made thanks to the great amount of high-quality geochemical and isotopic data gathered over the last twenty years has allowed important insights into igneous petrology (e.g., Continental Flood Basalt petrogenesis, mantle source characterization, geophysical models of mantle plume systems, primary melt compositions, isotopic systematics of crustal and mantle domains, etc.). This large dataset has also been used to relate the compositional characteristics of igneous rocks with specific tectonic settings and to infer the geodynamic processes involved. However, inferring a tectonic setting mainly on the basis of geochemical constraints may fail, because partial melts with substantial compositional differences can originate from the same source, and the same melts may have been generated in different tectonic settings. Moreover, geochemical characterization of the main mantle components is still hotly debated, even in terms of concepts such as asthenosphere and lithosphere. Asthenospheric mantle is quite often believed to be a geochemically homogeneous convecting domain, whereas the lithospheric one is thought to be a variably enriched, heterogeneous, non-convecting reservoir, capable of retaining geochemical and isotopic gradients for periods of time exceeding 2 Ga. These assumptions are clearly over-simplifications, particularly when relationships between physical and geochemical mantle characteristics are not properly constrained.

What emerges from recent literature is the «misuse» of petrological concepts tending towards the most convenient explanations and «effectively shutting out the entire creative thought process of the human mind» (Sheth, 1999). More appropriate use of petrological data is necessary to stimulate true scientific growth, especially as regards our knowledge of mantle-crust dynamics.

RIASSUNTO. — I progressi legati all'incremento di disponibilità negli ultimi vent'anni di dati geochimici ed isotopici di elevato standard qualitativo hanno permesso di raggiungere importanti conclusioni nel campo della petrologia ignea (es., petrogenesi dei basalti continentali, caratterizzazione delle regioni mantelliche, modelli geofisici sui sistemi di plume di mantello, composizione dei fusi primitivi, sistematica isotopica delle riserve crostali e mantelliche, etc.). Questa disponibilità di dati è stata anche utilizzata per mettere in relazione le caratteristiche composizionali delle rocce ignee a specifici ambienti tettonici e per ipotizzare i processi geodinamici attivi. Tuttavia, il riconoscimento di un ambiente tettonico sulla base dei dati geochimici può portare in errore, dal momento che fusi parziali prodotti dalla stessa sorgente possono avere sostanziali differenze composizionali. A questo va aggiunto il fatto che fusi con una particolare caratteristica

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possono essere prodotti in ambienti tettonici differenti. La caratterizzazione geochimica dei principali componenti di mantello è ancora fortemente dibattuta, anche in termini di concetti quali astenosfera e litosfera. Il mantello astenosferico è, piuttosto comunemente, ritenuto come un dominio convettivo geochimicamente omogeneo, mentre il mantello litosferico è considerato come una riserva mantellica variamente arricchita, eterogenea e non convettiva, capace di conservare gradienti geochimici ed isotopici per periodi superiori a 2 Ga. Queste assunzioni sono chiaramente sovrasemplificazioni, particolarmente laddove le relazioni tra le caratteristiche fisiche e geochimiche del mantello non sono chiaramente evidenziate.

Ciò che emerge dalla letteratura recente è l'uso scorretto di concetti petrologici tendenti verso le spiegazioni più convenienti. Un uso più appropriato dei dati petrologici è reso necessario per stimolare vera crescita scientifica, specialmente per quanto riguarda la nostra conoscenza delle complesse dinamiche di interazione crosta-mantello.

KEY WORDS: *Geochemistry, lithosphere, asthenosphere, CFB, mantle plume, metasomatism, experimental petrology.*

INTRODUCTION

The two major points addressed in this paper are: 1) how an identical set of raw geochemical data, when discussed by different authors, can lead to contrasting models and interpretations, and: 2) how the same material can yield confusing results when different methods are applied to it. The first «hot topic» is dealt with in the sections entitled «Care in using geochemical data» and «Continental basalt genesis». Many case studies are presented for which different conclusions are inferred from the same raw data (e.g., the Roman Comagmatic Province). With regard to continental basalt genesis, the contrasting geophysical and geochemical definitions of mantle reservoirs and the meaning of geophysical models viewed from a petrological standpoint (e.g., the concept of mantle plume) is highlighted.

The sections entitled «Experimental petrology», «geothermo barometers» and «Geochemical jargon and other diatribes» concern especially the second point addressed in this paper. Contrasting results obtained using differing experimental methods are briefly reported here, emphasizing analytical problems such as accuracy and reproducibility, as well as interpretations, e.g., equilibrium vs. disequilibrium, and the significance of calculated thermodynamic parameters.

CARE IN USING GEOCHEMICAL DATA

The igneous petrologist's dream is, first, to characterize as well as possible the system s/he is studying and to start with a model; then, to reconstruct the liquid line of descent of a magmatic series, to infer the contribution of the crust and lithospheric and asthenospheric mantle to magma chemistry, to hypothesize the reason for magmatic activity, to constrain temperature and depth of formation and, lastly, to arrange the above conclusions in a geodynamic scenario which takes into account other Earth Sciences disciplines, such as geophysics, field geology, and so on.

In the few last years, a general dearth of innovative ideas in some fields of Earth Sciences has become apparent, particularly in petrology. Many researchers have balanced the scarcity of stimulating and provocative models with increasingly accurate analytical methodologies. Nowadays, for example, it is not difficult to find papers with trace element concentrations reported in the range of ppt (parts per trillion) or LREE contents in mantle olivine (below 0.01 × times chondrite). On one hand, these ultraprecise data have certainly helped petrologists in their work; on the other, they too often become the only objective of the research.

Modern petrology cannot continue without the help of geochemistry, although this tool should be adopted with great care. For example, similar magma compositions may be found in very different tectonic settings (collisional *vs* extensional), depending on the age and type of the crustal-mantle dynamics which affected the site of magmatism and its style of differentiation. Moreover, magmatic records with extremely different trace elements and isotopic signatures are often found in a single plutonic complex or volcanic edifice. This highlights the need to enlarge one's point of view in order to avoid wrong interpretations. Thus, geochemical data must be tested against other Earth Sciences disciplines before they can be used to prove a model.

Too often petrological models are based only on geochemical data, without taking into account or without giving the right weight to field geology and diagnostic petrography. Examples may be taken from the Paleozoic evolution of Europe and the help given by petrology in palinspastic restorations identifying paleotectonic settings (e.g., Lustrino, 2000a). Active continental margins and within-plate settings have sometimes been inferred *only* on geochemical bases by studying magmatic products, even if they were equivocal. For instance, during the Paleozoic, a subduction-related setting of Cambro-Ordovician metavolcanics from the French Massif Central and of the Permian ophiolitic body of Finero (Ivrea zone, Italy) was postulated mainly on the basis of the Nb negative anomaly and LILE enrichment (Briand *et al.*, 1991, and Zanetti *et al.*, 1999, respectively). But the same geochemical characteristics of the same rocks were related by other authors (Pin and Marini, 1993, and Cortesogno *et al.*, 1998, respectively) to crustal contamination, excluding any primary «orogenic» imprint. Again, the nature of the subducted/obducted crust variously outcropping along Paleozoic Europe has been alternatively interpreted as originating in a back-arc setting (Bodinier *et al.*, 1988) or in a pure oceanic domain (Pin, 1990). Clearly, the two latter interpretations (compressional and extensional, respectively) configure two opposite paleotectonic environments and palinspastic restorations.

The same contrasting conclusions were drawn for the Eocene Colville igneous complex (western USA). Here, the «calc-alkaline signature» of mafic to sialic rocks was assumed to indicate contemporaneous subduction tectonics (e.g., Holder and Holder, 1988). More recent works, however, have pointed out the problems inherent in such a subduction model, questioning the existence of a volcanic arc in the area during the Eocene. The «calc-alkaline affinity» of these products, coupled with a «subduction signature» revealed by LILE-rich and HFSE-poor features, has been related to melting of heterogeneous Proterozoic crustal sources not requiring any contemporaneous subduction (Morris *et al.*, 2000).

The same problem also arises for more recent magmatic rocks for which a more clearly defined geodynamic setting exists. One example is the Miocene-Pliocene volcanic rocks (tholeiitic and calcalkaline basalts *s.l.* up to shoshonitic and lamproitic compositions) from the Betic orogen in SE Spain. The negative anomalies of Ta-Nb and other High Field Strength Elements have been linked either to subduction-related modifications (Torres-Roldàn *et al.*, 1986) variably modified by crustal contamination (Benito *et al.*, 1999) or to continental crust contamination of metasomatized asthenospheric mantle melts (Turner *et al.*, 1999).

