Petrogenesis of Cenozoic Basalts from Vietnam: Implication for Origins of a ‘Diffuse Igneous Province’

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RECEIVED JULY 8, 1996; REVISED TYPESCRIPT ACCEPTED AUGUST 28, 1997

Basalt magmatism occurred throughout east and southeast Asia after the early Tertiary India–Asia collision. This activity does not conform to the ‘Large Igneous Province’ model in view of lower eruption and melt production rates, wide dispersal of centres and the apparent absence of deep mantle upwelling. Age data for Vietnamese plateau basalts reflect spatial-temporal patterns consistent with a rotating stress field rather than supra-hotspot lithosphere migration. For most of the volcanic centres there are two eruptive episodes: an early series formed by high-SiO₂, low-FeO* quartz and olivine tholeiites—large melt fractions of refractory (lithosphere-like) mantle—and a later series made up of low-SiO₂, high-FeO* olivine tholeiites, alkali basalts and basanites—smaller melt fractions of more fertile (asthenosphere-like) mantle. Comparison of Mg-15 normalized basalt compositions with parameterized anhydrous and hydrous experimental melt compositions allowed calculation of melt segregation pressures and temperatures. Computed for anhydrous conditions these range from <4 GPa and ~1470°C (for alkali basalts) to <0.5 GPa and ~1400°C (quartz tholeiites), and for H₂O-undersaturated conditions, from <3.5 GPa and ~1450°C to ~1.5 GPa and ~1350–1400°C, respectively. Hydrous conditions are more realistic in view of high measured basalt H₂O⁺ contents, pressure estimates consistent with melting below a thinned mechanical boundary layer (MBL) and interpolated mantle adiabats of 2–3°C/km (compared with <1°C/km for anhydrous conditions), consistent with fluid dynamic constraints and a 1440°C potential temperature. After collision-induced ‘extrusion’ of east and southeast Asia, the lithosphere was probably thinned during heating and transtension; this converted refractory MBL into a low-viscosity thermal boundary layer (TBL), and caused upwarp penetration and polybaric melting of TBL–asthenosphere columns.

KEY WORDS: Vietnam; basalt; Cenozoic; geochemistry

INTRODUCTION

Neogene–Quaternary intraplate volcanism is widespread in east and southeast Asia (Fig. 1a) forming basalt plateaux associated with pull-apart, extensional rifts (Barr & McDonald, 1981; Whitford-Stark, 1987). Although widely dispersed the activity shares common source isotopic and lithosphere structural character with intraplate and back-arc volcanism in the western Pacific and has been referred to as a ‘diffuse’ igneous province (Hoang et al., 1996). The activity post-dates the early Tertiary India–Asia collision and may be related to asthenospheric and lithospheric tectonic extrusion processes (Tapponnier et al., 1982, 1986). The province is bounded to the east and southeast by active subduction at the Izu–Bonin, Mariana and Indonesian archipelagos and to the north-west by the collision-thickened Tibet plateau (Flower et al., 1998a). The Indochina and China plates appear to have been tectonically extruded along regional strike-slip faults, with concomitant opening of the South China
Sea, Japan Sea and Andaman Sea basins (Tapponnier et al., 1982, 1986; Briais et al., 1993).

Although widely dispersed, east Asian Cenozoic basalt magmatism does not conform to the ‘Large Igneous Province’ model (Coffin & Eldholm, 1994) in view of the lower rates of eruption and melt production, dispersal of eruptive centres and apparent absence of deep mantle upwelling (Su et al., 1994). In common with other ‘diffuse’ provinces (e.g. Ormerod et al., 1988; Hoernle et al., 1995) the east Asian activity reflects contemporaneous, rapid appearance of dispersed basalt centres, a transtensional setting, and proximity to major continent–continent plate collisions. In addition, east Asian continental and western Pacific back-arc volcanics appear to share a common asthenosphere isotopic signature, sometimes referred to as the ‘southeast Asia DUPAL anomaly’ (e.g. Hickey-Vargas et al., 1995; Castillo, 1996). This has been discussed by Hoang et al. (1996) with reference to Vietnamese basalts, and by Flower et al. (1998a) with respect to east Asia in general.

The causes of dispersed, relatively sudden mantle melting events have not been extensively discussed in the literature although there has been progress in understanding the constraints of pressure, temperature, $P(H_2O)$ and mantle fertility on melt composition and volume (e.g. McKenzie & Bickle, 1988; Latin & White, 1990; Wilson, 1995). Here, we advance the idea that “diffuse” volcanic provinces may reflect the combined effects of lithosphere transtension and asthenosphere decompression concomitant with collision-induced
extrusion of thermally anomalous mantle. Assuming that the volumes, supply rates and compositions of mantle melts are simple functions of mantle composition, thermal structure and the extent of lithosphere stretching [following McKenzie & Bickle (1988) and Latin & White (1990)], we develop a petrogenetic model for the Vietnamese basalts as a basis for understanding possible asthenospheric thermal responses during the closure of eastern Tethys.

CENOZOIC MAGMATISM IN VIETNAM

Ages and eruption rates of basalts
Basalt plateaux in southern and central Vietnam often exceed 100 km in diameter, are up to several hundred metres thick, and cover a total area of ~23,000 km² (Fig. 1a) (Lacroix, 1933; Carbonnel & Saurin, 1975; Quoc & Giao, 1980; Hoang & Han, 1990; Thi, 1991; and references given by Whitford-Stark (1987)]. Most of these centres are associated with pull-apart structures comprising short extensional rifts bounded by strike-slip faults (Rangin et al., 1995). The centres appear to have involved at least two eruptive episodes, referred to here as ‘early’ and ‘late’ eruptive series, thick palaeosols marking the intervening quiescent periods (Hoang, 1996). Early episodes usually produced quartz and olivine tholeiite flows, with rare alkali basalt, whereas later episodes erupted olivine tholeiite, alkali basalt, basanite and (rarely) nephelinite. This bi-episodal pattern is recognized at the Dalat, Phuoc Long, Pleiku, Buon Ma Thuot, Xuan Loc and Re Island centres, and probably other offshore localities, although at Buon Ma Thuot the compositional trend is inverted (see below). Tholeiitic eruptives are the most voluminous, with flows up to 30 m thick erupted from axial rifts. Alkali basalts, basanites, and rare nephelineitites formed thinner, more sporadic flows erupted...
mostly from small central volcanoes aligned on conjugate strike-slip faults (Quoc & Giao, 1980; Hoang & Han, 1990; Thi, 1991; Hoang, 1996).

Barr & Macdonald (1981) reviewed new and published K-Ar age data for southeast Asian Cenozoic basalts and concluded that basalt activity in Indochina appeared by at least 12 Ma, after the cessation of South China Sea opening, and peaked in the last 3 my with an equally rapid abatement. More recent K-Ar data (Novikov et al., 1989; Arva-Sos et al., 1990; Rangin et al., 1995a) and 24 high-precision Ar-Ar dates for our stratigraphically selected core samples (Lee et al., 1998) suggest the Vietnamese centres were active over the following intervals: Dalat (17.6-7.9 Ma), Phuoc Long (straddling the border with Cambodia) (<3.4 Ma), Buon Ma Thuot (5.8-1.67 Ma), Pleiku (4.3-0.8 Ma), Xuan Loc (0.83-0.44 Ma) and the Ile des Cendres (0.8-0 Ma), and confirm the bi-episodal eruptive pattern (Fig. 1b).

Although the bulk of volcanism in Indochina post-dates South China Sea spreading, Paleogene activity has been recorded by drilling on the southern Chinese and Vietnamese continental shelves (e.g. Zhang Qi Ming, Nanhai West Co., personal communication, 1989) and appears in Thailand, NW Vietnam, Yunnan and Sichuan close to the Aila Shan Red River (ASRR) and other strike-slip shear zones (Flower et al., 1998b). Historic activity in Vietnam is confined to the offshore Con Son swell (Ile des Cendres, Fig. 1a), sporadic ash eruptions in the central highlands (e.g. Pleiku, April, 1993), and submarine activity along the eastern seaboard (Koloskov et al., 1986). Late series undersaturated lavas carry mantle xenoliths, including garnet lherzolites, spinel lherzolites and harzburgite, and megacrysts of pyroxene, olivine, plagioclase, garnet, zircon and corundum (e.g. Sapozhinkov et al., 1979; Han & Hoang, 1985; Hoang & Han, 1990). The latter characterize centres in southern Vietnam (especially Pleiku, Xuan Loc, Buon Ma Thuot and Ile des Cendres), Cambodia and Thailand (Lacombe, 1967; Barr & Macdonald, 1981), and Hainan Island, Mongxi and other southern Chinese localities (Flower et al., 1992; compare Sutherland (1983) and Irving & Frey (1984)).

