Geodynamic setting of Izu–Bonin–Mariana boninites

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Abstract: The Izu–Bonin–Mariana (IBM) forearc is characterized by the occurrence of boninite-like lavas. The study of the Cenozoic setting of the genesis of these boninitic lavas in light of modern geodynamic contexts in the Tonga and Fiji regions lead us to define three tectonic settings that favour the formation of boninites in back-arc basins in addition to previous settings that involve the presence of a mantle plume: (1) propagation at low angle between a spreading centre and the associated volcanic arc; (2) intersection at a high angle of an active spreading centre and a transform fault at the termination of an active volcanic arc; and (3) intersection at a right angle between an active spreading centre and a newly created subduction zone. A geodynamic model of the Philippine Sea Plate shows that boninites in the Bonin Islands are related to the second mechanism mentioned above, whereas Mariana forearc boninites are relevant to the third mechanism. In the early Eocene, the transform plate boundary bounding the eastern margin of the Philippine Sea Plate at the location of the present-day Mariana arc evolved into a subduction zone that trends perpendicular to the active spreading centre of the West Philippine Basin, somewhere around 43–47 Ma. The presence of a mantle plume in the vicinity of the subduction zone bounding the northern IBM arc explains boninites that erupted in its northern part, but only in early Eocene time.

Boninitic magmatism represents a distinctive style of subduction-related magmatism, which is interpreted to result from the melting of strongly depleted mantle that is variably metasomatized by slab-derived fluids or melts (Crawford et al. 1989; Pearce et al. 1992). Boninites are therefore a rare subduction-related magma type as they are the most H₂O rich, and require the most refractory sources, compared with normal island arc suites. These primitive arc rocks have distinctive geochemical characteristics, such as high magnesium, compatible element contents (Ni, Cr, Co) and Al₂O₃/TiO₂ ratios, low TiO₂, intermediate (andesitic) SiO₂ content (>53 wt%), U-shaped rare earth element (REE) patterns, extreme high-field-strength (HFSE) element depletions, and by the presence of very magnesian olivine phenocrysts (Crawford et al. 1989).

The genesis of boninites requires unique thermal and petrological conditions: a depleted mantle peridotite, a source of (C-O-H) volatiles and an abnormally high geothermal gradient at relatively shallow levels in the mantle wedge. Based on experimental studies, it is now widely accepted that boninite petrogenesis requires temperatures of 1200–1350°C, even c. 1480°C according to Falloon & Danyushevsky (2000), and pressures below 10 kbars that are attained at about 25 km in depth (Crawford et al. 1989; Pearce et al. 1992; Hawkins 1994). Such temperatures are several hundred degrees higher than those postulated in geophysical models for the thermal structure of the mantle wedge under the modern forearc regions (Crawford et al. 1989). Boninite generation requires, therefore, a mechanism capable of raising the ambient temperatures in the shallow mantle wedge by up to 500°C.

Concerning the degree of depletion of the boninite source peridotites, its variations are reflected by a wide range of CaO/Al₂O₃ values (0.4–0.85, Crawford et al. 1989) in primitive boninites. On the basis of the CaO/Al₂O₃ ratio, boninites have therefore been divided into two groups: low-Ca and high-Ca boninites with a boundary set at CaO/Al₂O₃ c. 0.75 (Crawford et al. 1989). Low-Ca boninites are interpreted as being produced from relatively more depleted sources than high-Ca boninites. Variations in CaO/Al₂O₃ values and TiO₂ content recorded in several boninite suites (e.g. Howqua, Australia: Crawford & Cameron 1985; Izu–Bonin forearc: From: LARTER, R.D. & LEAT, P.T. 2003. Intra-Oceanic Subduction Systems: Tectonic and Magmatic Processes. Geological Society, London, Special Publications, 219, 163–185. 0305-8719/03/$15.00 © The Geological Society of London 2003.
Umino 1986; Pearce et al. 1992; Taylor & Mitchell 1992; Taylor et al. 1994) have been explained as the result of progressive source depletion during boninite magma genesis. The nature of depleted mantle sources involved in boninite petrogenesis is a subject of debate. Most workers propose a depleted mantle wedge, residual from prior extraction of mid-ocean ridge basalts (MORB) (Crawford et al. 1989). However, it is also suggested that at least some high-Ca boninites could originate from a refractory ocean island basalt (OIB) mantle source (Sharaskin et al. 1983a, b; Falloon & Crawford 1991; Macpherson & Hall 2001). A possible involvement of OIB-related melts as enriching components in boninite petrogenesis has been proposed in several studies (Hickey & Frey 1982; Rogers et al. 1989; Falloon & Crawford 1991; Stern et al. 1991; Kostopoulos & Murton 1992). However, many authors suggest that the source of boninite enrichment is the result of the metasomatism of the subforearc mantle by hydrous fluids or melts derived from the subducting Pacific Plate (Bougault et al. 1981; Sharaskin 1981; Wood et al. 1981; Hickey & Frey 1982; Murton et al. 1992).

Boninites or boninite-like lavas occurrences have been reported in a variety of settings, including modern forearc regions of intraoceanic island arcs (Falloon & Crawford 1991; Pearce et al. 1992), back-arc basins (Kamenetsky et al. 1997), Phanerozoic and Proterozoic supra-subduction zone ophiolites (Rogers et al. 1989; Ballantyne 1991; Meffre et al. 1996; Bédard et al. 1998; Eissen et al. 1998; Bédard 1999; Wyman 1999), Achaean greenstone belts (Kerrich et al. 1998) and in continental or epicontinental settings (Rogers & Saunders 1989: Piercey et al. 2001). The best studied locations of Cenozoic–recent boninite lavas are the Izu–Bonin–Mariana forearc (Fig. 1) (Crawford et al. 1981, 1989; Umino 1985, 1986; Tatsumi & Maruyama 1989; Stern et al. 1991; Pearce et al. 1992; Stern & Bloomer 1992; Taylor et al. 1994; Hawkins & Castillo 1998; Hickey-Vargas 1989), the North Tonga Ridge (Fig. 2b) (Falloon et al. 1989: Falloon & Crawford 1991; Sobolev & Danyushevsky 1994; Danyushevsky et al. 1995), the southern termination of the New Hebrides island arc (Fig. 2c) (Monzier et al. 1993), the Valu Fa Ridge in Lau Basin (Fig. 2b) (Kamenetsky et al. 1997) and the Setouchi volcanic belt of Japan (Tatsumi & Maruyama 1989; Tatsumi et al. 2001).