Bimodal magmatic activity (e.g., basaltic and rhyolitic *s.l.*) has generally been related to within-plate to extensional tectonic settings in many papers (e.g., Lorenz and Nicholls, 1984; Matte, 1991; Pin and Marini, 1993). Although it is true that bimodal activity during extensional tectonics is a common aspect (e.g., the Paranà-Etendeka magmatic province, with abundant basalts and associated quartz latite-rhyolite: Piccirillo *et al.*, 1988; Ewart *et al.*, 1998; Marques *et al.*, 1999), it is also common to find bimodal activity in compressional regimes, e.g., back-arc settings, as in Hokkaido (Japan: Takagi *et al.*, 1999). In this case, mantle-derived rhyolite, with extremely unradiogenic Sr ($^{87}Sr/^{86}Sr$ = 0.7034-0.7049) and radiogenic Nd (ε_{Nd} from +2.5 to +5), resulting from low-degree partial melting of the metasomatized mantle wedge, is associated with coeval basaltic (*s.l.*) rocks. This rock association at present clearly does not indicate a purely extensional (e.g., within-plate) tectonic setting.

More classical evidence of the «opinability» of geochemical data is the interpretation of the potassic and ultrapotassic Italian Cenozoic to Present magmatism (e.g., see Lustrino, 2000b). These magmatic products were first hypothesized to reflect within-plate extensional settings, without any evidence of subduction-related modifications: a mantle plume track (Vollmer, 1976) or an East Africa rift-like setting (e.g., Cundari, 1994; Stoppa and Cundari, 1995; Stoppa and Woolley, 1997; Ayuso *et al.*, 1998). But other authors (e.g., Beccaluva *et al.*, 1991; Conticelli and Peccerillo, 1992; Nelson, 1992; D'Antonio *et al.*, 1996, 1999; Peccerillo, 1999) strongly favoured subduction-modified mantle sources to explain the major and trace element and isotopic characteristics of the potassic to ultrapotassic magmatism in Italy. The reasons for the opinion of the first group of authors, who favour a within-plate setting, are the similarities of the Italian potassic volcanic rocks with those from the western branch of the East Africa rift (e.g., Rogers *et al.*, 1998), in terms of major and trace elements (except for HFSE) and Sr-Nd-Pb isotope ratios. For the ultra-alkaline magmatic activity in central Italy, Stoppa and Lavecchia (1992) proposed a primary origin linked to decarbonation of carbonate-melilitic magma. In their opinion, the high Cr and Ni contents of these rocks do not represent primary features, but are connected to the presence of numerous peridotitic xenoliths. The low Cr-Ni contents of the original melts, coupled with high LILE/HFSE ratios, suggest slight participation of olivine and spinel during partial melting processes and, together with the low Nb, indicate a fractionation feature with respect to a carbonatitic mantle component. The radiogenic Sr character of the same rocks is related to metasomatism by an agent with aged primary high Rb/Sr, and thus related to stable continental settings (such as the Toro-Ankole region in Uganda and the Sapucai-Asunçion graben in Paraguay) rather than to subduction tectonic settings (Stoppa and Lavecchia, 1992). The greatest difference between the potassic Italian and African volcanic rocks (i.e., higher LILE/HFSE and lower Nb and Ti contents of the Italian rocks) has also been interpreted in terms of the key role of Ti-phases (e.g., rutile, ilmenite and other phases) during magma production (Di Battistini *et al.*, 1998).

Nelson (1992), while observing the presence of potassic to ultrapotassic igneous rock suites in both tectonic settings («anorogenic» *vs* «orogenic»), argues that the peculiar features of both rock groups are inherited from subducted lithosphere, including sediments. Anorogenic sub-group potassic magmas have Sr and Nd isotopic ratios and incompatible trace element characteristics similar to those with long-term storage of subducted sediments within their sources (Nelson, 1992). The absence of any ocean island with pure potassic affinity remains, according to Nelson (1992), as the strongest evidence that potassic to ultrapotassic magmatism is unlikely to have been derived from hot-spot plume sources unrelated to subduction metasomatism.

The subduction-related setting of the Italian potassic rocks is mainly based on fractionated hygromagmatophile elements with low HFSE and high LILE/HFSE, typical geochemical features of subduction-related melts (e.g., Rogers and Setterfield, 1994; fig. 1). Interaction between crustal and mantle reservoirs is thought to be one of the main processes in the genesis of Italian potassic magmatism, although crustal assimilation *en route* to the surface is ruled out by simple mass balance calculations. Rather, this volcanism has been modelled by the partial melting of mantle sources metasomatized by bulk melting of subducted sediments of upper crustal composition (Beccaluva *et al.*, 1991; Conticelli and Peccerillo, 1992).

In particular, the aspect of low HFSE contents is thought to result from the presence of accessory residual phases with high K_D^{HFSE} in the source. It is important to mention here the origin of the fractionation between HFSE

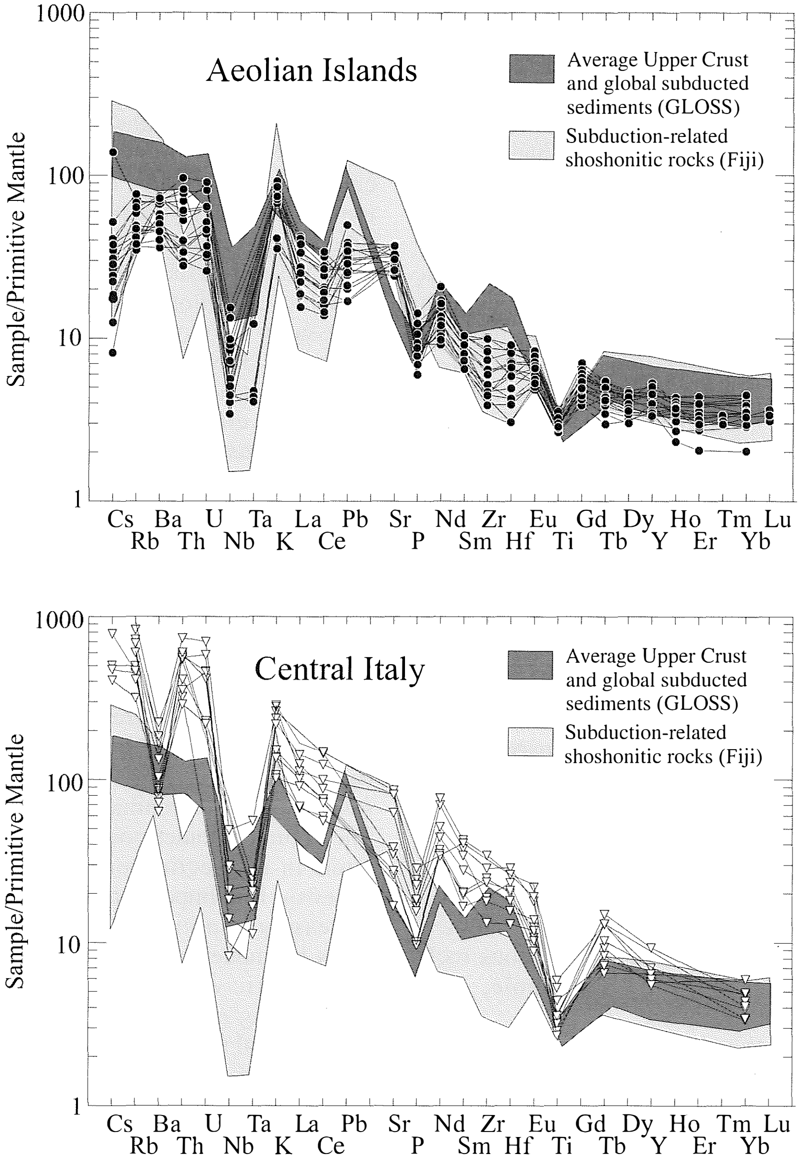

Fig. 1 – Primitive mantle (Sun and McDonough, 1989) normalized diagram for selected Pliocene to Present Italian mafic volcanic rocks (references in Lustrino, 2000b). Heavy shaded field = upper crust estimates (Wedepohl, 1995; Gao *et al.*, 1998a) and GLOSS composition (GLOSS = GLObal Subducting Sediments; Plank and Langmuir, 1998); light shaded field = typical shoshonitic volcanic and plutonic rocks from subduction-related setting (Tavua volcano, Fiji; Rogers and Setterfield, 1994). Pliocene to present Italian mafic volcanic rocks display typical trace element features of both subduction-related magmas and upper crust. Simple bulk mixing between N-MORB mantle-like composition and upper crustal composition has been ruled out on geochemical modelling grounds by several authors (e.g., Conticelli and Peccerillo, 1992). Instead, for Italian rocks, an origin from mantle sources metasomatized by slab-derived fluids/melts is preferred. The origin of these modifications has been related either to ancient processes or to coeval (Miocene to Present) subduction processes responsible for the opening of the Tyrrhenian sea as a back-arc basin.