Three factors allow fairly precise estimates of magmatic volumes erupted in Vietnam: first, the large number of cored hydrologic sections drilled through the basalts to basement, second, the well-documented areal extent of the eruptives, and third, their essentially basaltic character. Provisional estimates of magmatic volume for the principal centres are 1500 km$^3$ for Dalat, 2200 km$^3$ for Phuoc Long, 1500 km$^3$ for Buon Ma Thuot, 2000 km$^3$ for Pleiku and 500 km$^3$ for Xuan Loc, and for smaller centres to the north are, for example, 180 km$^3$ for Song Cau and 160 km$^3$ for Kong Plong (Fig. 1a and b). Despite a lack of age data for some smaller centres, the known age and volume relationships suggest magma production rates of up to ~2700 km$^3$/my, assuming a total of at least 8000 km$^3$ basalt were erupted at the major centres. However, such estimates underplay the effects of fractional crystallization and also ignore the presence of trapped melt consolidated at depth (see Latin et al., 1993).

**Structural and dynamic controls on volcanism**

The Indochina peninsula comprises fragments of Gondwana, made up of Precambrian, Palaeozoic and Mesozoic crust that migrated northwards during the Palaeozoic and accreted to pre-Tethyan Eurasia (e.g. Gatiński et al., 1984; Hutchison, 1989; Tung & Tri, 1992). These are preserved as distinct lithospheric sectors separated by tectonic sutures of known age (Fig. 1a and b) (e.g. Tung & Tri, 1992). A northern sector (I) comprises an accreted Archaean, Proterozoic and Palaeozoic complex in northern Vietnam and Laos, and southern China through which Cenozoic basalts were erupted at Dien Bien Phu, Phu Quy, Con Co island and Khe Sanh (Fig. 1a), and also localities in Hainan, Yunnan and Guangdong provinces (South China), and the South China Sea (e.g. Flowe et al., 1992; Tu et al., 1992). A central sector (II) comprises the Kontum Massif, a quasi-cratonic block with a 2.8 Ga core (Archaean) (Tung & Tri, 1992) surrounded by concentric ‘sub-sectors’ separated by Proterozoic, Cambrian and Permo-Triassic sutures (Fig. 1b). Cenozoic basalts were erupted within this sector at Song Cau (Archaean), Buon Ma Thuot, Pleiku and Kong Plong (Proterozoic), and at smaller centres in Quang Ngai and Re Island, and offshore north of latitude 15ºN (Cambrian) (Fig. 1a and b). A southwestern sector (III) includes the eastern part of the Khorat Plateau, which probably has a Precambrian core (in Thailand) enclosed by accretionary Palaeozoic and Mesozoic belts (in Cambodia and southwest Vietnam) (Hutchison, 1989). This sector contains the largest single basalt complex in Indochina (Phuoc Long; 200 km across and up to 500 m thick) which straddles the Cambodia Vietnam border. Smaller centres (as yet unstudied) occur in central and western Cambodia and Thailand (Barr & MacDonald, 1981; Mukasa & Zhou, 1994; Intasopa et al., 1995; Mukasa et al., 1996) (Fig. 1a). A southeastern sector (IV) probably includes most of southern Vietnam and consists of accreted Proterozoic, Palaeozoic and Mesozoic basement (Tung & Tri, 1992). With estimated crustal thicknesses of ~30 km (Tien, personal communication, 1993) this sector includes basalts of both the oldest (Dalat) and youngest (Xuan Loc) onshore complexes, and the active offshore Ile des Cendres and Katwit centres (Fig. 1a and b).

Tapponnier et al. (1982, 1986) proposed that the Indochina plate was extruded southeastwards and rotated...
PETROGRAPHY OF BASALTS AND XENOLITHS
Quartz tholeiites are dominant in Dalat and Phuoc Long and are generally aphyric (≤3% phenocrysts) to moderately phryic (≤10% phenocrysts) with plagioclase (An81-72), olivine (Fo78-74) and augite (Wo42-41En42-38Fs13-14). Most are petrographically similar to mid-ocean ridge basalt (MORB), with plagioclase preceding clinopyroxene at the low-pressure liquidus. Quartz tholeiites from early Phuoc Long and Dalat series and Dien Bien Phu, however, may contain small, unreacted phenocrysts of orthopyroxene (~En42-40Fs16-19), in which respect they resemble volcanic arc rather than MORB tholeiite (e.g. Kushiro, 1990). Orthopyroxene-phryic tholeiites are highly unusual in intraplate tectonic settings and their petrogenesis may reflect high P/H2O in the mantle source (see below). Olivine tholeiites are aphyric to sparsely phryic and are dominant at most other centres, usually interlayered with quartz tholeiite and lesser amounts of alkali basalt. Phenocrysts rarely exceed 10–15% by volume and are mostly olivine (Fo89-70) and augite (Wo44-41En42-46Fs16-18), with lesser amounts of plagioclase (An89-60). Alkali basalt and basanites are common in late series of Xuan Loc and Re Island centres, whereas alkali basalts are also prominent in the Pleiku late series and Buon Ma Thuot early series. These are moderately phryic, with 7–15% olivine phenocrysts (Fo89-70), several generations of which may be distinguished on the basis of morphology and composition, together with lesser amounts of augite.

Mantle xenoliths include garnet lherzolite, spinel lherzolite and harzburgite, along with eclogite of unknown provenance and cumulate xenoliths comprising wustite, websterite and pyroxenite. Megacrysts include olivine, Al-rich clinopyroxene, orthopyroxene, Ti-amphibole, anorthoclase, phlogopite, sapphire and zircon (see Flower et al., 1992). Spinel lherzolites are abundant at Phu Quy and Con Co island (sector I), Quang Ngai, Re Island, Buon Ma Thuot and Pleiku (sector II), Dalat, Xuan Loc and Ìle des Cendres (sector IV) along with cumulates and megacrysts. Eclogites were found in Pleiku basalts and comprise idiomorphic garnet (40%) and euhedral to subhedral clinopyroxene (60%), with or without orthopyroxene (see Halton & Gurney, 1987). Feldspar megacrysts were also found in Pleiku and range from sanidine to anorthoclase in composition. Clinopyroxene, orthopyroxene, and Ti-amphibole megacrysts were encountered mostly in Pleiku and Dalat centres, often with amphibole-free lherzolites, and are common in Ìle des Cendres basalts.

GEOCHEMISTRY OF BASALTS AND XENOLITHS
Geochemical data for basalts and xenoliths are used to define compositions and stratigraphic relations of the erupted magmas, identify compositions of their unfractonated (mantle-equilibrated) parent melts, and from
this basis estimate melt segregation conditions and thermal state of the asthenosphere. Assuming the entrained lherzolites and harzburgites represent lithospheric mantle fragments and that high-pressure undersaturated melts are probably generated in the asthenosphere, estimated melt segregation conditions help define the ‘thermal boundary layer’ (TBL) between convecting asthenosphere and a rigid ‘mechanical boundary layer’ (MBL). Sr, Nd and Pb isotopic data for representative basalts—summarized below and discussed in detail by Hoang et al. (1996)—confirm the presence of anomalous (DUPAL-like) asthenosphere beneath Indochina and record evidence for the interaction of asthenospheric melts with enriched lithospheric mantle and continental crust. Thermobarometric estimates of xenolith equilibration, a basis for estimating conductive geotherms in the lithospheric mantle, are being published separately along with chronologic interpretations of xenolith Re-Os and Sm-Nd isotopic decay systematics (Hoang et al., 1998).

**Sampling and analysis**

Most samples analysed were selected from fresh, unweathered drill-core from the Dalat, Phuoc Long, Buon Ma Thuot, Pleiku and Xuan Loc plateaux, surface outcrops in these and other basalt plateaux, and dredge hauls from the South China Sea. Stratigraphic sections were developed from hydrologic drill records of the Vietnamese Geological Survey and are accessible with geochemical data at the *Journal of Petrology* website (http://www.oup.co.uk/jnls/list/petroj). Dredge samples were collected by Vietnamese and Soviet scientists during cruises of the R.V. *Vulkanolog* between 1981 and 1987 (Koloskov et al., 1986) (Fig. 1a and b).

Representative major and trace element analyses were published with isotopic data by Hoang et al. (1996) together with descriptions, precisions and accuracies of the analytical techniques employed. Whole-rock major and trace elements were determined using X-ray fluorescence spectrometry (Michigan State University) and instrumental neutron activation analysis (University of Michigan), H2O+ and CO2 using a CHN analyser (Argonne National Laboratory), and phenocryst compositions using a scanning electron microscope with energy-dispersive attachment (UIC).