A variety of models for the genesis of boninites have been proposed, such as arc infancy and catastrophic melting (Pearce et al. 1992; Stern & Bloomer 1992), ridge subduction beneath young and, hence, hot plate (Crawford et al. 1989), back-arc basin formation (Coish et al. 1982), mantle plume–island arc interaction (Kerrich et al. 1998; Macpherson & Hall 2001) and ridge propagation into a forearc region (Fallon & Crawford 1991; Monzier et al. 1993; Meffre et al. 1996). Concerning the genesis of Izu–Bonin–Mariana (IBM) arc boninitic lavas, different tectonic settings have been proposed (Pearce et al. 1992; Stern & Bloomer 1992; Macpherson & Hall 2001). However, the lack of constraints about the Tertiary behaviour of the West Philippine Basin, that was bordered on its eastern margin by the IBM arc during the production of boninites (Fig. 1), prevents an accurate definition of the geodynamic setting and, consequently, the cause of their eruption. Here, we propose a model for the formation of the IBM boninites with the help of tectonic reconstructions of the West Philippine Basin, as well as comparisons with modern analogues of boninite formation in the North Tonga Ridge, the Valu Fa Ridge (Lau Basin) and the New Hebrides Arc (North Fiji Basin).

Geodynamic setting of the Tonga and New Hebrides boninites

Occurrences of boninites are reported from the North Tonga Ridge (Falloon et al. 1989; Falloon & Crawford 1991; Sobolev & Danyushevsky 1994; Danyushevsky et al. 1995), the Valu Fa Ridge in the Lau Basin (Kamenetsky et al. 1997) and from the southern New Hebrides arc (Monzier et al. 1993) (Fig. 2). All these boninites were emplaced in the vicinity of active back-arc spreading centres intersecting active island arcs. They all are younger than Pliocene. From these modern examples, we aim to understand the relationships between the geodynamic context and the occurrence of boninites.

Intersection between the Tofua arc and the NE Lau spreading centre

High-Ca boninites, arc tholeiites and OIB-like lavas were dredged from the northernmost part of the Lau Basin (Falloon et al. 1989; Sharaskin et al. 1983b; Danyushevsky & Sobolev 1987; Fallon & Crawford 1991; Danyushevsky et al. 1995). Boninites occur close to the northern termination of the trench, where it swings west into a major transform fault (Fig. 2b). Boninites then erupt at the location of a tear fault within the Pacific Plate, just above the subducting plate edge (Millen & Hamburger 1998) (Fig. 3). They occupy a 50 km-long section of the trench slope.
Fig. 1. Tectonic setting of the Philippine Sea Plate, with location of volcanic sites mentioned in the text and in Tables 1 and 2. Boninites are found along the Izu–Bonin–Mariana arc and within the Zambales ophiolite (Luzon). Tectonic and magnetic features in Shikoku, Parece–Vela and Mariana basins are indicated. They provide constraints about the direction of opening of these basins, and consequently about the former location of boninitic lavas sites along the Palau–Kyushu Ridge before the opening of these basins since 30 Ma. The arrow indicates as an example the path chosen to reconstruct the former position of sites 458/459 along the Palau–Kyushu Ridge.

Arc tholeiites were dredged 50 km east of the boninite outcrops (Sharaskin et al. 1983a). Alkaline OIB-like basalts, which have been shown to originate from the adjacent Samoan plume, are located 60 km to the west of boninite outcrops (Zlobin et al. 1991). Ages of boninites range from 1.4 to 2 Ma (Ar/Ar ages) (Danyushevsky et al. 1995).

The geodynamic context as the location of boninitic lavas is complex as the lavas occur not
Fig. 3. Perspective cartoon illustrating the geometry of the subduction trench-transform transition in the vicinity of an active spreading centre. The open edge above the tear within the Pacific Plate favours upward asthenospheric flow and high heat flow. In the northern termination of the Tonga Trench, this tear allows the southward migration of the hot asthenospheric material related to the nearby Samoan plume. Boninites were formed at the termination of the active volcanic arc, at the intersection of the active spreading centre and the transform zone that accommodates the opening of the back-arc basin.

only in the vicinity of a trench-transform fault transition, but also near the Samoan active plume, and also at the termination of an active back-arc spreading centre (Falloon & Crawford 1991; Wright et al. 2000; Pelletier et al. 2001) (Fig. 2b). Because of this particular tectonic setting, different hypotheses have been proposed to explain the genesis of boninites in this northeastern part of the Lau Basin.

The first one, proposed by Falloon & Crawford (1991), is based on the isotopic (Sr, Nd) composition of lavas. The boninites mantle sources are part of a regional OIB mantle domain upwelling beneath the Tonga subduction zone system. This mantle source was of refractory lherzolite composition, depleted in basaltic components by prior generation of Lau Basin crust. It has been enriched in incompatible elements by one or more metasomatic phases, suggested to be a hydrous fluid from the subducting lithospheric slab, a carbonatite melt and a small-degree silicate partial melt, both derived from OIB source mantle. The genesis of boninites is explained by the presence of the NE Lau Spreading Centre, which intersects the northernmost Tofua Arc (Fig. 2b). Upwelling asthenospheric mantle beneath the spreading axis may, indeed, cause partial melting of the refractory peridotite located in the mantle wedge above the subducting Pacific Plate at shallow depth. According to this model, the source of heat flow that is necessary to melt the refractory mantle is thus the ascending diapirs of MORB magmas beneath the spreading ridge.

The other main hypothesis to explain the formation of boninites at the northern end of the

Fig. 2. Tectonic setting of the Lau and North Fiji basins. (a) is modified from Pelletier et al. (1998). Blue boxes indicate the location of maps (b) and (c). (b) Two-dimensional shaded bathymetric map of the Lau Basin, with location of boninitic lavas. Boninitic lavas are found at two places along the Tofua arc: at its northern end near the NE Lau Spreading Centre (NELSC) termination, and more to the south, close to the southern end of the Valu Fa Ridge. (c) Two-dimensional shaded bathymetric map of the southern part of the North Fiji Basin. Boninites are found at the intersection between the active Central Spreading Ridge (CSR) and the Hunter Fracture Zone, i.e. the eastern continuation of the New Hebrides (Vanuatu) Trench. The Vanuatu arc north of 21°N displays only normal island arc tholeiitic lavas. Bathymetric data are satellite derived (Smith & Sandwell 1997).
Lau Basin is suggested by Danyushevsky et al. (1995). These authors agree with Falloon & Crawford (1991) concerning the mantle source composition of boninitic magmas, but the source of heat flow to melt this refractory mantle differs in their model. They suggest that abnormal high heat flow results from the intrusion of hot residual plume mantle diapirs in the mantle wedge above the subduction zone. The open edge of the mantle wedge beneath the transform fault would have, indeed, allowed the Samoan plume to intrude southward above the subducted slab. This intrusion would also be favoured by the eastward rollback of the Tonga Trench and the southward asthenospheric flow beneath the Lau Basin (Smith et al. 2001). The variably enriched boninitic melts would be due to the mixing of these magmas with earlier formed OIB-like melts, during their ascent to the surface. This hypothesis of Danyushevsky et al. (1995) is supported by the presence of many recent seamounts in the northern part of the Lau Basin, whose lavas are isotopically similar to Samoan OIB-like basalts. This shows that either plume mantle, or mantle plume-derived melts, occupy a large area of the mantle wedge in this region.