and other incompatible trace elements (e.g., LILE and REE). A characteristic of island-arc igneous rocks is their low HFSE/LILE and HFSE/REE, together with fractionation between $HFSE^{4+}$ (Zr, Hf) and $HFSE^{5+}$ (Nb, Ta). There are two hypotheses explaining this decoupling: *a)* partial melting of either the subducting eclogitic slab or the overlying mantle wedge, in the presence of rutile as residual phase; *b)* dehydration metamorphic reactions in the subducting amphibolitic slab, even in the presence of rutile. The high K_D rutile/melt and rutile/fluid for HFSE (e.g., Foley *et al.*, 2000; Stalder *et al.*, 1998, respectively) is in fact well-known. Moreover, rutile can also fractionate $HFSE^{4+}$ from $HFSE^{5+}$ in both melt and fluid pairs: the accepted D values are: $D_{Nb,Ta} > D_{Zr,Hf} >> D_{LILE,REE}$, with a difference of 1-2 orders of magnitude between Nb-Ta and Zr-Hf and 2-3 orders of magnitude between HFSE and LILE-REE.

The role of volatile-bearing phases (such as amphibole or mica) as repositories of HFSE is now gaining consensus (e.g., Ionov *et al.*, 1997). In particular, the role of Ti-pargasite and kaersutite in fractionating $HFSE^{4+}$ from $HFSE^{5+}$ and HFSE from LILE and REE has recently been evidenced (e.g., Tiepolo *et al.*, 2001). However, these new experimental data on mantle amphibole do not indicate the possibility of achieving the trace element fractionation seen in ocean-arc igneous rocks taking into account only amphibole, since D^{amph}_{HFSE} is generally < 1.

Whatever the case, melt or fluid, the presence of residual rutile is essential in explaining the HFSE troughs in normalized incompatible element variation patterns. Experimental studies show that rutile is not a stable phase during partial melting of peridotite, because of the reaction: *rutile+olivine = orthopyroxene+ilmenite* (Ryerson and Watson, 1987), but it can coexist at mantle depths with eclogitic or non-peridotitic mantle (e.g., pyroxenitic bodies or veins). This experimental evidence has often been used to infer an «orogenic» setting for igneous rocks with negative HFSE spikes, where other disciplines of Earth Sciences have failed to propose definite settings. In summary, fractionation intra-HFSE and between HFSE/LILE-REE has been considered to reflect subduction-related metasomatism, caused either during eclogitic slab partial melting or dehydration of amphibolitic subducting oceanic crust.

If these conclusions seem to be valid in clear-cut subduction systems, they are not necessarily so in other contexts where negative spikes of HFSE are present. This signature, seen in continental igneous rocks may in fact be related either to: *a)* contamination with anatectic crustal melts *en route* to the surface, or *b)* presence of pyroxenitic-eclogitic slices in the deep mantle. Continental crust is characterized by extremely low HFSE contents compared with LILE and REE (e.g., Rudnick, 1995; Gao *et al.*, 1998a) and an overall trace element similarity with island-arc igneous rocks (fig. 1). Mixing or AFC-like processes of mantle melts with crustal lithologies of crustal-derived melts may therefore be responsible for the HFSE signature of continental igneous rocks not requiring subduction-modified sources or active continental orogenic settings. The argument under point *b)* will be addressed later, in the section on «continental basalt genesis». It is important to recall here only the increasing consensus gained in the last few years by models requiring the involvement of pyroxenitic-eclogitic slices (possibly representing paleo-subducted slabs or sunken lower crustal restites) disseminated in the upper mantle to explain the geochemical characteristics of continental and oceanic igneous rocks (e.g., Cordery *et al.*, 1997; Hofmann, 1997; Kogiso *et al.*, 1997; Lassiter *et al.*, 2000; Tatsumi, 2000; Yaxley, 2000). Rutile in pyroxene-rich veins in the mantle (e.g., eclogitic or pyroxenitic assemblages) can potentially produce patterns typical of island-arc basalts (Foley *et al.*, 2000).

The presence of minerals with high K_D^{HFSE} in the residual mantle has been proposed not only in subduction-related settings (e.g., the

Italian case), but also in within-plate magmatic settings. In fact, Gibson *et al.* (1999) hypothesized a lithospheric origin for the potassic rocks (of lamproitic affinity) from the Lages district of southern Brazil. In this sense, the negative peak of HFSE of these Brazilian rocks in primitive mantle-normalized diagrams was related to the presence of residual Ti-phases during melting, without any evidence of modifications of the source by subduction-related processes (Gibson *et al.*, 1999). Instead, the same geochemical signature (low HFSE contents) of the lamproites from Tuscany (Italy) and the Betic Orogen (SE Spain) were related by Peccerillo (1999) and Turner *et al.* (1999) to contamination of upper crustal materials in the lithospheric mantle during subduction processes.

It is clear not only that the same magma type may be produced in different tectonic settings and that different melts may be produced in the same tectonic settings, but, above all, that interpretation of geochemical data is in some cases, if not frequently, biased. Moreover, to think that the geochemistry of a magma type reflects a particular mantle source does not imply that the processes responsible for its features acted contemporaneously with magmatic activity. As an example, if the assumption of a genetic relation between the geochemical features (K-enrichment and LILE/HFSE decoupling) of Italian rocks and a subduction-modified source is accepted, a space-time relation between magmatism and subduction processes is arbitrary (e.g., Tamburelli *et al.*, 2000). The subduction signature of Plio-Pleistocene magmatism in Italy may be the consequence of ancient compressive tectonics recorded in the lithospheric mantle (e.g., Di Battistini *et al.*, 1998).

I would now like to discuss one of the most intriguing topics in Earth Sciences and the role of petrology in solving it, i.e., the *recycling of continental crust into the mantle*. In particular, I would like to address the possibility of lower crust recycling via delamination and detachment of an overthickened lithospheric keel in a collisional zone. Recycling of lower crustal lithologies into the mantle has gained much consensus in the last ten years (e.g., Wedepohl, 1995; Gao *et al.*, 1998b; Tatsumi, 2000) The model proposes the gravitational instability of the dense lithospheric root accreted in continent-continent collisional zones. This keel may, under its own weight, collapse and detach from the overlying crust, at the point where the lower continental crust, which has ductile behaviour, accumulates major stresses. In this way, lower crust (plus possibly the uppermost lithospheric mantle) is returned deep into the mantle. These lithologies are distinguishable both in terms of major and trace elements and isotopic ratios from the «normal» asthenospheric mantle (see below) and thus their inference in mantle dynamics, at least in theory, is clear from a geochemical point of view. There are many researchers who, on geochemical grounds, have hypothesized recycling of lower crust into the mantle mainly on the basis of: 1) relative Eu, Sr and transition metal deficiencies in continental crust estimates, related to the detachment of a plagioclase-rich cumulate (with high $K_D^{Eu,Sr}$), lost from the base of the crust and recycled into the mantle (e.g., Wedepohl, 1995; Gao *et al.*, 1998a, 1998b); 2) the evolved nature of the continental crust in terms of major elements, explained by intracrustal differentiation to form evolved melts, ascending buoyantly to upper crustal levels and complementary residual paragenesis (eclogitic) recycled into the mantle (see discussion in Rudnick, 1995); 3) silica-rich glass inclusions in ultramafic xenoliths from Sierra Nevada with Eu/Eu* < 1, low Rb and very low Nb, which would reflect the interaction of mantle mineralogy with lower crustal melts (Ducea and Saleeby, 1998); 4) anomalous compositions of late Archean syenites of Australia (in particular low HFSE and Y) related to delamination by convective thinning of a dense, garnet-rich, lower crust, after partial melting (Smithies and Champion, 1999); 5) unradiogenic Pb ($^{206}Pb/^{204}Pb$ ~17.5) and Ba-rich compositions ($(Ba/Nb)_N > 1$) of within-plate Plio-Pleistocene volcanic rocks