**Major elements**

CIPW normative variation defines a range of quartz tholeiite (QT) (qz-normative), olivine tholeiite (OT) (ol + hy-normative), alkali basalt (AB) (ne up to 5%) and basanite (BA) (ne-normative >5%) at most centres, reflecting an overall increase in undersaturation from older to younger centres (Fig. 2). The Phuoc Long and Dalat plateaux are dominated by quartz tholeiite with subsidiary olivine tholeiite, basalts from Pleiku and Buon Ma Thuot include quartz and olivine tholeiites and alkali basalts in approximately equal amounts, and Xuan Loc and offshore seamounts comprise olivine tholeiite and alkali basalt (Fig. 2).

‘Chemical types’ were established as a basis for stratigraphic correlation between drill sites (Hoang, 1996) and for comparing primitive magma types within and between centres (this work). Where possible, chemical types were taken to include stratigraphic intervals of similar composition—cooling unit batches identified from drill core stratigraphy, or coherent groupings identified from surface sampling of mapped flows—whose internal variation is consistent with crystal-liquid redistribution or mixing. Distinctions were based on major element oxides and CIPW normative character, although surface sample groups were further constrained by element ratios such as K/Na, Rh/Sr and Ba/Zr. This approach is exclusive rather than inclusive such that chemical types identified at different sites may represent different parts of the same erupted magma batch (e.g. types XL-B and -D at Xuan Loc; types DT-A, -B and -C at Dalat) whereas a single type is unlikely to include compositions from distinct eruptive episodes. Type averages are given in Table 1, with representative incompatible element ratios, ages (where known) and melt segregation pressures and temperatures—the last computed from simulated primitive magma compositions (see below).

Figure 3a–f shows chemical type variation in plots of SiO2 and TiO2 vs MgO (wt %) annotated for age where possible, with early and late series outlined by continuous and dashed lines, respectively. Dalat centre basalts range between 6 and 9 wt % MgO and are exclusively tholeiitic (Fig. 3a), including four chemical types from drill sites 711, 736 and 756. Early series eruptives (10-5-14-0 Ma) include low-Ti olivine and quartz tholeiite types DT-A and -B, whereas late series (1-8-2-6 and <1-8-0 Ma) comprise low- and high-Ti olivine tholeiite types DT-C and -D. Basalts from Phuoc Long section 804 also range between 6 and 9 wt % MgO and show a quartz tholeiite type (PH-A) and two olivine tholeiite types (PH-B and -C), types PH-A and -B forming a 15-15-1 Ma early series and PH-C a late series of ~5 Ma (Fig. 3a). Pleiku basalts are more variable, ranging between 5 and 12 wt % MgO and showing a broad range of SiO2 and TiO2 contents (Fig. 3b). Early series eruptives (4-8-2-6 Ma) include quartz tholeiite (chemical types PL-A and -B), olivine tholeiite (PL-D and -E) and low-Ti alkali basalt (PL-F), whereas the late series type PL-C (~2-5 Ma), a high-Ti alkali basalt, succeeds type A at site 90. The early series types appear to comprise two groups, with 5-8 wt % MgO (types PL-A and -B) and 8-12 wt % MgO (PL-C, -D and -F), types PL-F and -B, and PL-D and -A representing possible cogenetic pairs. Type
Fig. 2. CIPW normative variation for analysed basalts from early and late eruptive series of Phuoc Long (SW sector), Pleiku, Buon Ma Thuot, Song Cau, Quang Ngai and Re Island (central sector), Xuan Loc, Dalat and Île des Cendres (SE sector), and Dien Bien Phu, Khe Sanh and Con Co island (northern centres) (Appendices A and B: http://www.oup.co.uk/jnls/list/petroj) (computed for Fe$^{3+} = 0.15$Fe$_{total}$), showing the range between quartz tholeiite, olivine tholeiite, alkali basalt and basanite.

PL-C, the youngest of the early series (2.6 Ma), resembles the late series type PL-E in SiO$_2$ but not TiO$_2$ content and appears to be unique. Buon Ma Thuot shows a similar compositional range, including quartz tholeiite (type BMT-B), olivine tholeiite (BMT-A and -C) and alkali basalt (BMT-D, -E and -F) (Fig. 3b). However, in contrast to the other centres, SiO$_2$-undersaturated eruptives are confined to the early BMT series (4.6±3.2 Ma) with late series made up exclusively of quartz tholeiite (1.9±0.3 Ma). Although Buon Ma Thuot chemical types were identified from several sections, this ‘inverted’ sequence is best seen at site 45, where quartz tholeiite (type BMT-B) overlies olivine tholeiite (BMT-A) and alkali basalt (BMT-D).

Xuan Loc early and late series are distinguished at site 507, where olivine and quartz tholeiites (types XL-A and -C) (2.4±2.2 Ma) precede basanites (types XL-B and -D) (1.1±0.4 Ma) (Fig. 3c). Offshore islands and seamounts show similar associations of tholeiite and alkali basalt although stratigraphic control is poor or lacking (Fig. 3c). An early tholeiite series (type RE-A) (12.0 Ma) and late alkali basalt series (type RE-B) (1.2±0.4 Ma) are recognized on Re Island, and analogous distinctions exist for Île des Cendres (IC-A and -B) and Katwit Island seamounts. Thus, with the important exception of Buon Ma Thuot, early eruptive series comprise quartz and olivine tholeiite with only rare alkali basalt (e.g. type PL-alkali basalt (BMT-D)). Although Buon Ma Thuot chemical types were identified from several sections, this ‘inverted’ sequence is best seen at site 45, where quartz tholeiite (type BMT-B) overlies olivine tholeiite (BMT-A) and alkali basalt (BMT-D).

In general, the alkali basalts show higher contents of TiO$_2$ and FeO$^*$ than quartz and olivine tholeiites at equivalent MgO contents, such that a rough inverse correlation exists between SiO$_2$, undersaturation and TiO$_2$ or FeO$^*$. Stratigraphic successions from low- to high-TiO$_2$ basalts are common in intraplate flood basalts and have been interpreted to reflect variable source rather than fractional crystallization or contamination effects (e.g. Hawkesworth et al., 1988). However, as noted, this pattern is not ubiquitous in Vietnam, as both Pleiku and
Table 1: Chemical Type averages for Vietnamese basalts