**Intersection between the Valu Fa Ridge and the Tofua arc**

The Valu Fa Ridge is the southernmost active spreading axis of the Lau Basin. It is located 40 km west of Ata Island, a volcano belonging to the Tofua arc that results from volcanism related to the westward subduction of the Pacific Plate along the Tonga Trench (Fig. 2b). Several lava samples have been dredged at two sites from the southern termination of the Valu Fa Ridge, c. 50 km north of its southward propagating tip and close to its intersection with the Tofua arc (Frenzel et al. 1990; Sunkel 1990; von Stackelberg et al. 1990; Fouquet et al. 1991; Kamenetsky et al. 1997) (Fig. 2b). These rocks mainly consist of normal island arc basaltic and andesitic suites with strong geochemical affinities with subduction zone magmas (Jenner et al. 1987; Boeseghug et al. 1990; Vallier et al. 1991). However, some rocks dredged close to the intersection between the active spreading axis and volcanic arc consist of low-Ca boninitic-like, primitive lavas (Kamenetsky et al. 1997). It is suggested that these boninitic lavas have been formed due to the melting of a shallow, refractory, hydrated mantle that has been metasomatized by fluids or melts coming from the subducting slab. The melting of the refractory mantle beneath the Tofua arc would be due to its juxtaposition against hot asthenospheric mantle that is currently supplying the Valu Fa Ridge (Kamenetsky et al. 1997).

**The southern termination of the Vanuatu Trench**

Active back-arc extension occurs in the North Fiji Basin, east of the Vanuatu (New Hebrides) subduction zone. The basin is characterized by the synchronous existence of several active spreading axes (Fig. 2 a, c). At least since 2 Ma, the Central Spreading Ridge of the basin has trended N–S and abutted the Hunter Fracture Zone, i.e. the eastern termination of the New Hebrides arc (Pelletier et al. 1998). It is active with a 8 cm a⁻¹ full spreading rate in a N72°E oblique direction (Fig. 2c).

In the southern New Hebrides arc, two spatially distinct arc magmatic suites are described by Monzier et al. (1993) from analyses of several dredged rocks (see location of dredges on Fig. 2c). Between 21°S and 22°S, a normal island arc tholeiitic magmatic suite is essentially similar to those occurring in the main part of the New Hebrides arc central chain volcanoes, whereas south of 22°S, at the southern termination of the arc, a high magnesian andesite suite appears with the more mafic end-members having mineralogical and compositional affinities with high-Ca boninites (Monzier et al. 1993) (see location on Fig. 2c). Monzier et al. (1993) suggest that most of the sampled volcanoes are probably younger than 2 or 3 Ma, considering the recent evolution of the southern termination of the arc. This seems to be confirmed by ages of the Matthew and Hunter volcanoes that range from 0.8 to 1.4 Ma (Maillet et al. 1986; Monzier et al. 1993). True high-Ca boninites have been dredged only 100 km east of Hunter Island, in the area where the Central Spreading Ridge abuts the Hunter Fracture Zone (Sigurdssoon et al. 1993). Monzier et al. (1993) suggest that the generation of high-Ca boninites or boninite-like magmas occurs by melting of a refractory mantle at a shallow level under the Hunter Ridge. The extra heat that is necessary to melt such a mantle would be provided by the rising diapirs supplying magmas to the intersecting spreading axis. Doleritic inclusions that are common in some Mg-rich acid andesites of Matthew and Hunter islands may represent quenched blobs of near parental boninitic high-Mg andesite magma incorporated into the acid andesite host magma prior to eruption. These inclusions are, indeed, certainly co-magmatic (Maillet et al. 1986; Monzier et al. 1993) and these authors...
suggest that magma mixing probably occurred along the termination of the New Hebrides arc.

*Intersection between arc and back-arc volcanism: a geodynamic context favourable to the genesis of boninites*

We learn from the above three recent tectonic settings of boninite eruptions that back-arc rifting or spreading occurs close to subduction-related arc volcanism:

- the southern Valu Fa spreading centre almost parallels and merges with the Tofua volcanic arc. Boninites are located at about 140 km above the subducting Pacific slab;
- the northern part of the NE Lau Spreading Centre and the southern part of the Central Spreading Ridge (North Fiji Basin) show similar geodynamic contexts. Boninites are found at the intersection between the back-arc spreading centre and a transform fault that connects at about right angle with the nearby subduction zones (Tonga or New Hebrides subduction zones). In northern Tonga, boninites erupt above the Pacific subducting slab whose top is 80 km deep, whereas, in southern New Hebrides, the Australian slab could reach depths around 80 km under the site of boninite eruption. Another similarity concerns their location just above the tear that allows the same plate to subduct along the trench and to slip along the transform fault (Fig. 3).

Earthquake distribution and source-mechanism determinations indicate progressive downwarping and tearing of the Pacific Plate as it enters the northernmost segment of the Tonga subduction zone (Millen & Hamburger 1998). Based on the same arguments, we suspect that the same situation occurs at the southernmost termination of the New Hebrides subduction zone. The occurrence of a tear is known to favour upward asthenospheric flow, unusual volcanic products and, thus, high heat flow (Lees 2000). As the subducting plate is torn beneath the transform fault, a slab window allows mantle material to flow around the exposed edge of the subducting plate. Slab rollback not only causes the entire wedge mantle to rise and decompress (Pearce et al. 1992; Bédard, 1999), but here also induces asthenospheric flow around the edge of the subducting plate, shearing of the edge and causing anomalously high magma production as seen at the Kamtchatka–Bering junction (Yogodzinski et al. 2001). The intrusion of the Samoan mantle plume into the mantle wedge above the Tonga slab evoked by Danyushevsky et al. (1995) is probably favoured by this asthenospheric flow.

We conclude that modern examples of boninitic eruptions all occur where back-arc spreading occurs above a subducting slab whose top is 80–140 km deep, commonly, but not always, above a tear in the subducting slab that could favour asthenospheric flow around the subducting plate edge, producing high heat flow (Fig. 3).

*Occurrence of boninites along the Izu–Bonin–Mariana arc*

The Izu–Bonin–Mariana (IBM) arc bounding the eastern margin of the Philippine Sea Plate is characterized by the presence of an important boninitic suite that occurs mainly within the present-day forearc domains (e.g. DSDP sites 782, 786, 458, Chichijima, Saipan, Guam, Palau islands, and dredge sites MD28, MD50, MD51, DM1403) (Fig. 1) (Dietrich et al. 1978; Sharaskin et al. 1980; Crawford et al. 1981, 1989; Umino 1985, 1986; Stern et al. 1991; Pearce et al. 1992; Taylor et al. 1994). The IBM region has been intensively studied and numerous geochronological and geochemical analyses were carried out. Boninites are found in close relationship with tholeiitic lavas, as well as other typical arc lavas, such as andesites, dacites and rhyolites (e.g. Crawford et al. 1981; Umino 1985; Pearce et al. 1992; Bloomer et al. 1995; Cosca et al. 1998).