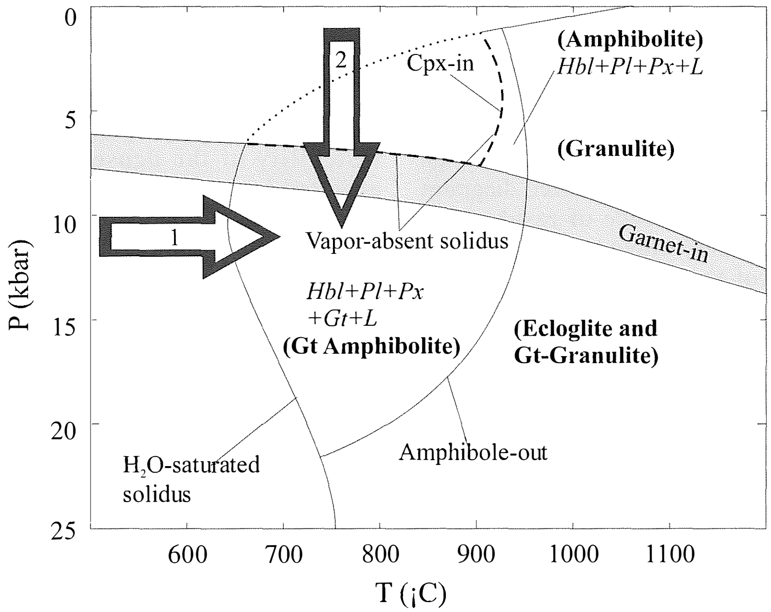

Fig. 2 – Composite phase diagrams for amphibolitic assemblages (whole-rock composition variable from alkali basalt to N-MORB) as deduced from several sources (see text). Continuous line, partly interrupted by dotted line: H_2O-saturated *solidus*. Garnet-in curve represented by shaded area because of contrasting results from experimental studies carried out on differing starting compositions.

Above garnet stability field, vapour-absent *solidus* coincides with H_2O-saturated *solidus*, because excess water provided by breakdown of amphibole may form new garnet during dehydration partial melting. Below garnet stability field, vapour-absent *solidus* is represented by dashed line, coinciding with appearance of clinopyroxene and garnet produced by reaction: *amph* + *pl* = *melt* ± *gt* ± *cpx* ± other phases. Italics: stable mineralogical assemblages; bold: rock types.

Arrow «1»: effect of arrival of hot basaltic batches on amphibolitic lower crust, when system is forced to melt partially by dehydration, not requiring aqueous pore fluid. Assemblage increases in density, due to increasing volumes of garnet substituting amphibole.

Arrow «2»: effect of tectonic piling following continent-continent collision and thickening of lower crust. Here too, lower crustal lithologies are forced to melt partially, with later increase in density of restitic material. If density gradient reaches a critical value, restitic lower crust (represented by garnet amphibolite or eclogite/granulite) may delaminate, detach and sink into upper mantle. See text for further explanations.

from Sardinia (Italy), related to mantle sources modified during the previous Hercynian orogeny and lower crustal digestion (Lustrino, 1999, 2000b; Lustrino *et al.*, 2000); 6) trace element and Sr-Nd-Pb isotopic modelling by Tatsumi (2000), who demonstrated that delamination of the mafic lower continental crust in a continental tectonic setting may explain both the intermediate composition of the continental crust and the isotopic characteristic of the EMI geochemical reservoir in the deep mantle.

Several geophysical data (e.g., anomalous high heat flow and low crustal thickness under continental collisional zones; Gao *et al.*, 1998a), together with several mathematical models (e.g., Marotta *et al.*, 1998; Meissner and Mooney, 1998; Schott and Schmeling, 1998) and experimental studies on natural compositions (e.g., Wolf and Wyllie, 1994), have also been used to «demonstrate» delamination and the possibility of detachment of the lower crust. In particular, lower crust sinking assemblages (eclogite and pyroxenite) are currently thought to represent restites after partial melting of amphibolitic sources. Indeed, experimental petrology indicates that: 1) amphibolite rocks (with alkali basalt to

MORB-like composition) start to melt in H_2O-undersaturated conditions at low temperatures (less than 750°C) at lower crustal pressures (~ 10-20 kbar); 2) dehydration-melting of these lithologies leaves an eclogitic (garnet + omphacitic clinopyroxene; $P \geq 15$ kbar) or granulitic/pyroxenitic (augite + orthopyroxene + garnet ± plagioclase; P ~ 10-15 kbar); 3) dehydration melting reactions produce a strong increase in density from ~ 3.1 (amphibolitic assemblage; amphibole + plagioclase ± quartz ± sphene) to > 3.5 g/cm^3 (granulitic/eclogitic assemblage; Wolf and Wyllie, 1994) of the former lower crust. This restite is clearly negatively buoyant and may thus sink into the upper mantle. The increased density of lower crustal lithologies (amphibolites) may therefore be related to a temperature increase (e.g., arrival of hot basaltic batches from below which forces the system to enter the anhydrous *solidus*), but it may also be due to an increase in pressure (e.g., as a consequence of tectonic piling), forcing the system to enter the garnet stability field (located for lower crustal assemblages at ~ 8-12 kbar in the interval 700-1200°C; Wolf and Wyllie, 1994; Rapp and Watson, 1995, and references therein; fig. 2). All these researches highlight delamination of the lower crust as «the most important process for recycling of the continental crust in the mantle» (Gao *et al.*, 1998b).

The problem arises if we consider one of the classical sites of crustal detachment, such as the Alboran Sea (western Mediterranean). Its formation has been related to crustal and lithospheric delamination and detachment of the Alpine-Betic Orogen mainly on tomographic bases (e.g., Spakman *et al.*, 1993; Zeck, 1996; Carminati *et al.*, 1998), although other geophysicists have proposed alternative solutions and demonstrated the fallacy of this model, on the basis of paleotectonic restorations and study of the structural and P-T evolution of the Alboran basement rocks (e.g., Doglioni *et al.*, 1997; Prosser *et al.*, 1999).

It is important to recall that the missing lower crust observed by seismology (e.g., Moho at shallower levels than expected) does not necessarily imply a true detachment process. Indeed, the shift of the Moho to shallower levels may also be the result of the increasing density of the lower crust up to values comparable with those of the upper mantle, thus making these two reservoirs seismically indistinguishable (Wolf and Wyllie, 1994).

In conclusion, as also noted by Rudnick (1995), effective recognition of crustal delamination remains a difficult proposition.

CONTINENTAL BASALT GENESIS

The genesis of continental basalts, particularly those associated with huge extrusion rates and volumes, is one of the most intriguing subjects of modern petrology. The main dispute focuses on the contribution of the lithospheric mantle and asthenosphere to their genesis and on the role of continental crust during their emplacement. One important aspect to be stressed is that the asthenosphere is geophysically defined as the mantle domain characterized by convective movements, but convective movements do not occur in the lithosphere. This geophysical definition of the lithosphere does not completely fit the geochemical definition of lithosphere (Enriched Mantle; EM component) and asthenosphere (Depleted Mantle; DM component) (Anderson, 1996). In order to solve this problem – i.e., the coherence of geophysical and geochemical definitions – it is important to consider mantle dynamics and the evolution of these reservoirs. First, mutual exchanges exist between lithospheric mantle and asthenosphere, in terms of both thermal and physical budgets, and they should thus be seen as an open system. In this sense, it is not possible to identify clear-cut, unquestionable evidence of asthenospheric and lithospheric mantle compositions. There are many possibilities of interaction and mixing between the two: 1) contamination of asthenosphere by delamination of the lithospheric keel in overthickened collisional regions (e.g., Kay and Kay, 1993; Marotta *et al.*, 1998); 2) contamination of asthenosphere by subduction processes (e.g., Hofmann, 1997); 3) contamination of lithosphere by asthenospheric metasomatic fluids and melts (e.g., Menzies and Hawkesworth, 1987); 4) contamination of asthenospheric melts after their passage through the lithosphere *en route* to the surface (e.g., Marzoli *et al.*, 2000). Moreover, the geochemical uniformity of the asthenospheric mantle largely depends on the style of cratonization of the region: geochemical gradients can be effectively removed by convecting cells only after geologically sound times (e.g., > 200-400 Ma). For example, for the Cenozoic European Volcanic Province, Wilson and Downes (1991) hypothesized an asthenospheric origin (mainly HIMU-DM components which variably interacted with EM compositions; see below). The EM imprint was related by these authors to modifications occurring during Paleozoic subduction processes. The relatively short period which elapsed since this orogeny (~ 300-400 Ma) caused the lack of homogenization of the asthenospheric mantle by convection. This unhomogeneized (Hercynian subduction-related) asthenospheric mantle was responsible for the HIMU-DM-EM transitional character of the Cenozoic European Volcanic Province.