| Magma series: | DT-I | DT-I | DT-I | DT-II | DT-II | DT-II | PH-I | PH-I | PH-II | PL-I | PL-I |
|--------------|------|------|------|-------|-------|-------|------|------|-------|------|------|
| Chem. type:  | DT-A/711 | DT-B/736 | DT-C/756 | DT-D/736 | DT-D/756 | DT-D/711 | PH-A/804 | PH-B/804 | PH-C/804 | PL-A/121 | PL-B/90 |
| Melt type:   | OT   | OT   | OT   | OT    | OT    | QT    | OT   | OT   | OT    | OT   | OT   |
| SiO₂         | 52.05 | 52.55 | 53.85 | 48.05 | 52.01 | 51.42 | 52.72 | 50.70 | 48.84 | 53.99 | 53.12 |
| TiO₂         | 1.86  | 1.88  | 2.04  | 2.48  | 2.04  | 2.33  | 1.80  | 2.16  | 2.36  | 1.59  | 1.78  |
| Al₂O₃        | 14.24 | 14.93 | 14.89 | 15.29 | 14.38 | 15.65 | 15.57 | 15.28 | 14.22 | 14.55 | 14.56 |
| FeO⁺         | 11.14 | 10.87 | 9.69  | 11.70 | 10.04 | 10.02 | 10.85 | 11.06 | 11.09 | 10.02 | 10.22 |
| MnO          | 0.18  | 0.16  | 0.16  | 0.19  | 0.14  | 0.23  | 0.18  | 0.16  | 0.15  | 0.15  | 0.16  |
| MgO          | 7.92  | 7.04  | 6.17  | 8.19  | 8.01  | 6.45  | 6.28  | 7.35  | 10.26 | 6.72  | 6.89  |
| CaO          | 9.12  | 8.40  | 9.08  | 8.72  | 8.57  | 7.49  | 9.07  | 8.64  | 8.47  | 8.71  | 8.79  |
| Na₂O         | 2.66  | 3.20  | 3.04  | 2.70  | 3.04  | 3.49  | 2.87  | 3.15  | 2.56  | 3.00  | 2.96  |
| K₂O          | 0.59  | 0.67  | 0.80  | 2.14  | 1.40  | 2.47  | 0.43  | 1.11  | 1.50  | 0.83  | 1.24  |
| P₂O₅         | 0.23  | 0.30  | 0.27  | 0.53  | 0.36  | 0.45  | 0.24  | 0.40  | 0.54  | 0.28  | 0.27  |
| K₂O/Na₂O     | 0.227 | 0.21  | 0.263 | 0.793 | 0.461 | 0.708 | 0.15  | 0.352 | 0.585 | 0.277 | 0.419 |
| Zr/Y         | 4.2   | 6.1   | 3.7   | 4.7   | 5     | 5.6   | 5     | 6.8   | 7.8   | 4.7   | 6.7   |
| Zr/Nb        | 185.1 | 193.2 | 138.6 | 101   | 99.1  | 272.2 | 502.1 | 317.1 | 233.2 | 168.8 | 221.2 |
| Rb/Sr        | 0.03  | 0.04  | 0.05  | 0.11  | 0.12  | 0.06  | 0.03  | 0.05  | 0.06  | 0.06  | 0.03  |
| Ba/Zr        | 1.91  | 1.82  | 2.31  | 3.3   | 3.46  | 2.74  | 0.61  | 1.32  | 2.3   | 1.71  | 2.65  |
| n            | 4     | 15    | 6     | 3     | 1     | 2     | 8     | 6     | 4     | 6     | 3     |
| Age (Ma)     | >10.5 | 10.5–14.0 | 18–26 | <1.8  | <1.8  | <1.8  | 15.1  | 15.5  | 7.0   | 3.9   | 2.6   |
| T°C (dry)    | 1427  | 1419  | 1378  | 1447  | 1386  | 1390  | 1418  | 1425  | 1442  | 1401  | 1421  |
| T°C (H₂O)    | 1417  | 1404  | 1345  | 1446  | 1357  | 1363  | 1403  | 1414  | 1439  | 1379  | 1408  |
| P (dry)      | 11.5  | 9.8   | 6.3   | 23.9  | 11.5  | 14.5  | 10.5  | 15.7  | 19.6  | 7.9   | 15.6  |
| P (H₂O)      | 17.9  | 16.9  | 14.8  | 25.2  | 17.9  | 19.7  | 17.3  | 20.4  | 22.7  | 15.8  | 20.3  |
| Magma series: | PL-I | PL-I | PL-I | PL-I | PL-II | BMT-I | BMT-I | BMT-II | BMT-II | BMT-II | BMT-II | BMT-I |
|--------------|------|------|------|------|-------|-------|-------|--------|--------|--------|--------|-------|
| Chem. type:  | PL-C/90 | PL-D/90 | PL-E/90 | PL-F/121 | PL-G/121 | BMT-A/63 | BMT-A/45 | BMT-B/911 | BMT-B/45 | BMT-B/858 | BMT-C/45 | |
| Melt type:   | QT   | AB   | OT   | OT   | AB    | OT    | OT    | OT     | OT     | OT     | OT     | OT    |
| SiO2         | 52.04 | 45.61 | 49.75 | 49.00 | 48.69  | 48.33  | 48.82  | 52.38   | 52.22   | 51.75   | 48.77 |
| TiO2         | 1.93  | 2.07  | 1.96  | 2.58  | 2.44   | 2.18   | 2.06   | 1.65    | 1.58    | 1.62    | 1.88  |
| Al₂O₃        | 15.19 | 13.36 | 13.99 | 14.55 | 13.52  | 14.04  | 14.35  | 15.03   | 15.22   | 15.38   | 13.99 |
| FeO*         | 10.34 | 13.16 | 11.49 | 11.75 | 11.33  | 11.51  | 11.29  | 10.04   | 10.15   | 10.12   | 11.64 |
| MnO          | 0.16  | 0.23  | 0.18  | 0.17  | 0.17   | 0.13   | 0.16   | 0.15    | 0.14    | 0.16    | 0.12  |
| MgO          | 6.44  | 10.21 | 10.00 | 7.76  | 9.76   | 10.05  | 8.68   | 7.09    | 6.99    | 7.46    | 9.89  |
| CaO          | 9.11  | 10.13 | 8.19  | 8.45  | 8.64   | 9.18   | 9.82   | 9.42    | 9.22    | 8.72    | 9.24  |
| Na₂O         | 3.14  | 3.01  | 2.74  | 2.99  | 2.75   | 2.63   | 2.83   | 2.97    | 3.08    | 3.15    | 2.54  |
| K₂O          | 1.33  | 1.40  | 1.28  | 2.00  | 2.04   | 1.45   | 1.43   | 0.95    | 1.09    | 1.28    | 1.56  |
| P₂O₅         | 0.27  | 0.74  | 0.43  | 0.59  | 0.93   | 0.50   | 0.55   | 0.31    | 0.31    | 0.37    | 0.35  |
| K₂O/Na₂O     | 0.424 | 0.465 | 0.467 | 0.669 | 0.741  | 0.551  | 0.505  | 0.320   | 0.354   | 0.406   | 0.614 |
| Zr/Y         | 6.5   | 6.5   | 8.7   | 7.7   | 9      | 7.1    | 6.2    | 4.2     | 4.8     | 5       | 5.1   |
| Zr/Nb        | 170.9 | 118.4 | 260.2 | 237   | 268.5  | 153    | 137.2  | 101.8   | 118.6   | 107.1   | 121   |
| Rb/Sr        | 0.02  | 0.07  | 0.06  | 0.07  | 0.05   | 0.06   | 0.06   | 0.06    | 0.06    | 0.07    | 0.06  |
| Ba/Zr        | 2.74  | 4.45  | 2.44  | 2.59  | 3.98   | 3.24   | 5.4    | 3.54    | 3.46    | 3.85    | 3.8   |
| n            | 5     | 4     | 7     | 7     | 8      | 1      | 3      | 7       | 4       |         |       |
| Age (Ma)     | ?     | 3.4   | ?     | <2.5  | 4.8    | 4.02   | 4.02   | 1.9-0.3 | 1.9-0.3 | 1.9-0.3 | 3.2   |
| T°C (dry)    | 1402  | 1435  | 1419  | 1439  | 1444   | 1436   | 1431   | 1389    | 1392    | 1391    | 1442  |
| T°C (H₂O)    | 1380  | 1428  | 1405  | 1434  | 1441   | 1430   | 1422   | 1361    | 1366    | 1363    | 1439  |
| P (dry)      | 11.1  | 20.1  | 17.2  | 20    | 22.9   | 20.4   | 20.5   | 10.4    | 11.2    | 12.4    | 20.4  |
| P (H₂O)      | 17.7  | 23    | 21.3  | 24.6  | 21.8   | 23.1   | 23.2   | 17.3    | 17.7    | 18.6    | 23.1  |
### Table 1: continued

| Magma series: | BMT-I | BMT-I | BMT-I | BMT-I | BMT-I | XL-I | XL-I | XL-II | XL-I | XL-I |
|---------------|-------|-------|-------|-------|-------|------|------|-------|------|------|
| Chem. type:   | BMT-C/63 | BMT-D/45 | BMT-D/63 | BMTD/658 | BMT-E/900 | XL-A/507 | XL-A/508 | XL-B/511 | XL-C/507 | XL-C/516 |
| Melt type:    | OT | AB | AB | AB | AB | OT | OT | BA | QT | OT |
| SiO$_2$       | 50.73 | 45.90 | 47.44 | 46.61 | 48.43 | 49.88 | 50.31 | 50.07 | 43.40 | 49.54 | 49.32 |
| TiO$_2$       | 1.80 | 1.99 | 2.19 | 2.00 | 2.40 | 2.69 | 2.19 | 2.17 | 3.04 | 2.27 | 2.22 |
| Al$_2$O$_3$   | 15.02 | 14.62 | 13.64 | 14.54 | 15.40 | 15.08 | 13.99 | 14.01 | 12.93 | 13.77 | 13.45 |
| FeO*          | 10.94 | 11.81 | 12.49 | 11.40 | 11.93 | 10.69 | 11.29 | 11.06 | 12.64 | 11.43 | 10.83 |
| MnO           | 0.21 | 0.19 | 0.26 | 0.18 | 0.18 | 0.16 | 0.21 | 0.16 | 0.24 | 0.24 | 0.17 |
| MgO           | 7.57 | 9.74 | 9.06 | 9.98 | 6.92 | 7.38 | 8.94 | 9.14 | 11.05 | 9.34 | 9.95 |
| CaO           | 9.31 | 10.31 | 9.93 | 10.03 | 8.54 | 8.37 | 8.51 | 8.57 | 12.28 | 9.12 | 9.30 |
| Na$_2$O       | 2.94 | 2.81 | 2.57 | 2.89 | 2.97 | 2.94 | 2.84 | 2.84 | 2.45 | 2.87 | 2.77 |
| K$_2$O        | 1.19 | 2.12 | 1.76 | 2.57 | 2.31 | 2.18 | 1.46 | 1.54 | 0.71 | 0.97 | 1.54 |
| P$_2$O$_5$    | 0.30 | 0.61 | 0.65 | 0.80 | 0.91 | 0.64 | 0.25 | 0.43 | 1.25 | 0.45 | 0.44 |
| K$_2$O/Na$_2$O| 0.405 | 0.754 | 0.626 | 0.889 | 0.786 | 0.741 | 0.514 | 0.542 | 0.230 | 0.338 | 0.556 |
| Zr/Y          | 5.3 | 5.7 | 5.8 | 6.2 | 9.4 | 11.3 | 7.9 | 6.9 | 9.4 | 9.2 | 6.2 |
| Zr/Nb         | 132.9 | 153.6 | 160 | 168 | 424.4 | 237.8 | 204.1 | 191.5 | 166.7 | 255 | 167.6 |
| Rb/Sr         | 0.06 | 0.07 | 0.06 | 0.06 | 0.08 | 0.07 | 0.1 | 0.07 | 0.1 | 0.11 | 0.05 |
| Ba/Zr         | 3.28 | 4.74 | 4.71 | 5.4 | 2.0 | 2.9 | 2.11 | 2.29 | 5.82 | 2.35 | 3.25 |
| n             | 6 | 4 | 4 | 2 | 3 | 2 | 2 | 4 | 3 | 3 | 2 |
| Age (Ma)      | 3.2 | 4.6 | 4.6 | 4.6 | 4.5 | >4.6 | >2.4 | >2.4 | >1.1 | 2.2-2.42 | 2.2-2.42 |
| T°C (dry)     | 1418 | 1445 | 1449 | 1431 | 1456 | 1432 | 1431 | 1418 | 1480 | 1433 | 1428 |
| T°C (H$_2$O)  | 1404 | 1443 | 1449 | 1423 | 1459 | 1424 | 1422 | 1403 | 1495 | 1426 | 1419 |
| P (dry)       | 15.9 | 25.7 | 30.9 | 26.6 | 23.8 | 18.8 | 15.3 | 16 | 38.8 | 18.2 | 18.2 |
| P (H$_2$O)    | 20.5 | 28.3 | 29.3 | 26.8 | 25.2 | 22.2 | 20.6 | 34 | 33.8 | 21.8 | 20.1 |
## Cenozoic Basalts from Vietnam