In Hole 458, on the Guam and on Chichijima islands, boninites are interbedded with tholeiites (see locations on Fig. 1). In Hole 786, they dominate the lowermost part of the deep crustal section through the outer arc, and are closely linked to bronzite andesites, andesites, dacites and rhyolites in the main part of the volcanic edifice. At this site, they are also found as younger sills or dykes that intrude the main volcanic edifice (Pearce et al. 1992). Radiometric dating of boninites and indirect palaeontological dating of their closely associated sediments show ages of emplacement that range between early Eocene and Oligocene times, with a principal phase of volcanism between 48 and 41 Ma, possibly followed by a second minor event around 35–34 Ma, which led to the emplacement of sills and dykes within the main volcanic edifice constructed during the first phase (Maruyama & Kuramoto 1981; Meijer 1983; Reagan & Meijer 1984; Mitchell et al. 1992; Pearce et al. 1992; Cosca et al. 1998) (Tables 1, 2,
### Table 1. Ages of arc volcanism along the Izu–Bonin Arc

<table>
<thead>
<tr>
<th>Site/sample</th>
<th>Nature of sample</th>
<th>K/Ar (Ma)</th>
<th>Ar/Ar (Ma)</th>
<th>Confidence</th>
<th>Ref.</th>
</tr>
</thead>
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<tr>
<td>GDP-8-12</td>
<td>plagiogranite</td>
<td>37.4</td>
<td>–</td>
<td>–</td>
<td>Malyarenko &amp; Lelikov (1995)</td>
</tr>
<tr>
<td>N4-90</td>
<td>plagiogranite</td>
<td>42.7</td>
<td>–</td>
<td>–</td>
<td>Malyarenko &amp; Lelikov (1995)</td>
</tr>
<tr>
<td>N4-91</td>
<td>plagiogranite</td>
<td>31.2</td>
<td>–</td>
<td>–</td>
<td>Malyarenko &amp; Lelikov (1995)</td>
</tr>
<tr>
<td>N4-91</td>
<td>plagiogranite</td>
<td>26.2</td>
<td>–</td>
<td>–</td>
<td>Malyarenko &amp; Lelikov (1995)</td>
</tr>
<tr>
<td>D-60</td>
<td>plagiogranite</td>
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<td>–</td>
<td>–</td>
<td>Malyarenko &amp; Lelikov (1995)</td>
</tr>
<tr>
<td>D-76</td>
<td>tonalite</td>
<td>48.5</td>
<td>–</td>
<td>–</td>
<td>Malyarenko &amp; Lelikov (1995)</td>
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<tr>
<td>–</td>
<td>volcanic rock</td>
<td>49</td>
<td>–</td>
<td>–</td>
<td>Mizuno <em>et al.</em> (1977)</td>
</tr>
<tr>
<td>–</td>
<td>granodiorite</td>
<td>38</td>
<td>–</td>
<td>–</td>
<td>Shibata &amp; Okuda (1975)</td>
</tr>
<tr>
<td>same</td>
<td>granodiorite</td>
<td>–</td>
<td>51 (<em>fission tracks</em>)</td>
<td>–</td>
<td>Nishimura (1975)</td>
</tr>
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<td>Site 782A/B</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>45X-1</td>
<td>clast, dacite</td>
<td>18.1</td>
<td>–</td>
<td>3</td>
<td>Mitchell <em>et al.</em> (1992)</td>
</tr>
<tr>
<td>50X-1</td>
<td>clast andesite</td>
<td>30.7</td>
<td>–</td>
<td>–</td>
<td>Mitchell <em>et al.</em> (1992)</td>
</tr>
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<td>1W-1</td>
<td>clast, andesite</td>
<td>36.2</td>
<td>–</td>
<td>2</td>
<td>Mitchell <em>et al.</em> (1992)</td>
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<td>Site 786 A/B</td>
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<td>12X-1</td>
<td>clast, basalt</td>
<td>9.2</td>
<td>–</td>
<td>3</td>
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<td>16X-1</td>
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<td>3</td>
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<td>2R-2</td>
<td>flow, ICBzA</td>
<td>41.0</td>
<td>–</td>
<td>–</td>
<td>Mitchell <em>et al.</em> (1992)</td>
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<tr>
<td>2R-2</td>
<td>flow, ICBzA</td>
<td>–</td>
<td>46.6, 46.7, 45.7</td>
<td>–</td>
<td>Cosca <em>et al.</em> (1998)</td>
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<td>6R-2</td>
<td>dike, HCB</td>
<td>17.3</td>
<td>–</td>
<td>–</td>
<td>Mitchell <em>et al.</em> (1992)</td>
</tr>
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<td>8R-1</td>
<td>breccia, andesite</td>
<td>9.2</td>
<td>–</td>
<td>3</td>
<td>Mitchell <em>et al.</em> (1992)</td>
</tr>
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<td>11R-1</td>
<td>breccia, ICB</td>
<td>25.7</td>
<td>–</td>
<td>3</td>
<td>Mitchell <em>et al.</em> (1992)</td>
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<tr>
<td>16R-1</td>
<td>breccia, andesite</td>
<td>33.7</td>
<td>–</td>
<td>1</td>
<td>Mitchell <em>et al.</em> (1992)</td>
</tr>
<tr>
<td>20R-1</td>
<td>dike, ICB</td>
<td>12.0</td>
<td>–</td>
<td>3</td>
<td>Mitchell <em>et al.</em> (1992)</td>
</tr>
<tr>
<td>20R-1</td>
<td>flow, andesite</td>
<td>33.8</td>
<td>–</td>
<td>4</td>
<td>Mitchell <em>et al.</em> (1992)</td>
</tr>
<tr>
<td>21R-1</td>
<td>flow, andesite</td>
<td>16.7</td>
<td>–</td>
<td>3</td>
<td>Mitchell <em>et al.</em> (1992)</td>
</tr>
<tr>
<td>21R-2</td>
<td>dike, HCB</td>
<td>33.8</td>
<td>–</td>
<td>1</td>
<td>Mitchell <em>et al.</em> (1992)</td>
</tr>
<tr>
<td>30R-1</td>
<td>flow, ICBzA</td>
<td>39.8</td>
<td>–</td>
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### Sites 786/782

**Sediments over volcanics**
- nannofossils
  - z. CP13c: 44.4–47.0
  - z. P10/P11: 45–52
- foraminifers
  - Milner (1992)

**Sediments intercalated**
- microfossil
  - middle–late Eocene
  - Milner (1992)

**Mean values**
- 41.3 and 34.8

**Conclusion**
- 38–44 and 34–36
Table 1. continued

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* Location of sites in Fig. 1.
† HCB, high-Ca boninite; ICB, intermediate-Ca boninite; ICBzA, intermediate-Ca bronzite–andesite; LCB: low-Ca boninite; LCBzA: low-Ca bronzite andesite; COBA, clinopyroxene–orthopyroxene phyric basaltic andesite; COPBA, clinopyroxene–orthopyroxene–plagioclase phyric basaltic andesite.
‡ Numbers 1, 2 and 3 in the ‘Confidence’ column indicate: (1) well constrained; (2) poorly constrained (altered sample or discrepancies between isochron and plateau ages for examples); (3) very poorly constrained (altered samples); and (4) reset ages due to the second boninitic event, according to authors (e.g. Cosca et al. 1998).
§ In the K/Ar and Ar/Ar columns, P, i, and t indicate plateau, isochron and whole rock ages, respectively. Palaeontological ages are also indicated, when they are determined on sediments that are closely associated with boninitic lavas. Those ages that are poorly constrained are given in normal type, and ages that are not poorly constrained are given in bold type.