The lithosphere itself may be divided into an upper elastic region (Mechanical Boundary Layer), cooler than the 650°C isotherm, and a lower viscous region (Thermal Boundary Layer), with temperatures between the 650°C and 1250°C isotherms (Wilson *et al.*, 1995b). The Mechanical Boundary Layer is able to retain geochemical and isotopic heterogeneities for very long times (up to 2-3 Ga), in contrast with the Thermal Boundary Layer, which can only retain them for shorter periods (from ten to a few hundred Ma; Wilson *et al.*, 1995b). Therefore, the lowermost region of the lithospheric mantle is in some way transitional to the asthenosphere. Unequivocal geochemical tracers of a lithospheric or asthenospheric mantle are still lacking (e.g., Milner and le Roex, 1996). For example, the lithospheric mantle estimates of McDonough (1990, 1994) indicate, in primitive mantle-normalized diagrams, positive peaks at Nb, with $(Nb/La)_N > 1$. On this basis, the negative peaks at Nb, typical of many Continental Flood Basalts (CFB), were related by Cordery *et al.* (1997) to a mantle plume origin contaminated by paleo-subducted slabs, rather than to a lithospheric mantle origin. Arndt and Christensen (1992) also excluded the lithospheric origin of some CFB, first because of their virtually anhydrous composition and because they are too cold to melt partially, and second because they are characterized by negative Nb spikes, while many CFB show the opposite behaviour (Arndt and Christensen, 1992). It is well known that CFB with both positive and negative spikes do exist (e.g., Melluso *et al.*, 1995). The fact is that, in some cases, a lithospheric origin is ruled out considering only the low-Nb share and assuming a Nb-rich lithospheric mantle (e.g., Cordery *et al.*, 1997); in other cases, the same conclusions (no lithospheric inference in continental basalt genesis) are reached by considering only the high-Nb share of CFB and assuming a Nb-poor lithospheric mantle (e.g., Arndt and Christensen, 1992). In contrast with these conclusions, for many researchers CFB reflect only a lithospheric signature, confining the role of mantle plumes to the mere thermal anomaly necessary to trigger lithospheric partial melting (e.g., Marques *et al.*, 1999). The CFB provinces are now almost unanimously ascribed to features called «mantle plumes». In particular, this model (Campbell and Griffiths, 1990) has been proposed to explain the large volumes of magma produced over relatively small time intervals (generally of the order of a few Ma), although this concept is now also used for relatively small volumes of volcanic rocks (e.g., Ritter *et al.*, 2001; see below).

However, the presence of mantle plumes is not strictly necessary to explain the genesis of CFB: in fact, the existence of mantle plumes is fundamentally requested to explain an anomalously hot mantle responsible for the large volume of magma produced. But these features may also be explained by considering thinned lithosphere, extremely fertile mantle reservoirs, or geometric relations between lithospheric plates of different thickness (e.g., King and Anderson, 1995), thus reducing the role of mantle plumes in the Earth's history (Smith and Lewis, 1999).

The mantle plume concept has recently been criticized by Sheth (1999), whose perplexities mainly regard the following points: *a)* recent studies (e.g., Cordery *et al.*, 1997; Kogiso *et al.*, 1997; Yaxley, 2000) exclude the fact that CFB and OIB originate from melts coming exclusively from the lower mantle; they also require a contribution from subducted oceanic crust; *b)* the rapid turn-off of almost all the CFB (all the large igneous provinces of the Earth lasted only a few Ma and their activity peaked within 1-3 Ma) is very difficult to reconcile with the plume head theory, considering the considerable thermal inertia of igneous rocks; *c)* in the laboratory model of Campbell and Griffiths (1990), mantle plume-like structures can be obtained in non-natural conditions (e.g., no phase changes, no plate motions or structural discontinuities, injection of superheated fluid at the base of another fluid, etc.), whereas other authors (e.g., Anderson, 1996) argued in favour of non-stationary core-mantle boundaries and high Rayleigh numbers for natural systems; *d)* the geographic locations of CFB are not random but are all associated with mobile belts; *e)* in some cases (e.g., Siberian traps; Czmanske *et al.*, 1998), CFB volcanism was not preceded by surface uplift, as required by the classical plume model of Campbell and Griffiths (1990); *f)* the hot spot-CFB connection (required by the mantle plume model) is not a worldwide phenomenon (e.g., there are some igneous provinces, such as the Afar and Siberian traps, which do not appear to be related to hot-spot tracks; *g)* the presence of mantle plumes (i.e., anomalous hot mantle) under any flood basalt province has been ruled out by seismic tomographic evidence (Anderson *et al.*, 1992); *h)* the Hawaii hot-spot track (thought to represent the classical evidence of a plume tail) is not associated with any major flood basalt province (e.g., the impact of the plume head); *i)* regarding all active hot spots, none is currently in the plume-head phase; they all represent plume-tail stages.

All these points lead us to think that the importance of the mantle plume concept in the last ten years has not been properly constrained, particularly as regards CFB petrogenesis. Mantle plume-related origin has been proposed not only for large volumes of magmas but also for areally-limited products such as the Cenozoic volcanism of Europe (Hoernle *et al.*, 1995; Wedepohl and Baumann, 1999; Jung and Hoernes, 2000; Wedepohl, 2000; Ritter *et al.*, 2001). Fundamentally, the mantle plume model alone cannot explain both geochemistry (e.g., Cordery *et al.*, 1997; Marques *et al.*, 1999; Yaxley, 2000) and the geographic location of the CFB provinces (e.g., Vauchez *et al.*, 1997; Tommasi and Vauchez, 2001). In particular, mantle plume models explaining flood basalt provinces require unreasonably high potential temperatures to generate sufficient volumes of magma over a limited time interval. Fluid dynamic models on mantle plumes only composed of peridotite fail to produce CFB volumes of melts and melting rates (Cordery *et al.*, 1997). However, the presence of some basaltic component in a mantle plume source (e.g., 30% eclogite + 70% MORB pyrolite) may provide a viable source for some flood volcanics (Yaxley, 2000). This is because the eclogite *solidus* (which represents the basaltic composition stable at high pressures) is substantially lower than the peridotite *solidus*. Hybrid sources, with slices of subducted slabs stored at great depths within a peridotitic medium, would start to melt early with respect to normal pyrolite. The first portion to melt is, obviously, the basaltic component; partial melts of such a component are SiO_2-rich and highly reactive with peridotitic assemblages. Migration and infiltration of such compositions would produce reactions with mantle olivine and form an orthopyroxene-rich front. Experimental works have shown that the *liquidus* assemblage in equilibrium with such SiO_2-rich melts remains peridotitic (i.e., *opx* + *ol* ± *cpx*; Yaxley, 2000). Subsequent melting of such a metasomatized mantle would thus continue to produce nepheline-normative picritic melts in lower melt proportions, becoming hypersthene-bearing picrites or olivine tholeiites at higher melt fractions and lower pressures (Yaxley, 2000). The key role of the addition of a basaltic component to a MORB pyrolitic composition is not to lower the *solidus* temperature (by only a few tens of degrees) but to increase the melt fraction for a given temperature, enhancing melt productivity by 50-70% compared with normal pyrolite (Yaxley, 2000). In every CFB case study, strong lithospheric control (in terms of both structural and geochemical influence) does exist (e.g., King and Anderson, 1995; Melluso *et al.*, 1995; Vauchez *et al.*, 1997; Sheth, 1999; Tommasi and Vauchez, 2001).

Currently, it is well known that there is no general consensus explaining the genesis of continental flood basalts and their associated magmatic activity (e.g., alkaline plutonic and carbonatitic rocks), as the contribution of lithospheric and asthenospheric components may vary from one CFB province to another and even within a single province (e.g., Deccan traps; Melluso *et al.*, 1995).

EXPERIMENTAL PETROLOGY

The contrasting interpretations of the magmatic record may result from: 1) limited direct observation of the mantle (depths shallower than 100 km: mantle xenoliths and ophiolitic complexes); 2) disagreement about the chemical composition of primary melts; and 3) impossibility of incorporating the geologic time-factor into laboratory experiments. No widespread consensus regarding the composition of mantle melts yet exists. As an example, the contrasting compositional results of primary liquids produced by batch melting of peridotite may be emphasized. Although the composition of the partial melt of a peridotite is known to be basaltic (*s.l.*), great uncertainties still exist both on the location of the mantle *solidus* and the effect of peridotite composition (Hirschmann, 2000). Differences among various model mantle *solidi* reach ~ 60°C in the pressure range of 3-4 GPa and up to 80°C for higher pressures (9-10 GPa; fig. 3). Assuming a standard mantle adiabath of ~ 10°C/GPa, this bias implies differences in *solidus* depth of ~ 20 km and in partial melting productivities (Hirschmann, 2000). In some cases, experimental runs carried out on peridotitic compositions similar in terms of major element abudance (e.g. KLB-1, Kilbourne Hole peridotite, and KR-4003, Kettle River peridotite; Walter, 1998; Herzberg *et al.*, 2000) at the same nominal pressure of 5 GPa yielded differences of ~ 100°C in the anhydrous *solidus* temperature.