**Magma series:** XL-II, XL-II I. Cendres I. Cendres Katwit Is. Re Is.-II Re Is.-I QN Con Co Khe Sanh DBP

**Chem. type:** XL-D/511 XL-D/507 IC-A IC-B RE-A RE-B

**Melt type:** BA AB AB OT OT OT OT OT OT OT

| Component | XL-II | XL-II | I. Cendres | I. Cendres | Katwit Is. | Re Is.-II | Re Is.-I | QN | Con Co | Khe Sanh | DBP |
|-----------|-------|-------|------------|------------|------------|-----------|----------|----|--------|----------|-----|
| SiO₂      | 43.26 | 45.01 | 49.50      | 50.40      | 49.44      | 49.87     | 52.99    | 44.77 | 52.67  | 48.48    | 55.26 |
| TiO₂      | 2.76  | 2.93  | 2.34       | 1.99       | 2.07       | 2.29      | 1.57     | 2.85  | 1.78   | 2.68     | 2.04 |
| Al₂O₃     | 12.28 | 13.84 | 13.51      | 13.91      | 13.87      | 15.32     | 15.01    | 14.17 | 16.50  | 14.61    | 16.00 |
| FeO⁺      | 12.15 | 12.78 | 11.67      | 11.52      | 11.48      | 10.93     | 9.43     | 11.51 | 10.22  | 10.83    | 9.66 |
| MnO       | 0.20  | 0.15  | 0.16       | 0.15       | 0.16       | 0.14      | 0.14     | 0.22  | 0.12   | 0.15     | 0.16 |
| MgO       | 12.96 | 9.22  | 7.86       | 8.18       | 8.50       | 7.10      | 7.46     | 11.21 | 4.17   | 8.62     | 5.36 |
| CaO       | 10.99 | 9.86  | 9.00       | 8.87       | 9.13       | 8.48      | 8.90     | 10.12 | 5.67   | 9.45     | 6.15 |
| Na₂O      | 2.87  | 2.76  | 3.27       | 3.12       | 3.13       | 3.11      | 3.28     | 3.22  | 4.92   | 3.09     | 3.00 |
| K₂O       | 1.32  | 2.48  | 2.27       | 1.49       | 1.83       | 2.19      | 0.92     | 0.94  | 3.22   | 1.54     | 2.04 |
| P₂O₅      | 0.94  | 0.96  | 0.42       | 0.36       | 0.38       | 0.57      | 0.29     | 1.00  | 0.72   | 0.54     | 0.32 |

Averages of dry-weight normalized analyses, annotated as follows: QT, quartz tholeiite; OT, olivine tholeiite; AB, alkali basalt; BA, basanite. I, 'early series'; II, 'late series' eruption episodes. No stratigraphic distinction is observed for Île des Cendres seamounts, although low- and high-Ti groups are recognized. $P-T$ calculations are for Mg-15 normalized chemical type averages, explained in text. Basalt plateaux: DT, Dalat; PG, Phuoc Long; BMT, Buon Ma Thuo; PL, Pleiku; XL, Xuan Loc; QN, Quang Ngai; DBP, Dien Bien Phu. $^{39}$Ar-$^{39}$Ar age data on these samples from Lee et al. (1998). Chemical types named alphabetically with respect to drill site number. FeO⁺, total Fe oxide.
Fig. 3.

Buon Ma Thuot high-Ti alkali basalts precede lower-Ti variants in the sequence. The existence of both coupled and decoupled variation between TiO$_2$ and SiO$_2$ may result from the interplay of variable melt fraction, source fertility and $P$($H_2$O) in generating primitive melts. We suggest therefore that parental magmas reflect a spectrum between large melt fractions, low pressures and a (low-Ti, Fe) refractory source (on the one hand), and small melt fractions, high pressures and a (high-Ti, Fe), fertile source (on the other).

Clinopyroxene separates from representative spinel lherzolites and harzburgites are in the range En$_{0.6-4.5}$ Fs$_{94.5-1}$Wo$_{0.4-4.7}$, with mg-numbers [100 × Mg/(Mg + Fe$^{2+}$)] (80–95), Al$_2$O$_3$ (5·6–7·9 wt %) and CaO (19·7–22·2 wt %) overlapping with those of peridotites from mid-ocean ridge (e.g. McDonough & Frey, 1989; Johnson et al., 1990) and continental settings (e.g. Cao & Zhou, 1987; Hawkesworth et al., 1990; Qi et al., 1995). Cr$_2$O$_3$ contents range from 1 to 0·7 wt %, with cr-numbers [Cr$_2$O$_3$ × 100/(Cr$_2$O$_3$ + Al$_2$O$_3$)] in the range 1·2–10·38. The mg-number shows a positive covariance with SiO$_2$ and Cr$_2$O$_3$, and a negative covariance with Al$_2$O$_3$, consistent with control by partial melting (Fig. 4), whereas Fo contents of coexisting olivines are in the range 90·1–95·3. These data suggest that Indochinese lithospheric mantle is variably refractory and similar to Phanerozoic subcontinental mantle elsewhere (e.g. Song & Frey, 1989; Hawkesworth et al., 1990).

**Trace elements and isotopes**

Isotopic and trace element data for representative lava samples (Hoang et al., 1996) are reviewed here as a basis for evaluating mantle boundary layer models derived from the major element variation (see below). As-similation-fractional crystallization appears to have affected several flows in Pleiku (olivine tholeiite, type PL-A), Dalat (quartz tholeiite, DT-A), Buon Ma Thuot (quartz tholeiite, BMT-B), Xuan Loc (tholeiitic andesite, XL-A) and Dien Bien Phi (orthopyroxene-bearing quartz tholeiite), which show negative correlations of MgO/FeO* and $^{87}$Sr/$^{86}$Sr and strong enrichment in EM2 [Hoang et al., 1996; compare DePaolo (1981) and Arndt et al. (1993)]. Involvement of crustal components is also indicated by positive Rb and Ba anomalies in mantle-normalized incompatible element distributions (Fig. 5).
Fig. 3.

(e.g. in Pleiku, Buon Ma Thuot and Xuan Loc patterns) and in plots of Sr/Zr, Rb/Zr and Ba/Zr vs Ti/Zr (Fig. 6). However, covariance of MgO/FeO and Sr$^{87}$/Sr$^{86}$ (Fig. 7) suggests some of these reflect partial melting of an enriched source, generated by sediment-derived input. Covariance of MgO/FeO and 87Sr/86Sr (Fig. 7) and -II and, to a lesser extent, PL-I and XL-II series show marked increases in Sr/Zr consistent with crustal addition, whether resulting from wallrock reaction or mantle metasomatism, rather than wallrock contamination of ascending melt.