Fig. 4). The low reliability of many dates prevents us from confirming continuous boninitic activity during this period. Younger ages ranging between 18 and 33 Ma have been determined from boninitic lavas in several sites, but they often conflict with palaeontological ages, leading some authors such as Cosca et al. (1998) to question their reliability.
Table 2. Ages of arc volcanism along the Mariana arc

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<th>Site/sample*</th>
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<td>49.0¹, 49.3⁵, 49.9⁶</td>
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Location of sites in Fig. 1. On Guam Island, the Facpi Formation is Middle Eocene in age, and it consists in interbedded boninite pillow lavas and breccias, locally with tholeiitic rocks as the youngest member. The Alutom Formation paraconformably overlays the Facpi Formation, and is constituted by calc-alkaline, tholeiitic, and boninitic pyroclastic rocks and lavas (Meijer et al. 1983). On Palau, the Babelduap Formation is the oldest one, and contains boninitic lavas (Cosca et al. 1998). Boninites are also found in the Aimeliik Formation. The youngest unit is called Arakabesan Formation. On Saipan, the Sankakuyama Formation is the lowest unit on this island, and it consists in boninite and dacite series. The younger Hagman Formation is composed of andesitic pyroclastic rocks and lavas flows (Meijer et al. 1983). See Table 1 for the abbreviations used in the K/Ar, Ar/Ar and Confidence columns.
Fig. 4. Ages of volcanic events that are inferred along the Izu–Bonin arc between 55 and 15 Ma. Palaeontological and radiometric dating have been reported from Tables 1 (Izu–Bonin arc) and 2 (Mariana arc). Ages that are poorly constrained (in normal type in the K/Ar and Ar/Ar columns in Tables 1 and 2) are not reported on this figure.
Previous models for the formation of boninites along the Izu-Bonin-Mariana arc

The cause of the widespread boninitic volcanism along the IBM arc is the subject of considerable debate. Several hypotheses have been proposed. (1) According to Crawford et al. (1989), boninite formation is due to the subduction of an active spreading centre subparallel to the proto-Palau–Kyushu trench bounding the eastern margin of the Philippine Sea Plate at this time. As the hot lithosphere on either side of the spreading centre approached the trench, the dip of the slab probably decreased and isotherms in the mantle wedge were raised, causing partial melting of depleted subforearc oceanic lithosphere and generation of high-Ca boninites. (2) Pearce et al. (1992) and Taylor et al. (1994) suggested that the boninites formed due to the subduction of young, hot oceanic crust beneath hot, young, lithosphere of the Philippine Sea Plate. Subduction is supposed to have started along a ridge-to-ridge transform fault. (3) Stern & Bloomer (1992) proposed that boninites in the IBM forearc are due to the nucleation of subduction along an active transform boundary separating an active spreading ridge between Asia and Australia to the southwest, and the Kula–Pacific Ridge to the northeast. Because of its old age, the lithosphere east of the transform fault would have been gravitationally unstable and subsided, leading to the inception of a subduction zone. Migration of the asthenosphere over the subsiding lithosphere would have entailed adiabatic decompression, and water would have been released from the subducted lithosphere and sediments. These two processes should have led to extensive melting and production of boninites, particularly in the forearc region subjected to extension due to trench rollback. (4) Another hypothesis to explain the boninitic volcanism along the IBM arc is proposed by Macpherson & Hall (2001). These authors observe that regional uplift, ocean island basalt-style magmatism and high heat flow characterized the northern part of the Philippine Sea Plate in middle Eocene times. They also notice that the reconstructed middle Eocene location of the IBM arc and West Philippine Basin lies close to the present location of the Manus Basin, where there is evidence for the existence of a mantle plume. They therefore speculate that middle Eocene magmatism along the IBM arc developed its particular character because subduction was initiated above a mantle plume. The thermal anomaly in the mantle in the subduction zone would have provided the extra heat necessary to melt highly depleted mantle to form boninitic magmas along the volcanic arc. The OIB-like lavas that were recovered from the West Philippine Basin, and the uplift of the Amami–Oki–Daito province would also be due to the presence of the same mantle plume.

Geodynamic setting of the Philippine Sea Plate at the epoch of boninite formation: reconstruction

Any model of boninite genesis requires a good knowledge of the geodynamics of the Philippine Sea Plate during their emplacement. As the Parece–Vela and Shikoku basins started to open at c. 30 Ma (Okino et al. 1998), i.e. after the end of boninitic magmatism (Cosca et al. 1998), we focus on the Palaeogene history of the West Philippine Basin that was forming the Philippine Sea Plate at the time of boninitic volcanism. Deschamps et al. (1999, 2002), Fujioka et al. (1999) and Okino et al. (1999) studied the spreading mechanisms within the West Philippine Basin based on new data such as high-resolution bathymetry, magnetism and gravity, and on new radiometric dating on rocks recovered from the basin. Deschamps & Lallemand (2002) reconstructed the tectonic history of the Philippine Sea Plate between 54 and 30 Ma. We integrated results of these studies, as well as all available data within the region, such as geochronological and geochemical data, tectonic events and plate movements along the basin margins, and palaeomagnetic data from the basin and its surroundings.

We present in Figure 5 the reconstructed Philippine Sea Plate from early middle Eocene to late Eocene time, i.e. the time when most boninitic lavas were emplaced along the proto-IBM arc. The global model, and data used to establish it, are discussed in Deschamps (2001) and Deschamps & Lallemand (2002). Here we

Fig. 5. Evolutionary model from the Philippine Sea Plate between 55 and 34.35 Ma. The movement of the plate is constrained by palaeomagnetic data acquired in the West Philippine Basin, and from its northern and southern margins. The plate was restored every 5 Ma by successive rotations on a sphere that is tangential to the WGS84 ellipsoid (Deschamps 2001; Deschamps & Lallemand 2003). The Benham mantle plume is supposed to have remained fixed with respect to the mantle frame. Tholeiitic and boninitic volcanic events are reported, after Shikoku, Parece–Vela and Mariana basins are ‘closed’ perpendicularly to the spreading fabric and magnetic lineations, and parallel to the fracture zones (see Fig. 1).
focus on the Philippine Sea Plate itself and on
evidence for arc and mid-ocean ridge lava erup-
tion within the plate, in order to constrain the
geodynamic setting of boninite eruption. In
order to determine the location of the different
parts of the proto-IBM arc before the opening of
the Shikoku and Parece-Vela basins at 30 Ma
(Okino et al. 1998) and of the Mariana Basin
after 6 Ma (Hussong & Uyeda 1981), we have
‘closed’ these basins perpendicularly to the mag-
etic anomalies (when available) and to the
spreading fabric (Shikoku and Parece–Vela
basins: Chamot-Rooke et al. 1987; Okino et al.
1998; Sumisu Rift: Taylor 1992; Mariana Basin:
Martinez et al. 1995; unpublished data of
JAMSTEC and Kobe University) (Fig. 1).