Differences in peridotite (and mixtures of peridotite + basalt) compositions are key variables affecting peridotite *solidus*, melt productivity and melt composition, but there is no strong consensus about the effective weight of alkali contents in bulk peridotite, modal clinopyroxene and source refractory degree (expressed by the Mg#). Indeed, the positive correlation between Na_2O+K_2O in the source and *solidus* temperatures at variable pressures found in many experimental studies (e.g., Hirose and Kushiro, 1993; Robinson and Wood, 1998; Robinson *et al.*, 1998) are not observed in other cases (e.g., Pickering-Witter and Johnston, 2000). Moreover, the small effect of basalt addition in lowering the peridotite *solidus* temperature (~ 50°C; Yaxley, 2000) contrasts with increasing alkalies in the mixture (1-2 times higher than normal peridotite). This happens because of the contrasting effect of the clinopyroxene mode, which has the effect of making Na more compatible in residual solids, thus effectively diminishing the *solidus*-lowering ability of Na (Hirschmann, 2000).

Other contrasting results among experimental petrologists emerge with regard to other important parameters such as the SiO_2 contents of peridotite partial melts (Table 1). The SiO_2 of melts related to hydrous peridotite

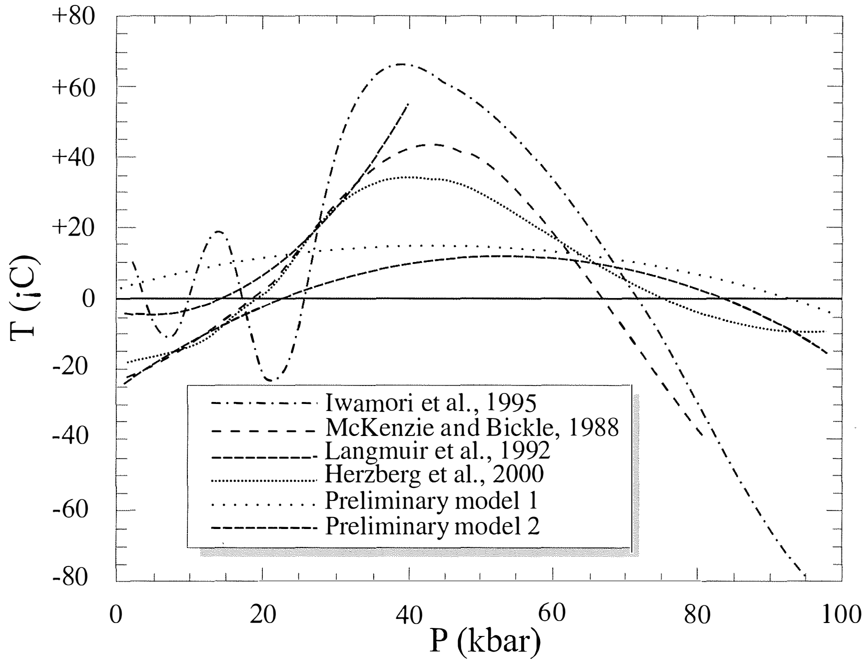

Fig. 3 – Comparison of recommended peridotite *solidus* of Hirschmann (2000) relative to other selected models of peridotite *solidi* from literature. Recommended *solidus* results from peridotite data, with exclusion of enriched and depleted compositions. Positive values of plotted parameter indicate temperatures hotter than Hirschmann's (2000) proposed best-fit model. Preliminary models 1 and 2 are *solidi* obtained by Hirschmann (2000) using entire range of experimental results from literature, not filtered for enriched and depleted compositions, and the *solidi* obtained excluding MORB-pyrolite and similar compositions, respectively. For a complete list of references, see Hirschmann (2000).

was found to be higher (e.g., Hirose, 1997), broadly similar (Kushiro, 1990) or lower (Gaetani and Grove, 1998) compared with that of melts derived from anhydrous peridotite. Moreover, the SiO_2 of near-*solidus* melts, compared with magmas resulting from a higher degree of melting, has been found to be higher (Raterron *et al.*, 1997; Gaetani and Grove, 1998), roughly the same (Takahashi and Kushiro, 1983; Baker and Stolper, 1994) or lower (Mysen and Kushiro, 1977). Lastly, the major element composition of the melt is thought to be independent of (e.g., Hirose and Kushiro, 1993; Kogiso *et al.*, 1997) or dependent on (Robinson *et al.*, 1998) the bulk composition of the starting peridotite. The main discrepancies seem to be related to:

a) experimental technique adopted (diamond aggregate method and «sandwich technique»; Hirose and Kushiro, 1993; Robinson *et al.*, 1998). Both techniques require extensive testing before they can be accepted. On one hand, in the diamond aggregate method, the melt segregates in a layer of diamond powder placed in the charge and there quenches into analyzable pools without interaction with the coexisting minerals (olivine and pyroxenes). However, the melt may not equilibrate at the nominal pressure of the experiment, as the pore space pressure is not isobaric. In addition, the resulting melt is quartz-normative (at 1 GPa; Baker *et al.*, 1995), whereas other studies have generally determined nepheline or hyperstene/olivine normative compositions produced by peridotite partial melting at similar pressures and temperatures (e.g.,

Table 1

Differing results from experimental studies on natural and simplified compositions. Notwithstanding important constraints obtained by experimental petrologists on igneous petrogenesis, note clearly contrasting results on some basic assumptions regarding production of magma and its evolution.

Compared to melts derived from anhydrous peridotite, those derived from partial melting of hydrous peridotite are thought to have SiO_2:		
Higher	Broadly the same	Lower
-Hirose, 1997	-Kushiro, 1990	-Gaetani and Grove, 1998
Compared to magmas formed by higher melting degrees, near-*solidus* partial melts of peridotite are thought to have SiO_2:		
Higher	Broadly the same	Lower
-Raterron *et al.*, 1997 -Gaetani and Grove, 1998	-Takahashi and Kushiro, 1983 -Baker and Stolper, 1994	-Mysen and Kushiro, 1977
Compared to bulk composition of peridotite, the SiO_2 content of partial melts is thought to be:		
Dependent	Independent	
-Robinson *et al.*, 1998	-Hirose and Kushiro, 1993	
Near-minimum anhydrous partial melts in equilibrium with peridotitic assemblages at 10 kbar are though to be:		
Nepheline normative	Quartz normative	
-Falloon *et al.*, 1999 (basaltic composition s.l.)	-Baker *et al.*, 1995 (andesitic composition s.l.)	
The temperature dependence of the partitioning-coefficient for the Fe^{2+}-Mg exchange ($KD^{Ol\text{-}Liq}_{Fe\text{-}Mg}$) is thought to be:		
Positive	Zero	Negative
-Grover *et al.*, 1980	-Roeder and Emslie, 1970 -Ulmer, 1989	-Ford *et al.*, 1983

Falloon *et al.*, 1999, and references therein). On the other hand, the «sandwich technique», which consists of placing a layer of basaltic component (about 10% of the charge) in contact with about 90% of peridotite (in order to force the basaltic melt to be saturated with mantle mineralogy) also presents problems: the experimental bulk composition (i.e., basalt + peridotite) departs from that of the representative mantle material (Baker and Stolper, 1994) and the growth from the liquid of metastable phases during quenching of runs means that determination of partial melt composition by direct electron microprobe analysis is in some cases impossible (Yaxley, 2000);

b) artifact of volatile-free basis normalization of hydrous experimental melts. Hydrous peridotite partial melts are thought to be SiO_2-richer than anhydrous melts, implying strong constraints on the genesis of subduction-related andesite. This conclusion derives mainly from the results of the pioneering studies of Kushiro (1969), who showed the persistence of the incongruent melting of enstatite and the increase of the stability field of olivine up to 3 GPa in hydrous experiments in the Fo-Di-SiO_2 system. Instad, Gaetani and