Ti/Zr typically decreases from early to late series eruptions (except in Buon Ma Thuot) and is generally matched by increasing ratios of more- to less-incompatible elements (e.g. Fig. 6a–c). In addition to source heterogeneities these arrays reflect a decrease in melt fraction from tholeiite to alkali basalt (indicated by dashed parallel lines with decreasing Ti/Zr) and contrast with divergent trends towards higher Ti/Zr, reflecting the addition of crustal components (CC). For example, negative covariation bands of Rb/Zr and Ba/Zr vs Ti/Zr include members of both early and late series at each centre (Fig. 6a and b), whereas sharp increases in Rb/Zr and Ba/Zr with increasing or near-constant Ti/Zr (shown by DT-I, BMT-I, BMT-II and XL-II series) (Fig. 6a and b) are consistent with the addition of crustal material. In contrast, again with the exception of Buon Ma Thuot, Sr/Zr ratios show little overall variation with Ti/Zr and lie mostly between two and four. Whereas variation is slight to negligible in Dalat and Phuoc Long, BMT-I and -II and, to a lesser extent, PL-I and XL-II series show marked increases in Sr/Zr consistent with crustal input.

Crust addition, whether resulting from wallrock reaction or mantle metasomatism, is thus significant at Buon Ma Thuot and Xuan Loc, and is further supported by the presence of isotopic EM2 (Hoang et al., 1996). However, two points should be noted. The presence of systematic isotopic differences between basalts from distinct lithospheric sectors suggests that the effects of enriched lithospheric mantle predominate over wallrock reaction. Second, at most centres there is a secular change from EM2-rich early series to EM1-rich late series compositions. Hoang et al. (1996) proposed that, irrespective of whether EM2 is incorporated by rising melt or derives from lithospheric mantle metasomes, EM1-like components must have been present in the convecting asthenosphere and were perhaps delaminated from the Sino-Korean craton.

As samples of sub-Indochina lithospheric mantle, the spinel lherzolite xenoliths show low Rb/Sr (0·001–0·006) and high Sm/Nd (0·3–0·65) values, with Sr$^{87}$/Sr$^{86}$ and
Fig. 3. Plots of SiO$_2$ and TiO$_2$ vs MgO (wt %) showing ‘chemical types’ defined in the text for cored sections of Vietnamese basalt plateaux (average compositions in Table 1). These are labelled alphabetically with drill site numbers in parentheses (shown in Fig. 1b) and age ranges based where possible on Ar-Ar data [Lee et al., 1998]. ‘Early series’ types are shown by continuous outlines, ‘late series’ by dashed outlines. Tholeites are shown by fine lines and alkali basalts and basanites by bold lines.

$^{143}$Nd/$^{144}$Nd ratios corresponding to N-MORB ± EM2 (Hoang et al., 1998). Although these differ from the EM1-rich host alkali basalts, the (decoupled) negative covariance of Sm/Nd and mg-number suggests cryptic EM2-rich metasomatism of the refractory lithospheric >7.5 MgO wt % which shows negligible effects of mantle (Fig. 8) (see Frey & Green, 1974).

PRIMITIVE MELT COMPOSITIONS
To interpret basaltic melt segregation conditions, we have considered the compositions of primitive, mantle-equilibrated, melts to be a simple function of lithospheric stretching ($\beta$) and asthenosphere potential temperature ($T_p$), in turn reflecting melt segregation pressure, melt fraction and bulk source composition (after McKenzie & Bickle, 1988; Latin & White, 1990). By comparison with published experimental results, estimating melt segregation pressures and temperatures.

Variation of Mg-15 normalized eruptive compositions
Primitive melt compositions with assumed 15 wt % MgO were simulated by adding forsteritic olivine to averages of chemical types based on samples with >7.5 MgO wt %. The rationale for this procedure (following Scarrow & Cox, 1995) was that: (1) olivine (Fo$_{85-90}$) is the only significant phenocryst phase in basalts with >7.5 MgO wt % which shows negligible effects of clinopyroxene and plagioclase fractionation, (2) parent magmas are almost certainly more magnesian than the most Mg-rich erupted lavas, (3) liquids with >13 wt % MgO are likely to have equilibrated with Fo$_{65-90}$ (Roeder & Emslie, 1970), and (4) experimental peridotite melts fall in the range 12-17 wt % MgO. These criteria apply to both anhydrous and hydrous conditions (Kushiro, 1990; Hirose & Kushiro, 1993; Hirose & Kawamoto, 1995). The effect of adding olivine is relatively trivial for most oxides but provides a basis for establishing the variation of primitive melts, and by comparison with published experimental results, estimating melt segregation pressures and temperatures. Olivine (Fo$_{30}$) was added to the eruptive (i.e. chemical type) averages until MgO = 15 wt %, assuming $K_{\text{Olivine}}$ = 0.30 (Roeder & Emslie, 1970), according to the equations.

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Fig. 4. Plots of $mg$-number [$Mg \times 100/(Mg + Fe^*)$] vs oxides wt % for clinopyroxenes separated from spinel lherzolites [data published by Hoang et al. (1998)]. $\bigcirc$: northern and central sectors; $\triangle$: Dalat; open crosses: Île des Cendres.

\[ \log\left(\frac{FeO}{Al_2O_3}\right) = K \log\left(\frac{MgO}{Al_2O_3}\right) + \log \left(\frac{FeO}{MgO}\right)_p \left(\frac{Al_2O_3}{MgO}\right)_p^{K-1} \]  

\[ \langle Al_2O_3/FeO \rangle_i = K \langle MgO/FeO \rangle_p \langle Al_2O_3/MgO \rangle_p + \langle Al_2O_3/FeO \rangle_p (1 - K) \]  

where $p$ denotes the primary melt composition, $i$ denotes the initial (i.e. erupted) liquid composition, and $K$ is the distribution coefficient $K_{Mg/Fe}^{liq/ol}$ (from Pearce, 1978).

Figure 9a-d shows Mg-15-normalized variation from which three distinctive features are evident. First, discrete ‘suites’ of chemical type averages are recognized in terms of FeO*, SiO2, TiO2, K2O and P2O5 at most of the basalt centres. Second, several of these range from relatively SiO2-rich (quartz or olivine tholeiite) to SiO2-poor (olivine tholeiite, alkali basalt or basanite) end-members, replicating the overall pattern of regional variation. Third, at several centres (e.g. Dalat, Phuoc Long, Pleiku, Xuan Loc and Re Island) the low- and high-FeO* suites correspond, respectively, to early and late eruptive series. Whereas the covariation of Si-15 with Fe-15 is negative, that of Ti-15 and P-15 is mostly positive, and K-15 appears to vary independently of Fe-15 (Fig. 9a-d). Covariance of Si-15 and K-15 (not shown) is mostly negative within magmatic suites, except for the Xuan Loc and Pleiku upper series basalts, which show a positive
Fig. 5. Primitive mantle-normalized incompatible element distributions for representative basalts from: (a) Phuoc Long (SW sector), (b) Pleiku, (c) Buon Ma Thuot, (d) Quang Ngai and Re Island (central sector), (e) Xuan Loc and Dalat (SE sector), and (f) Dien Bien Phu and Con Co island (northern centres), and Ile des Cendres (SE sector) (Appendix B). Normalizing data are from Hofmann (1988). (Note the negative Ba anomaly in Phuoc Long, high Rb anomaly in Xuan Loc, and positive Sr anomaly in Buon Ma Thuot basalts.)

deflection consistent with isotopic and trace element indications for wallrock reaction (Hoang et al., 1996).

A simple model was developed as a basis for petrogenetic interpretation in terms of relevant published experimental data (in this paper), and Sr, Nd and Pb isotopic variation (Hoang et al., 1996). The within-suite variation of SiO₂ saturation and incompatible element contents represents a spectrum of melts generated from a discrete volume of decompressing mantle, reflecting minor isotopic and trace element source heterogeneities, and the combined effects of variable melt fraction, segregation pressure and temperature (Langmuir et al., 1992; Scarrow & Cox, 1995). In contrast, the between-suite differences in Fe-15 were attributed to differences in mantle fertility as reflected by the bulk source mg-number, and contents of CaO, Al₂O₃ and Na₂O. These contrasting vectors are analogous to those recognized in oceanic (Wilkinson, 1991; Nicolson & Latin, 1992) and continental (Turner & Hawkesworth, 1995) intraplate basalts.