At c. 50 Ma (Fig. 5), an efficient spreading
system is established in the West Philippine
Basin. The basin is characterized by the pres-
ence of two spreading axes. The driving force for
its opening is provided by slab rollback along the
two surrounding subduction zones. A mantle
plume is active in the northernmost part of the
basin. Its activity is responsible for the eruption
of OIB-like lavas at sites 446 (Minami–Daito
Basin) and 294/5 (northern West Philippine
Basin) (Hickey-Vargas 1991, 1998; Macpherson
& Hall 2001). The OIB-like lavas that erupted a
few millions years later at Site 292 (Benham
Rise) are probably also due to the same plume.
Similarly enriched MORB (E-MORB) lavas
found at Site 291 are interpreted to result from
mixing of OIB and N-MORB sources. There are
no age data for normal MORB (N-MORB)-like
lavas recovered at DSDP Site 1201 (Shipboard
Scientific Party 2001) or at dredge site D12
(Bloomer & Fisher 1988), but their locations in
the northern and southwesternmost part of the
basin, respectively, suggest that they erupted at
about 50 Ma, according to the reconstruction
(Deschamps & Lallemand 2002). Petrological
and geochemical analyses of these samples
demonstrate that their area of emplacement was
not influenced by the mantle plume. Arc volcan-
ism occurred along the northern part of the
Palau–Kyushu Ridge after c. 51 Ma, possibly
45 Ma (see Table 1, Fig. 4). Boninitic or
boninitic-like lavas of this age occur at sites 786
and 782, and in the Bonin Islands, but there is
no evidence for arc volcanism along the southern
part of the proto-Palau–Kyushu Ridge. Relative
movement between the Philippine Sea Plate and
the Pacific Plate, as well as the orientation of the
proto-Palau–Kyushu Ridge, indicate that strike-
slip movement was occurring along this bound-
ary, accommodating the opening of the West
Philippine Basin. It is very possible that a small
amount of opening occurred within the
Minami–Daito Basin in the early Middle
Eocene, as indicated by ages of rocks at Site 446.
We, unfortunately, lack reliable constraints
about the opening of this basin.

At c. 46 Ma (Fig. 5), the transform fault separ-
atting the Philippine Sea Plate and the Pacific
Plate still accommodated the opening of the
West Philippine Basin at an intermediate
spreading rate (Hilde & Lee 1984; Deschamps
2001). N-MORB lavas erupted at Site 447
located in the eastern part of the basin, and also
possibly in its westernmost part. Some 46 Ma N-
MORB lavas occur within the Coto Block of the
Zambeles ophiolite (Fuller et al. 1989), which is
shown to be an autochthonous part of the Philip-
Palau-Kyushu Ridge (Hickey-Vargas 1991, 1998; Okino
et al. 1987; Okino et al.
1998) and of the Mariana Basin:
Martinez et al. 1995; unpublished data of
JAMSTEC and Kobe University) (Fig. 1).

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Similarly enriched MORB (E-MORB) lavas
found at Site 291 are interpreted to result from
mixing of OIB and N-MORB sources. There are
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Scientific Party 2001) or at dredge site D12
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slip movement was occurring along this bound-
ary, accommodating the opening of the West
Philippine Basin. It is very possible that a small
amount of opening occurred within the
Minami–Daito Basin in the early Middle
Eocene, as indicated by ages of rocks at Site 446.
At c. 34/35 Ma (Fig. 5) spreading rate in the West Philippine Basin continued to decrease. OIB-like lavas were still erupting in the vicinity of the Benham Rise, demonstrating the presence of the active mantle plume in this area. Rifting started in the Caroline Sea at c. 34 Ma (Hegarty & Wessel 1988). Hegarty & Wessel (1988) suggest that an active mantle plume was present under the basin at that time, and interacted with the spreading system, being responsible for important excess volcanism during spreading. Boninitic or boninitic-like lavas erupted at several places along the Palau–Kyushu Ridge (at least at sites 786/782, 458/459 and at Bonin and Guam islands). They are found as sills and dykes intruding former volcanic formations. However, the main volcanic products at this epoch are normal island arc lavas, such as tholeiites, andesites and dacites. Adakitic lavas erupted in Catanduanes Island, probably due to the onset of subduction of the young and hot lithosphere of the West Philippine Basin along the east Luzon Trough.

Inconsistencies between proposed models for boninites formation and the Cenozoic geodynamic setting of the Philippine Sea Plate

Our reconstruction shown in the previous section is generally incompatible with the existing models of boninite generation.

- The model of formation of boninitic lavas by subduction of a very young plate beneath the Philippine Sea Plate fails for several reasons: (i) The hypothesis implies that the new subduction zone was initiated close to an active spreading centre located on the subducting plate, and that very young and buoyant crust was the first crust to enter the trench over its entire length (Macpherson & Hall 2001). The formation of IBM boninites due to the subduction of an active spreading axis would thus require that the axis was almost parallel to the proto-IBM trench over a 700 km length, or even 2000 km according to Macpherson & Hall (2001), which is not very plausible. (ii) Kinematic reconstruction (Deschamps & Lallemand 2002) shows that, shortly after initiation of subduction along the proto-Palau–Kyushu Ridge (i.e. IBM arc), northward slab rollback was the main driving force for the opening of the West Philippine Basin between 54 and 33–30 Ma. It is now widely accepted that such a strong trench retreat cannot occur if the subducting crust is very young and, hence, buoyant. The subduction of a young plate and of its active spreading centre would therefore prevent the important trench rollback that is necessary to allow the basin to: (a) open since Early Eocene times; (b) migrate northward as indicated by palaeomagnetic data (e.g. Louden 1976, 1977; Haston & Fuller 1991; Haston et al. 1992; Koyama et al. 1992; Hall et al. 1995; Shipboard Scientific Party 2001); and (c) maintain a significant amount of convergence along the northern margin of the Philippine Sea Plate. Kinematic models (Deschamps 2001; Hall 2001, 2002; Deschamps & Lallemand 2003) show that the West Philippine Basin was surrounded by two subduction zones at the onset of its (rapid) opening, one being located along the Philippine arc and the other along the proto-IBM arc. The opening of the West Philippine Basin is essentially due to the northeastward migration of the proto-IBM trench. This strong rollback of the subducting slab probably requires the subduction of old lithosphere.