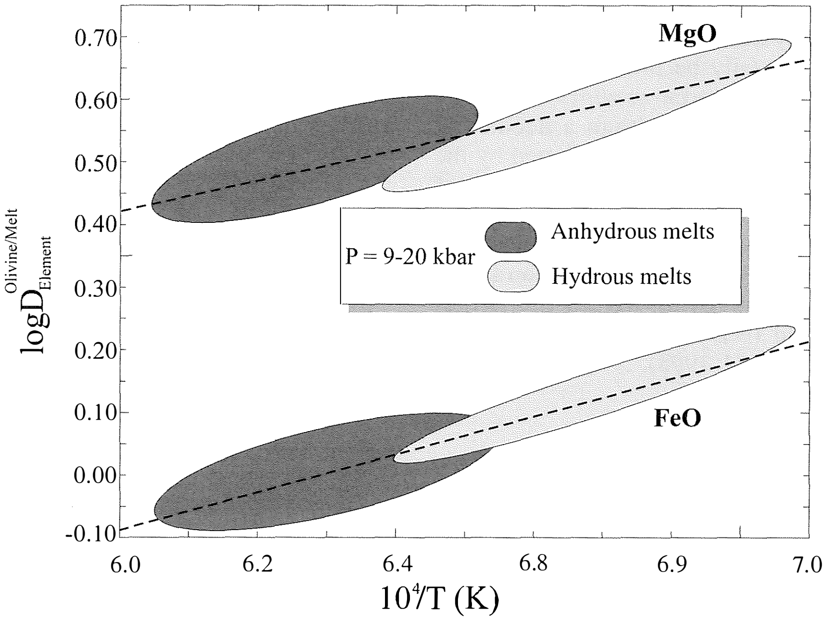

Fig. 4 – Plot of $\log D^{Olivine/Melt}_{FeO}$ and $\log D^{Olivine/Melt}_{MgO}$ vs. inverse temperature. Dashed lines: best-fit lines from Gaetani and Grove (1998), calculated from nominally anhydrous (Kinzler and Grove, 1992; Gaetani and Grove, 1998) and H_2O-bearing experimental silicate melts (Gaetani and Grove, 1998) saturated with a mantle peridotite mineral assemblage. Experimental conditions: P = 9-20 kbar and T = 1170-1370°C. Positive correlation of $\log D^{Olivine/Melt}_{FeO}$ and $\log D^{Olivine/Melt}_{MgO}$ with inverse temperature: melts in equilibrium at relatively low temperatures (e.g., obtained in water-saturated experimental conditions) have lower MgO+FeO and higher SiO_2/(MgO+FeO) rather than higher SiO_2 contents. Modified from Gaetani and Grove (1998).

Grove (1998) demonstrated that the characteristic of hydrous peridotite partial melts, compared with melts from anhydrous assemblage, is not higher SiO_2 but the higher SiO_2/(MgO+FeO) ratio: the apparently higher SiO_2 content of H_2O-bearing peridotite partial melts is an artifact due to normalizing the compositions on a water-free basis. The lower MgO+FeO of these melts (and the higher SiO_2/(MgO+FeO) ratio is a consequence of the lower *solidus* temperature of hydrous starting compositions, as $\log D^{oliv/melt}_{MgO}$ and $\log D^{oliv/melt}_{FeO}$ are negatively correlated with temperature (Gaetani and Grove, 1998; fig. 4).

c) composition of capsules. The analytical problem of capsule material for experimental runs is well known. In particular, solid graphite has often been used instead of Ag-Pd mixtures as a sample container in order to prevent Fe loss of capsule material (e.g., Draper and Green, 1997). But the choice of graphite is not without its problems either, in view of possible reactions between silicate melt and carbon, as indicated by Herzberg *et al.* (2000): $Fe_2O_{3(silicate\ melt)} + C_{(capsule)} = CO_{2(melt)} + 2FeO$. This reaction has been proposed to explain the presence of carbonate as the quench phase from a melt that contained dissolved CO_2 obtained according to the above reaction;

d) uncertainties regarding thermocouple measurements (Baker and Stolper, 1994) and pressure calibration (Herzberg *et al.*, 2000), and

e) partly differing T-P-fO_2 conditions, quenching problems, slow reaction rates in solids, and interlaboratory calibration

(Takahashi and Kushiro, 1983; Hirschmann, 2000). An anhydrous or volatile-rich peridotite source cannot easily be inferred from liquid composition (e.g., Lustrino *et al.*, 1999).

Notwithstanding more than a decade of studies, these uncertainties are reflected by the lack of consensus about the MORB origin of primary mantle melts or polybaric differentiated products, as originating within the garnet stability field (e.g., Bourdon *et al.*, 1996; Salters, 1996) or within the spinel stability field (e.g., Blundy *et al.*, 1998; Robinson and Wood, 1998).

GEOTHERMOBAROMETERS

Another example of extreme difficulty in understanding mantle dynamics arises from the geothermobarometers available for magmatic rocks and mantle xenoliths. In particular, these methods are used in the first case to constrain source regions, and in the second to anchor the geotherm. In both cases, the researcher is left with the possibility of choosing among the most «convenient» results. There are many geothermometers, but currently available geobarometers for basaltic compositions are few. Their models are based on: 1) major element composition of the magma (e.g., Helz and Thornber, 1987; Grove and Juster, 1989; Albarède, 1992; Gudfinnsson and Pressnal submitted); 2) inter-mineral phase equilibria (e.g., Nimis, 1995; Loucks, 1996); 3) relations between phases in equilibrium with the liquid (e.g., Putirka *et al.*, 1996). The results obtained by these and other methods are sometimes not coherent and occasionally yield unrealistic values. There are four main reasons for the failure of geothermometers in many natural systems: *a)* the haplobasaltic or oversimplified starting conditions on which they are based; *b)* analytical difficulties; *c)* different blocking-exchange temperatures recorded by the various mineral pairs; *d)* difficulty in finding true primary melts which have not undergone fractional crystallization.

a) First, oversimplified assumptions represent the largest obstacle in determining true estimates, as the natural system is much more complicated and its parameters cannot easily be defined. The 100-150°C difference between mantle *solidus* obtained for the CaO-MgO-Al_2O_3-SiO_2 system (CAMS; Milholland and Presnall, 1998) and natural peridotites (Herzberg *et al.*, 2000; Hirschmann, 2000) may be taken as an example of the bias between experiments on natural and synthetic systems.

b) Analytical difficulties regard especially electron microprobe work with very low concentrations of elements (some thousands of ppm) in the mineral lattice (e.g., Ca contents in olivine used as a geobarometer; Simkin and Smith, 1970; Kohler and Brey, 1990), alkali loss under the probe beam (e.g., Morgan and London, 1996), or partitioning of ferrous and ferric iron in both lavas and minerals (e.g., Sack *et al.*, 1980; Droop, 1987; Ulmer, 1989).

c) The geothermobarometers based on inter-mineral equilibria mainly concern the orthopyroxene-clinopyroxene join. However, there are many other models based on other mineral pairs: olivine-augite, olivine-spinel, olivine-orthopyroxene-spinel, Fe-Ti oxides, or three feldspars. There are also many geothermobarometers based on liquid-mineral equilibrium exchange: olivine-liquid (Roeder and Emslie, 1970; Ulmer, 1989), clinopyroxene-liquid (Putirka *et al.*, 1996), plagioclase-liquid (Housh and Luhr, 1991), etc.. The much faster diffusive re-equilibration of oxides with respect to silicates (e.g., Scowen *et al.*, 1991) and the different rates of diffusion in the various mineral pairs mean that these models become imprecise when applied to natural systems.

d) Last but not least, it is difficult to find fresh basaltic glass not «spiked» with phenocrysts, and it is almost impossible to sample primitive liquids which have not undergone fractionation and which are free of any phenocryst phase. In addition, all geothermobarometers are based on the assumption that melt and crystals are in equilibrium – a situation which, in natural rocks, is not always the case.

Results obtained with these algorithms must be used with caution, bearing in mind all these approximations. Worthy of note are the results of Neumann *et al.* (1999) on crustal xenoliths from Olot (NW Spain) using three different geobarometers: 1) Grove *et al.* (1989) = 1.8-3.2 kb, with many negative values; 2) Nimis (1995) = 1-7 kb; 3) Soesoo (1997): 2-12 kb, all indicating differing pressures of formation of crustal xenoliths from the atmosphere to the crust-mantle boundary.

GEOCHEMICAL JARGON AND OTHER DIATRIBES

Since the original definition of geochemically distinguishable mantle reservoirs by Zindler and Hart (1986), terms such as DMM (Depleted Morb Mantle), HIMU (HIgh μ), EMI, EMII, EMIII (Enriched Mantle I, II and III), PREMA (PREvalent MAntle) and FOZO (FOcal ZOne) have become more and more frequently applied in igneous petrology. Although these end-member compositions were originally defined only for oceanic volcanism (e.g., Hofmann, 1997), they are currently also applied to the petrogenesis of continental volcanic and plutonic rocks (e.g., Thompson *et al.*, 1998).