Source fertility and melt segregation conditions

Establishing pressure and temperature conditions of primitive melt segregation, although problematic, can be approached in at least two ways. The first involves
comparing natural melt compositions with those produced experimentally at known pressures, temperatures and $P(\text{H}_2\text{O})$, from a source of known composition, assuming experimental melt compositions are realistic, that batch equilibrium melting is a valid analogue for mantle melting (even if fractional melting is more realistic) and that an independent measure of $P(\text{H}_2\text{O})$ is available (e.g. Albarède, 1992; Scarrow & Cox, 1995; Turner & Hawkesworth, 1995). A complementary approach involves mathematical inversion of erupted melt compositions with respect to an assumed mantle source and known subsolidus assemblages determined experimentally as a function of pressure and temperature, assuming fractional melting of a polybaric melt column.
Pressure and temperature calibrations

Covariation of FeO* in melts with both temperature and pressure has been recognized in experimental studies (e.g., Jacques & Green, 1980; Falloon et al., 1988) although simple temperature-dependent functions have been elusive as a result of problems such as Fe loss from experimental capsules. For a given bulk composition the temperature dependence of Fe is pronounced at low pressures (0.5–1 GPa) (Jaques & Green, 1980) but less so at higher pressures where garnet appears at the sub-solidus. Scarrow & Cox (1995) discussed this relationship with respect to Hirose & Kushiro’s (1993) data and projected isopleths for Fe and Mg in pressure–temperature space for melts equilibrated with fertile lherzolite HK-66. For melts of restricted MgO range (e.g. 14–16 wt %) over a pressure range of 1–3 GPa (implying decreasing melt fraction with increasing pressure) the Fe–temperature relationship is quasi-linear for both fertile and refractory bulk compositions (Fig. 11a; see below). Equivalent melt fractions formed in hydrous ‘sandwich’ and ‘diamond aggregate’ experiments also show this relationship over the pressure range 1.2–2.5 GPa, although temperatures are 50–100°C less for melts of equivalent FeO* content (Fig. 11a). The relationship between SiO₂ and pressure in experimental melts has been quantified by Albarené (1992) and Scarrow & Cox (1995), and is shown in Fig. 11b (see below) for anhydrous and hydrous conditions. Compositions of experimental melts from fertile peridotite (HK-66) were therefore calibrated for pressure and temperature according to the reported experimental conditions of Kushiro (1990) and Hirose & Kushiro (1993). It seems reasonable to apply these relationships to the natural melt compositions, assuming the latter were derived from sources of similar fertility, segregated over a similar pressure interval, and represent a similar range of melt fractions to those of the experiments.

Comparison of natural and experimental melts

In general, the Vietnamese melts are closer in composition to FeO*-rich melts generated from HK-66 than to those

Fig. 6. Plots of (a) Rb/Zr, (b) Ba/Zr and (c) Sr/Zr vs Ti/Zr ratios for samples from Dalat (DT), Phuoc Long (PS), Pleiku (PL), Buon Ma Tho (BMT), Xuan Loc (XL) and Re Island (RE), distinguishing early (I) and late (II) eruptive series.
Fig. 7. Plots of $\delta^{87}$Sr/$\delta^{86}$Sr vs MgO/FeO* (wt %) for representative basalts from Phuoc Long (SW sector), Pleiku, Buon Ma Thuot, Song Cau, Quang Ngai and Re Island (central sector), Xuan Loc, Dalat and Île des Cendres (SE sector), and Dien Bien Phu, Khe Sanh and Con Co island (northern centres), indicating the effects of crustal wallrock reaction [see Hoang et al. (1996) for details].

Fig. 8. Plots of Sm/Nd vs mg-number [$Mg \times 100/(Mg + Fe^{2+})$] for clinopyroxenes separated from spinel lherzolites [data published by Hoang et al. (1998)]. ◆: northern and central sectors; △: Dalat; open crosses: Île des Cendres.
Fig. 9. Plots of (a) FeO* vs SiO₂, (b) FeO* vs TiO₂, (c) FeO* vs K₂O and (d) FeO* vs P₂O₅ (in wt %) for basalt chemical type averages normalized to Mg-15 for Xuan Loc (XL), Pleiku (PL), Buon Ma Thuot (BMT), Dalat (DT), Phuoc Long (PH), Île des Cendres (Cendres), Quang Ngai (QN), Re Island (RE) and Con Co (CC). Compositions of early (I) and late (II) series are outlined where possible.

from refractory KBL-1 (Fig. 10). However, the inverse covariance of Fe-15 and Si-15 in the natural melts is less steep than that of anhydrous HK-66 melts and resembles those of melts equilibrated with fertile peridotite under hydrous conditions (Fig. 10). According to the anhydrous calibrations, melt segregation pressures range from nearly 4 GPa and temperatures of ~1470°C (e.g. alkali basalts from Xuan Loc) to <0.5 GPa and ~1400°C (quartz tholeiites from Dalat and Phuoc Long), representing a pressure–temperature slope of ~0.75°C/km. In contrast, segregation conditions based on the H₂O-undersaturated conditions may thus offer useful insights concerning ambient mantle Tₚ, topology of the MBL–TBL interface, and the relationship between convecting asthenosphere and extended, conductively cooled lithosphere.
**MELTING DYNAMICS AND MANTLE BOUNDARY LAYER MODEL**

A petrogenetic model for Vietnamese basalts needs to explain the relatively high volume basalt plateaux occupying extensional pull-apart nodes, the common progression from tholeiite-dominated early series, tapping refractory lithosphere-like sources, to later alkali basalt-dominated series, tapping relatively fertile, asthenospheric sources, a \( T_p \) of \( \approx 1440^\circ \text{C} \) and adiabatic gradient of >2°C/km, and LILE- and H\(_2\)O-rich asthenosphere. In the absence of data for lithosphere thickness, we developed a simple boundary layer model for Indochina that is consistent with the petrologic and geochemical data and general considerations of rheology and thermal state.

The progression from FeO*-poor to FeO*-rich (OIB-like) basalts in continental settings is commonly interpreted to reflect a change from lithospheric to asthenospheric mantle sources (e.g., Perry et al., 1987; DePaolo, 1988; Kempton et al., 1991; Gallagher & Hawkesworth, 1992; Turner & Hawkesworth, 1995). However, the problem of melting relatively non-ductile, refractory MBL has been raised as an objection to this interpretation and led to models invoking H\(_2\)O to lower the peridotite solidus (Gallagher & Hawkesworth, 1992) or mobilization of relict melt fractions ('mafic components') (Harry & Leeman, 1995). Harry & Leeman (1995) have contended that H\(_2\)O does not play a significant role in generating magma, as they believe that during initial stages of extension volatile fractions are quickly exhausted by early-formed melts. They suggested, rather, that significant melt fractions are produced at the solidi of 'mafic components' present in the lower lithosphere, depending on the amount and duration of lithospheric extension. However, despite the presence of pyroxenite (etc.) in peridotite restites, geothermal gradients based on xenolith thermobarometry (e.g., Ionov et al., 1998) differ significantly from adiabats of the type indicated by Vietnamese melt segregation conditions, suggesting non-ductile MBL is not decompressed to the extent needed to produce magma. In contrast, consideration of mantle dehydration reactions supports a key role for H\(_2\)O in melting refractory lithospheric mantle. Water is held in amphibole or mica at most lithospheric pressures (Lambert & Wyllie, 1970; Gallagher & Hawkesworth, 1992). Beyond the stabilities of these phases, small melt fractions saturated with H\(_2\)O and fusible mafic components will be able to form if temperatures exceed the H\(_2\)O-saturated solidus. However, even where such
As a framework for reconciling petrologic, thermal and geodynamic constraints, we adopt a lithospheric MBL whose mantle component is isotopically depleted and refractory, and rigidly attached to the crust (McKenzie & Bickle, 1988; Hawkesworth et al., 1990). Below this, the TBL is hypothetically characterized by attenuated shear-wave velocities and taken to mark the rheologic transition from dominantly convective to conductive heat transfer. It also marks the closest approach of the inflected thermal gradient to the peridotite solidus, allowing the highest potential for melting, and is assumed to be weak and isotopically enriched (Anderson, 1995). The asthenosphere is generally held to be isotopically depleted, unless lithospheric components have been added—either recycled via plumes or delaminated directly from above (e.g. Storey et al., 1989; Hart et al., 1992)—and residual to the extraction of primordial crust (Hofmann, 1988).

Hydrous phase breakdown would also produce changes in mantle rheology at similar depths (with or without melting) such that the TBL also marks the intersection of geotherm and dehydration curves (Lambert & Wyllie, 1970; Anderson, 1995). It is thus helpful to look at the combined effects of $T_p$ and $H_2O$ on peridotite solidi, which for a particular thermal gradient determine the depth and extent of melting. These relationships are shown in Fig. 12 along with the mantle adiabat interpolated for Vietnamese melts. The depth of phlogopite dehydration ranges from $\sim 3.8$ GPa ($\sim 110$ km depth) at normal $T_p$ to $<3$ GPa ($<90$ km depth) at $T_p > 1400^\circ C$ (Fig. 12). The interpolated Indochina $T_p (~1440^\circ C)$ is significantly higher than normal for subcontinental asthenosphere (e.g. McKenzie & Bickle, 1988; Wilson, 1993) suggesting an ambient MBL thickness of $\sim 80$ km ($\sim 2.6$ GPa) (Fig. 12). A conductive type 1 geotherm is believed to be typical for small or negligible lithospheric stretching factors ($\beta$) and is consistent with xenolith thermobarometric data from analogous settings (e.g. Ionov et al., 1998). It also agrees with thermal gradients computed as a function of uniform stretching (Latin & White, 1990), which suggest type 2 geotherms (Fig. 12) result at $\beta$ values of 2–3 (Fig. 13).