The tectonic setting of the Izu–Bonin–Mariana system in the Eocene appears, therefore, to preclude Tertiary subduction of an active spreading centre on the Pacific Plate along the proto-Palau–Kyushu Ridge, suggesting that another mechanism is required to explain the generation of boninitic magmas. The hypothesis of genesis of boninites proposed by Stern & Bloomer (1992) does not account for the necessity of a high heat flow at the time of generation of the boninitic magmas. Secondly, according to this model, boninites would be expected at every subduction zone that is initiated at a boundary between plates of different ages. At present we do not have such evidence. Thirdly, the episode of boninite formation should be short in time (a few million years), but dating shows that it lasted for at least 14 Ma.

- Concerning the hypothesis of formation of boninitic lavas due to the presence of a mantle plume near the proto-IBM arc (Macpherson & Hall 2001), it can explain the formation of the first boninitic lavas erupted along the northern part of the IBM arc before c. 48 Ma. However, it fails to explain the genesis of younger boninitic lavas along the arc. As a matter of fact, the Amami–Oki–Daito region located in the northern part of the Philippine
Sea Plate was influenced by a mantle plume in Middle Eocene times (Hickey-Vargas 1991, 1998; Macpherson & Hall 2001) (Fig. 5). This is supported by geochemical analyses of rocks and by the shallow bathymetry of this region at this epoch. Concerning the West Philippine Basin itself, OIB-like lavas were described only in its northern and western parts. At c. 50 Ma, OIB-like lavas erupted only in the northern Philippine Sea Plate and the 46 Ma reconstruction clearly shows that the area influenced by the mantle plume is restricted to the western part of the plate (Fig. 5). This is demonstrated by: (i) the occurrence of typical N-MORB lavas in the eastern part of the basin; and (ii) the greater depth of the eastern part of the basin, contrasting with the much shallower western part. Moreover, detailed studies of the spreading mechanisms within the basin (Deschamps 2001; Deschamps & Lallemand 2002) show that the spreading system has been strongly disorganized in the western part of the basin due to the presence of an active hot spot during spreading, whereas its eastern part has been formed by normal seafloor spreading. We therefore suggest that an active mantle plume was present beneath the West Philippine Basin in middle Eocene times, but its influence was restricted to the northern part at c. 50 Ma, and then to the western part of the basin at c. 46 Ma and later. From the bathymetry and the distribution of different types of lavas, we propose that the zone of influence of the mantle plume was at most c. 500 km in diameter. An important consequence of this observation is that the boninitic volcanism that occurred after 48 Ma along the proto-IBM arc cannot be explained by the presence of a mantle plume interacting with the subduction zone (Fig. 5).

We have shown that the formation of boninites along the proto-IBM arc can be explained neither by the subduction of a very young and hot plate, nor by the nucleation of subduction along a transform boundary. It can be explained by the presence of a mantle plume in the vicinity of the subduction zone along the proto-IBM arc, but only in early middle Eocene times. We thus need to propose an alternative mechanism explaining the occurrence of boninitic lavas along the IBM arc after 48 Ma.

**Geodynamic setting of boninites lavas**

From our reconstruction, boninites observed in the Bonin Islands were generated near the termination of a volcanic arc, at the transition between a subduction zone and a transform fault (Fig. 5). As represented in our reconstruction, it is very possible that a spreading centre was active in the Minami-Daito Basin near the subduction–transform transition, at the time of boninite formation. This tectonic context appears to be similar to the setting of boninites from North Tonga and the southern New Hebrides (Figs 2, 3). The high heat flow that is necessary for the genesis of boninites may thus have been provided by the hot mantle upwelling beneath the spreading axis, and also by the active mantle plume in the northernmost part of the basin during early middle Eocene times. A slab window that formed due to the subduction–transform transition probably allowed hot mantle material to flow around the exposed edge of the subducting plate. Boninites found in the Zambales ophiolite are shown by our model to be formed by a similar mechanism. Reconstruction at c. 43–46 Ma (time of the boninitic episode) indeed shows that this region was at western termination of the Philippine arc, at the transition between the related subduction and a transform boundary. The eruptions of boninites at sites 458 and MD28, and Guam and Saipan in middle Eocene times are most probably due to incipient subduction along the transform plate boundary (along the proto-Palau–Kyushu Ridge) in the vicinity of a well-established spreading centre that connects at about a right angle with the subduction zone. In this case, the hot mantle upwelling beneath the spreading axis provided high heat flow. Boninites that erupted at c. 40 Ma near Site 458 would also have formed due to the interaction between the hot mantle feeding the active spreading centre and the subduction zone. Boninites erupted at sites 782 and 786 at c. 50 Ma would have originated due to the presence of a mantle plume in the vicinity of the subduction zone bounding the northern Philippine Sea Plate. Concerning boninites that were emplaced at the same location in the middle Eocene, it is possible that these are due to the interaction between hot mantle feeding a minor spreading axis in the Kita–Daito Basin and the same subduction zone. This hypothesis cannot be easily verified because of a diversity of age determination in the Amami–Oki–Daito region and the lack of reliable data. The occurrence of a minor episode of boninitic volcanism along the Palau–Kyushu Ridge at c. 34–35 Ma is still problematic. These boninites are observed as sills and dykes that intrude older volcanic sequences. However, our reconstruction at this epoch does not show any particularity in tectonic setting that can explain the occurrence of
GEODYNAMIC SETTING OF IZU-BONIN-MARIANA BONINITES

Fig. 6. Definition of three types of settings that favour the genesis of boninites lavas, without requiring the influence of a mantle plume. Black circles indicate location of possible boninite eruptions. Type 1 setting is most likely to favour the genesis of boninites along the central part of the Tofua arc. Type 2 setting is probably responsible for the production of boninitic lavas in the Bonin Islands, in the northern termination of the Tofua arc, in the southern termination of the Central Spreading Ridge in the North Fiji Basin and, perhaps, in the Zambales ophiolite. Similar settings are reported from the northern and southern margins of the Caribbean Plate, and near the southern termination of the Andaman Sea spreading ridge. We could thus expect to find boninitic lavas at these locations. Type 3 setting is probably the most unusual one, as this has probably operated for a limited length of time. It is likely to explain the genesis of boninites in the Mariana Islands.

Discussion

The study of the Cenozoic tectonic setting of IBM arc boninite genesis in light of modern geodynamic contexts in Tonga and Fiji regions has led us to define three types of settings that favour the genesis of boninitic lavas, none of which require the influence of a mantle plume in the region (Fig. 6). In all three settings, the activity of a back-arc spreading centre is a key element for formation of boninites.

Type 1 (Fig. 6)

Most back-arc rifts and spreading centres propagate at a low angle (subparallel) to the associated volcanic arcs. The Valu Fa Ridge in the southern part of the Lau Basin is a typical example of a back-arc spreading system slightly
oblique and close to the volcanic arc. It is propagating toward the arc from north to south. Boninites could erupt for a short period when the spreading axis is very close to the active arc. While the basin is opening, the spreading axis moves away from the arc and no more boninitic melts will be formed. However, if the tip of the spreading centre still propagates to the south, one may expect that the same favourable context will also propagate to the south. Similarly, further back in time, the same situation should have occurred more to the north, but no dredging or sampling is available to confirm this hypothesis. Many back-arc basins (Sumisu rift, Mariana Trough, Okinawa Trough, Havre Trough) begin to form along volcanic arcs, as attested by the presence of remnant arcs. Conversely, no boninites were found in these basins. One explanation is that boninites erupted during the rifting phase and were then buried shortly after their emplacement beneath N-MORB basalts when spreading starts. An alternative solution is that rifting conditions (for example, temperatures) are not compatible with boninite generation. One may observe that the Valu Fa Ridge already has the characteristics of a well-developed spreading centre with high heat flow.