With regard to OIB, among the HIMU, EMI and EMII end-member components, only HIMU is *relatively* well characterized in terms of isotopic ratios and trace element characteristics, whereas enriched compositions (EMI and EMII) largely overlap in trace element ratios and cannot be identified on geochemical grounds alone (see discussion in Woodhead and Devey, 1993). Moreover, the formerly known «HIMU-OIB» end-member is actually distinguished on both geochemical and isotopic grounds into two groups: Group 1) S. Helena-like, with extremely radiogenic lead ($^{206}Pb/^{204}Pb > 20$), high μ ($\mu = {}^{238}U/^{204}Pb$; 34-44), relatively low Ce/Pb (~32), and $\Delta 7/4$ generally above the northern Hemisphere Reference Line (NHRL; Hart, 1984); and Group 2) with high but not extreme $^{206}Pb/^{204}Pb$ (19-20), lower m (15-36), higher Ce/Pb (30-45) and negative $\Delta 7/4$ (Vidal, 1992; Thirlwall, 1997). Also, EMI cannot be considered as a single component, and various $^{207}Pb/^{204}Pb$ end-members with low $^{206}Pb/^{204}Pb$ have been proposed (Mahoney *et al.*, 1996). This enriched component is believed to reflect pelagic (Weaver, 1991) or ancient sediments (Dostal *et al.*, 1998; Hofmann, 1989), ancient lithosphere (Milner and le Roex, 1996), lower crust (Hart and Zindler, 1989; Lustrino *et al.*, 2000; Tatsumi, 2000) or primitive mantle (e.g., Roden *et al.*, 1994).

In summary, the EMI signature identifies the imprint of recycled ancient crust (plus or minus ancient pelagic sediments and the ancient lithosphere budget) lodged in the lithospheric mantle (Zhang *et al.*, 1995) or at even greater depths in the deep mantle (Milner and le Roex, 1996) or interaction with lower crustal lithologies (Lustrino *et al.*, 2000). The EMII mantle component is also subduction-related and contains small but significant amounts of recycled continental crust (Hofmann, 1997).

Another debated topic of modern igneous petrology regards the role of volatiles in the upper mantle and the presence of hydrous phases in mantle mineralogy (e.g., Draper and Green, 1997; Lustrino *et al.*, 1999). In particular, the presence of hydrous phases (mainly phlogopite and amphibole) is closely linked to metasomatic processes, although there is a dilemma of the «egg and chicken» type (Wilkinson and Le Maitre, 1987): are hydrous phases the frozen counterparts of metasomatic agents, or is metasomatism the result of destabilization of hydrous phases?

The origin of K itself in the upper mantle remains a question for debate. Potassic compositions are generally attributed to subduction-modified mantle sources (e.g., the Roman Comagmatic Province), whereas the presence of phlogopite and amphibole in MARID xenoliths found in kimberlites (e.g., Phillips, 1991) or their presence in spinel harzburgites from the Mid-Atlantic Ridge (e.g., Bazylev *et al.*, 1999) requires a deep origin, not necessarily related to subduction metasomatism.

Both negative and positive K spikes in primitive mantle-normalized diagrams have been related to the presence of phlogopite and/or amphibole in the mantle source. In the first case (negative spikes), the hydrous phases are thought to remain in the residual mineralogy, thus retaining K in their lattice (e.g., Wilson *et al.*, 1995a; Späth *et al.*, 1996). In the second case (positive spikes and potassic compositions), these phases are thought to participate in the melting process (e.g., Gibson *et al.*, 1999). In some situations (e.g., the Comorean lavas), negative K spikes have been explained as due to the presence of residual phlogopite and amphibole (Späth *et al.*, 1996) or a «normal» mantle signature not requiring any metasomatic hydrous phase (Melluso and Morra, 2000). In yet other cases, in the same district (e.g., Lages, SE Brazil) the occurrence of coeval potassic and sodic magmatism has been related to the alternative role of phlogopite as melting phase (in the first case) or residual phase (in the second) in mantle mineralogy (Gibson *et al.*, 1999).

The constancy of the ratio of two highly incompatible elements is also a quite hotly debated topic. For example, with regard to the Ce/Pb ratio, it has been assumed that cerium is not significantly fractionated with respect to lead during partial melting, due to their very similar K_D (Hofmann, 1988; Miller *et al.*, 1994), whereas Sims and DePaolo (1997) demonstrate the possibility of fractionating these two elements for different degrees of partial melting, lead being more incompatible than cerium. Problems arising from fractional crystallization modelling are mainly based on the uncertainty in determining K_D.

Geochemical and isotopic complexity may also derive from carbonatite metasomatism, particularly *in auge* for the European sublithospheric continental mantle and Italian volcanic rocks (e.g., Stille and Schaltegger, 1996; Trua *et al.*, 1998). The effects of carbonatite metasomatism are difficult to ascertain. For example, according to Hauri *et al.* (1993), carbonatite metasomatism shifts the original Sr-Nd isotopic compositions towards the enriched mantle quadrant (i.e. $\varepsilon_{Sr}>0$ and $\varepsilon_{Nd}<0$), but some African carbonatites are clearly characterized by $\varepsilon_{Sr}<0$ and $\varepsilon_{Nd}>0$ (Simonetti and Bell, 1994; Smithies and Marsh, 1998). Moreover, according to Rudnick *et al.* (1993), one of the peculiar influences of carbonatite metasomatism is to increase the original Nb contents whereas, according to Wiechert *et al.* (1997), carbonatite melt metasomatism produces the opposite result. Moreover, with regard to the tectonic implications of carbonatite metasomatism, according to Rudnick *et al.* (1993), increased Nb contents rule out subduction-related settings. But Zanetti *et al.* (1999) believe that carbonatite metasomatism with increased Nb is also possible in subduction zones, as a consequence of the increased CO_2/H_2O ratio after crystallization of hydrous minerals (amphibole and phlogopite) and CO_2 saturation of the residual melt, with the subsequent separation, by immiscibility, of CO_2-bearing phases (carbonates; Zanetti *et al.*, 1999).

In this context lies the diatribe concerning the origin of the Ca-rich rocks of central Italy, alternatively believed to be true carbonatitic rocks (with clear (?) within-plate tectonic implications; e.g., Stoppa and Woolley, 1997), or mixtures of silicate magmas and geochemically depleted carbonate material (limestone) (e.g., Peccerillo, 1998) with no geotectonic implications. The presence of syn-orogenic carbonatite in the Himalayan belt (NW Pakistan; Tilton *et al.*, 1998) is more evidence of the wrong axiom: carbonatite magmatism = anorogenic setting (see also Mattioli *et al.*, 1989; Zanetti *et al.*, 1999).

CONCLUDING REMARKS

This paper discusses some of the currently most debated topics of igneous petrology. The aim was not to discern which of the various theories are right, but only to investigate possible reasons for these diatribes.

With regard to the geodynamic approach to modern igneous petrology, it is of particular importance to proceed with care when geotectonic setting inferences are based on geochemical data *alone*. This is because similar compositions may be produced in different tectonic settings, and similar tectonic settings may produce igneous rocks with wide compositional ranges. Moreover, the presence of magmatic activity with a peculiar geochemical signature (e.g., calc-alkaline) does not imply a closed relation with a coeval tectonic setting (e.g., subduction tectonics). The geochemical features of igneous rocks simply reflect the geodynamic history of the area and the cratonization style of their sources.

Even considering the tool of geochemistry, the contrasting definition of mantle and crustal reservoirs from a geophysical and petrological viewpoint must be stressed.

Another topical aspect of modern igneous petrology regards the definition of asthenospheric and lithospheric mantle, as well as concepts such as mantle plumes and their role during continental basalt genesis. For example, it is now accepted that there is no single lithospheric or asthenospheric origin for CFB. Involvement of the lithosphere and asthenosphere as sources or contaminants heavily depends on the previous geodynamic evolution of the region in question, and inferences drawn for one magmatic province only represent «pin-point» conclusions, not necessarily valid in other contexts. In a continental region where the basaltic component of the lithospheric mantle is greatly depleted, the subcontinental mantle is probably not suitable to reach *solidus* in order to generate other melts. Conversely, a lithospheric mantle enriched by subduction processes and metasomatized is able to produce new batches of basaltic liquids.

Similar care must be taken with regard to experimental petrology. This is because there is no large consensus yet as regards mantle *solidus* temperature, composition of peridotite-derived partial melts, the effective role of key parameters in generating basaltic composition (e.g., alkali contents, refractory and clinopyroxene modes of the source), differences in experimental techniques, absence of interlaboratory calibration, standardization of temperature and pressure measurements).

Universally accepted petrologic rules are still a long way off.

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