An important implication of the model is that uniform stretching causes changes in MBL bulk composition as a result of its partial conversion to TBL. This effect is illustrated in Fig. 12, where stretching (producing type 2 geotherms) causes upward migration of the rheologic boundary, separating solid from partially molten regions, relative to the pre-stretching (type 1 geotherm) boundary which was both rheologic and compositional. In other words, the upward migration of dehydration depth resulting from a local increase in $\beta$ converts the lowermost lithospheric mantle into rheologically weak TBL. Incorporation of refractory peridotite into the ductile region produces a 'macro'-heterogeneous column, which, with

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**Fig. 11.** Plots of (a) FeO$^*$ (wt %) vs temperature ($^\circ C$), and (b) SiO$_2$ (wt %) vs pressure (GPa) for experimental partial melts produced under hydrous and anhydrous conditions. Open circles, anhydrous melts of fertile peridotite HK-66 (Hirose & Kushiro, 1993); crossed circles, H$_2$O-undersaturated melts equilibrated with HK-66 ‘sandwiches’ (Kushiro, 1990). For H$_2$O-undersaturated experimental results (e.g. Kushiro, 1990) only runs with $<5$ wt % added H$_2$O and $K_{OL\rightarrow MELT}^\beta > 0.27$ were used. The regressions used in the melt segregation $P$-$T$ calculations are shown (see text).
Fig. 12. Schematic asthenosphere–lithosphere boundary model showing experimental solidus for fertile peridotite HK-66 (Hirose & Kushiro, 1993), stability fields of phlogopite and amphibole in peridotite (Modreski & Boettcher, 1973; Milhollen et al., 1974), spinel–garnet transition, and an H₂O-undersaturated solidus (assuming all H₂O in hydrous phases). H₂O-undersaturated melt segregation conditions of ‘primitive’ chemical type averages, estimated using the regressions in Fig. 10, are plotted, and suggest an asthenosphere potential temperature \( T_p \) of ~1440°C. The \( P-T \) field of incipient, H₂O-saturated melting is shaded. At \( T_p = 1440°C \) and \( \beta \approx 1.0 \) ambient conductive heat flow may be ~70–90 mW/m² (Pollack & Chapman, 1977) characterizing lithospheric mantle and producing geotherms of type 1. Adiabatic asthenospheric geotherms of type 2 may result from uniform stretching at \( \beta \approx 2.5 \), causing significant melting at pressures of between ~1–5 and 3–5 GPa. Mantle sections to the right show the dependence of MBL–TBL interface depth, spinel lherzolite transition and potential enrichment of the TBL, on the geotherm type (hence \( \beta \)) for a given \( T_p \).

We therefore propose that ambient lithospheric thicknesses of 80–100 km are reasonable at \( T_p \) of ~1440°C, and that thinning at transtensional ‘nodes’ leads to the penetration by and advanced polybaric melting of low-viscosity TBL and asthenosphere. Decompression melting would yield plateau basalt sequences of low-pressure, large-fraction tholeiites and high-pressure, low-fraction alkali basalts and basanites. This conclusion is consistent with element inversions on the Vietnamese basalt data conducted by D. McKenzie at Cambridge University which indicate a \( T_p \) of ~1450°C and partial melt column extending between garnet lherzolite and spinel lherzolite facies. Our conclusion that a major thermal anomaly exists beneath Indochina is supported by geomorphologic evidence for ~600 m uplift since the late Neogene (Bao & Hai, 1991) and high heat flow (Duchkov et al., 1992; Uyeda & Nagao, 1994) in south-central Vietnam.

**CONCLUSIONS**

1. Cenozoic basalt plateaux in southern and central Vietnam appeared over a total area of ~23,000 km² as part of a widespread regional volcanic episode. Eruptions at discrete centres appear to have involved at least two episodes separated by thick palaeosols, referred to as ‘early’ and ‘late’ series. The bi-episodal pattern is recognized at the Dalat, Phuoc Long, Pleiku, Buon Ma Thuot, Xuan Loc and Re Island centres, and probably other offshore localities, although at Buon Ma Thuot the compositional trend is ‘inverted’.

2. K–Ar and Ar–Ar age data indicate activity occurred over the following intervals: Dalat (17.6–7.9 Ma), Phuoc Long (<3–14 Ma), Buon Ma Thuot (5.8–1.67 Ma), Pleiku (4.5–0.8 Ma), Xuan Loc (0.83–0.14 Ma) and Île des Cendres (0.8–0 Ma), and reflects clockwise rotation of transtensional fractures. Palaeomagnetic data indicate little or no tectonic rotation since the India–Asia collision, although this can be reconciled with the extrusion model
Fig. 13. Adiabatic upwelling as a result of differential stretching of a convective geotherm generated from a 100 km thick MBL for an interior potential temperature $1440^\circ C$ (after Latin & White, 1990). Geotherms for respective stretching factor values ($\beta$) (continuous lines) and the locus of melt fraction ($F$) of $0.25$ (dot-dashed lines) are shown together with the dry peridotite solidus (bold dashed line) and typical H$_2$O-undersaturated solidus (fine dashed line) [adapted from McKenzie & Bickle (1988) and Latin & White (1990)]. Xenolith $P$-$T$ equilibration estimates reflecting conductive thermal gradients are expected to correspond to $\beta$ values between 1 and $\sim 1.5$ (type 1 geotherm in Fig. 12). Melt segregation $P$-$T$ estimates based on H$_2$O-undersaturated experiments (Fig. 11) indicate adiabats consistent with $\beta$ values between 2 and 3 (type 2 geotherm in Fig. 12) for $T_p = 1440^\circ C$. No melting would be expected for this $\beta$ range at $T_p = 1280^\circ C$ (see Latin et al., 1993).

If Indochina behaved as a non-rigid plate. The basalts probably reflect stretching associated with the change from left- to right-lateral motion on the Ailao Shan–Red River shear zone.

(3) With the exception of Buon Ma Thuot, early series basalts comprise high-SiO$_2$ and low-FeO* quartz and olivine tholeiites, tapping a relatively refractory (lithosphere-like) source, and a later series of low-SiO$_2$ and high-FeO* olivine tholeiites, alkali basalts and basanites, tapping a fertile (asthenosphere-like) source.

(4) Comparison of Mg-15-normalized basalts compositions with experimental melts allowed estimation of melt segregation pressures and temperatures: (a) anhydrous conditions: $< 4$ GPa and $\sim 1470^\circ C$ (for alkali basalts from Xuan Loc) to $< 0.5$ GPa and $\sim 1400^\circ C$ (quartz tholeiites from Dalat and Phuoc Long), and (b) H$_2$O-undersaturated conditions: $< 3.5$ GPa and $\sim 1450^\circ C$ to $\sim 1.5$ GPa and 1350–1400$^\circ C$. Hydrous rather than anhydrous conditions are favoured by: (a) high basaltic H$_2$O contents, (b) minimum melt segregation pressures consistent with melting below the thinned MBL, and (c) interpolated mantle adiabats of 2–3$^\circ C$/km (compared with $< 1^\circ C$/km for anhydrous conditions) consistent with fluid dynamic models. High potential temperatures may result from collision-extruded asthenosphere rather than a deep plume.

(5) Phlogopite stability provides a possible model for the base of the lithosphere MBL. Given the $P$-$T$ shape of phlogopite dehydration interpolated potential temperatures suggest ambient MBL thicknesses of $\sim 80$ km. Thinning of the MBL at transtensional ‘nodes’ leads to penetration and advanced polybaric melting of low-viscosity TBL and asthenosphere. Decompression melting of ‘macro’-heterogeneous columns yields plateau basalt sequences of low-pressure, large-fraction tholeiites and high-pressure, low-fraction alkali basalts and basanites.

ACKNOWLEDGEMENTS
We thank sponsors of the Indochina Research Consortium (Agip, Amerada Hess, Amoco, Arco, British Gas, BHP, British Petroleum, DuPont-Conoco, Chevron, Elf Aquitaine, Enterprise, Mobil, Petofina, Petronas, Phillips, Occidental, Shell, Total and Unocal) for support, and Dr Nguyen Trong Yem of the Institute of Geology of the National Centre for Science and Technology (Hanoi) and Dr Nguyen Xuan Bao of the Geological...
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Survey Division in Ho Chi Minh City for authorizing the study of samples collected. Nguyen Hoang acknowledges a Predoctoral Fellowship from the Carnegie Institution of Washington. We thank Keith Cox and Ikou Kushiro for invaluable discussions, and Colin Devey, Samuel Makasa and Godfrey Fitzton for reviews. Dan McKenzie is thanked for conducting element inversions on a major and trace element data set. Last but not least, Marje Wilson is thanked for her review and for extensive editorial input.
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