Type 2 (Figs 3 and 6)

A favourable context for longer periods of boninite eruption is an intersection at a high angle between an active spreading centre and the transition between a subduction zone and a transform. There are a few examples of ridge segments almost perpendicular with arcs (Andaman Sea, Cayman Trough), all caused by a high obliquity in convergence. In these cases, short ridge segments are offset by long fracture zones more or less parallel to the arc. In other words, when a back-arc spreading ridge intersects a plate boundary, it is likely to be a transform fault, like in northern Tonga, southern New Hebrides or Scotia Sea, and rarely a subduction zone, or if it is, spreading segments are short indicating highly oblique plate convergence. In any case, spreading cannot occur in the forearc domain because the mantle wedge is absent. The subducting plate, indeed, cools the front of the overriding plate. Besides, no modern example of forearc spreading has been yet reported. Interestingly, nature offers us two modern examples where back-arc spreading ridges intersect the transition zone between subduction and strike-slip. At these transitions, the slab edge dips less than the rest of the slab and thus arc volcanism associated with boninite eruptions could occur during a significant amount of time as in northern Tonga (Millen & Hamburger 1998). We believe that a similar context occurred at the elbow along the Palau–Kyushu Ridge that coincides with the Bonin Islands after closure of the Shikoku Basin. We propose that subduction occurred along the palaeo-Izu–Bonin trench, whereas strike-slip occurred along the palaeo-Mariana segment between 55 and 47 Ma. We assume that a slow rift was active over a long period near the Palau–Kyushu Ridge elbow. There is evidence for a plume in this region at that time (Hickey-Vargas 1991, 1998; Macpherson & Hall 2001) that could have triggered either the generation of boninites by itself or by feeding such slow rifts. According to our reconstructions, the Zambales boninites were also in the same geodynamic context of subduction to strike-slip transition at the time of their emplacement (44–46 Ma).

Type 3 (Fig. 6)

Since 47 Ma, the main spreading centre of the West Philippine Basin was far from any subduction zone and intersected a major transform zone that evolved into a subduction zone between 47 and 43 Ma, according to volcanic records (Hussong & Uyeda 1981; Seno & Maruyama 1984; Stern & Bloomer 1992; Taylor 1992; Bloomer et al. 1995) and changes in Pacific Plate kinematics (e.g. Engebretson et al. 1985; Clague & Dalrymple 1989; Norton 1995; Rowley 1996; Gordon 2000; Hall 2002). Over a short period, boninites erupted near the edge of the spreading centre that formerly intersected the transform zone, which was progressively inactivated by incipient subduction of the Pacific Plate. When subduction started along the Mariana Trench, the spreading rate in the West Philippine Basin suddenly decreased from 4.5 to 2 cm a⁻¹ (Hilde & Lee 1984; Deschamps 2001; Deschamps & Lallemand 2002). The transform motion required by the generation of oceanic crust on both sides of the spreading ridge of the West Philippine Basin was probably accommodated behind of the volcanic arc. Genesis of boninites in this type of setting is helped by migration of asthenosphere over the subsiding lithosphere, which would entail adiabatic decompression, and also by the presence of water released from the subducted lithosphere and sediments at a depth of 40 km (Peacock 1990; Stern & Bloomer 1992). The source of extra heat flow is provided by the nearby asthenospheric flow beneath the active spreading ridge in the forming upper plate.
Conclusions

Geodynamic models of the Philippine Sea Plate in Eocene times help us to define the tectonic setting of the Izu–Bonin–Mariana boninite genesis. It appears that in the early Eocene, the interaction between a subduction zone and a mantle plume could have contributed to the formation of boninitic lavas along the northern part of the IBM arc. However, our model clearly shows that the subduction zone along the arc was beyond the influence of this mantle plume since the middle Eocene. The genesis of boninites in the Bonin Islands at this time is thus rather due to the intersection at high angle between an active spreading centre and a subduction–transform fault transition. Modern examples of boninite tectonic settings show that the intersection between an active spreading axis and such a plate boundary is a context favourable to the genesis of boninites. The high heat flow associated with the upwelling of hot asthenospheric mantle beneath the spreading axis probably provides the extra heat that is necessary for the melting of a very depleted mantle, and the dehydration of the subducting slab provides fluids that help melting. The occurrence of a slab window at the location of the tear also probably favours upward asthenospheric flow around the exposed edge of the subducting plate, and hence high heat flow. Synchronous eruption of boninitic lavas within the Zambales ophiolites (Luzon) is likely to be due to a similar mechanism. The boninites that erupted along the central part of the IBM arc in the middle Eocene are probably formed due to the intersection between the active spreading centre of the West Philippine Basin and the transform fault that has accommodated the basin opening, and which was reconverted into a subduction zone at this time.

Finally, the study of the Cenozoic setting of the IBM arc boninites, considered in light of modern geodynamic contexts in the Tonga and Fiji regions, has led us to define three tectonic settings that favour the formation of boninitic lavas in back-arc basins, without requiring the influence of the mantle plume. The first setting is the propagation at a low angle of a spreading centre towards its associated volcanic arc. The second is the intersection at a high angle between an active spreading centre and the transition from subduction to a transform fault, which means the intersection between an active axis and the termination of an active volcanic arc. The third is the intersection at a high angle between an active spreading centre and a subduction zone. This last case of boninitic volcanism should be quite rare, as such a tectonic setting is probably limited in time. The definition of these three tectonic settings that favour the formation of boninitic lavas implies that such volcanic products should be found each time that an active spreading axis closely interacts with an active volcanic arc, or its termination near a transform boundary. We can then expect to find boninites at the southern termination of the Mariana Basin, where the active spreading axis approaches the southern termination of the active Mariana Arc. In the same way, boninites could be found along the border of the Andaman Sea, where the central valley (or Central Andaman Trough) approaches at a high angle to the active Andaman arc near the Nicobar Islands. The Caribbean–South America plate boundary could also be a context favourable to the genesis of boninites, at the southern termination of the Granada back-arc basin where the Lesser Antilles arc swings into a transform boundary. For some reasons, boninites could also occur at the northern end of the same arc, near the Virgin Islands.

We thank Y. Tatsumi and Y. Tamura for valuable discussions about conditions of boninitic lavas genesis. This contribution has been improved by constructive reviews by R. Taylor and J. Bédard. Some maps were made using Wessel & Smith (1995) GMT 3.1 software. Smith & Sandwell’s (1997) satellite-derived bathymetric data were used to draw some maps.

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