Geochemistry and tectonic significance of basalts in the Poya Terrane, New Caledonia

Jean-Philippe Eissen a,*, Anthony J. Crawford b, Joseph Cotten c, Sébastien Meffre b, Hervé Bellon c, Mireille Delaune d

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(Text continued on inside back cover)
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Abstract

The Norfolk–New Caledonia Ridge represents a continental slice which drifted away from Australia during the Late Cretaceous breakup of the eastern Gondwana margin. The presence of widespread basaltic rocks beneath the main ophiolite nappe of New Caledonia has been long known but the origin and the age of the Poya Terrane basalts (PTB herein) remained controversial. Recent palaeontologically determined ages date the PTB as Late Cretaceous (Campanian). New geochemical data show that two main discrete groups constitute the PTB: a MORB-like tholeiitic suite, and a more alkaline intra-plate basaltic suite distinguished mainly on immobile HFSE and REE elements. Furthermore, low εNd and high Th/Nb relative to MORB, and weak negative Nb anomalies, reflect limited assimilation of continental crust by these otherwise MORB-like tholeiites. Inter-PTB sedimentary rocks all have a pelagic or hemipelagic origin; detrital material originated from the nearby Norfolk–New Caledonia ridge basement. The PTB form a parautochthonous sheet below the main harzburgitic nappe constituting the New Caledonian ophiolite. They are genetically unrelated to the ophiolite, and are interpreted to be 70–85-Ma-old rift tholeiites formed during of the easternmost continental part of Mesozoic Gondwana, and opening the East New Caledonia Basin. The Norfolk–New Caledonia Ridge formed the western passive margin of this new oceanic basin, but the rifted-off eastern block is less easily identified. It may form part of the basement of the Western Belt of the New Hebrides island arc (Vanuatu). The cessation of rifting of the eastern Australian margin around 56 Ma was followed by an eastward-directed subduction which produced boninitic melts and its associated refractory harzburgitic mantle, in the forearc of the primitive Loyalty–d’Entrecasteaux arc. Following the major Pacific plate motion reorganization around 42 Ma, collision of the Norfolk–New Caledonia Ridge with the forearc region of the intra-oceanic Loyalty–d’Entrecasteaux arc around 40 Ma led first to westward thrusting of the PTB as a slice picked up from the upper crustal section of the colliding Norfolk Ridge. Subsequent collisional tectonism led to detachment of the main New Caledonian harzburgitic nappe from its forearc location in the Loyalty arc, and westward emplacement of this nappe over the PTB nearby allochthon. The presence of parautochthonous sheets of basalts unrelated to immediately overlying forearc-derived, boninite-bearing harzburgitic ophiolites is briefly discussed in the light of two other examples in arc–continent collision settings. © 1998 Elsevier Science B.V. All rights reserved.

Keywords: SW Pacific; New Caledonia; Cretaceous; Eocene; allochthonous terrane; geodynamic significance; MORB; backarc basalts; intra-plate basalts; boninites; ophiolites; geochemistry

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1. Introduction

New Caledonia is a microcontinental island which drifted away from Australia during the Late Cretaceous breakup of the eastern Gondwana margin and opening of the Tasman Sea and New Caledonia Basin (Paris, 1981; Kroenke, 1984; Mignot, 1984; Rigolot, 1989; Cluzel et al., 1994). It is constituted by rocks which record at least two arc–continent collisions, the first in the early Mesozoic, the more recent in the mid-Eocene (Cluzel et al., 1994; Aitchison et al., 1995a; Meffre et al., 1996). The latter collision resulted in SSW-directed emplacement of a massive harzburgite-dominated nappe probably at least 6 km thick (Fig. 1) (Avias, 1967; Guillon, 1975; Prinzhofer et al., 1980). Although this sheet appears to be broadly continuous with crust of the adjacent South Loyalty Basin to the east of New Caledonia,

Fig. 1. (a) Location of New Caledonia in the SW Pacific. DEZ = d’Etrecasteaux Zone; SLB = South Loyalty Basin. (b) Simplified geological map of New Caledonia showing the location of the Poya Terrane basalts along the western coast between Bourail and Koumac. Ne = Népoui. Sample sites: G = Gomen; O = Ouaco; Tn = Taen; Bo = Boyen; V = Voh; Tr = Temala; P = Pinjen peninsula; K = Kone; P = Poya (including samples from a 150-m-deep drilling done by the BRGM on the Honfleur sulphide deposit).
(Collot et al., 1987), there are no known occurrences of volcanics associated with the nappe on New Caledonia, and magmatic rocks are restricted to limited occurrences of cumulates and occasional dolerite dykes (Prinzhofer et al., 1980; Dupuy et al., 1981).

The presence of widespread basaltic rocks beneath the New Caledonia ophiolite nappe has been long known and well documented (Routhier, 1953; Lillie and Brothers, 1970; Guillou, 1972; Guillou and Gonord, 1972; Rodgers, 1975; Parrot and Dugas, 1980; Paris, 1981; Kroenke, 1984; Maurizot et al., 1985; Black, 1993; Cluzel et al., 1994; Aitchison et al., 1995a,b). Marine geophysical data indicate that this basaltic formation, as well as the main harzburgite nappe, extend 300 km northwestward in the basement of the northern lagoon of New Caledonia (Collot et al., 1988). Guillon (1972) first recognized that these basaltic rocks form an extensive nappe, outcropping over much of the west coast of New Caledonia, where these rocks have been called the 'Formation des Basaltes de la Côte Ouest' (West Coast Basalts Formation). More recent studies (Cluzel et al., 1994; Meffre, 1995; Aitchison et al., 1995a) have confirmed that the West Coast Basalt Formation represents the western part of a 20–500-m-thick nappe, which extends over much of the island as a slice beneath the main harzburgite nappe. Basalts of similar petro-geochemical characteristics outcrop also locally along the east coast of New Caledonia as a thin slice beneath the harzburgites (Cluzel et al., 1994, 1995; Meffre, 1995). This basaltic nappe has been termed the Poya Terrane (Cluzel et al., 1994; Aitchison et al., 1995a). Ongoing controversy surrounds the affiliations, origin and tectonic significance of the Poya Terrane basalts (PTB herein).

2. Poya Terrane

The Poya Terrane was emplaced as a relatively thin thrust sheet along with but beneath the main harzburgite sheet of the New Caledonia ophiolite, between 38 and 46 Ma (Cluzel et al., 1994, 1995; Aitchison et al., 1995a; Meffre, 1995). The direction and sense of its emplacement was determined by careful examination of the geological contacts, a kinematic analysis of the normal faults, the polarity of the metamorphism which affects the PTB, which increases northward along the east coast (Cluzel et al., 1994, 1995; Meffre, 1995) and its strong link with the emplacement of the main harzburgite sheet from the north-northeast (Guillon, 1975; Prinzhofer et al., 1980; Collot et al., 1987).

Poya Terrane sequences are dominated by low-grade metamorphosed, often tectonized, pillow basalts, with associated hyaloclastites, fine-grained tuffaceous sediment and calcareous sediments known as the ‘Koné facies’ (Carroué, 1972), radiolarian cherts and more locally massive basalts, gabbros and dolerites. Occasional sheeted basaltic flows and serpentinites have also been observed.

The age of the Poya Terrane has been a controversial issue, with arguments being presented in the literature for Eocene–Paleocene and/or Cretaceous ages (Routhier, 1953; Espirat, 1963; Coudray and Gonord, 1967; Carroué, 1972; Guillou, 1972; Cluzel et al., 1994; Aitchison et al., 1995a; Meffre, 1995). Palaeontological ages range between 88 and 45 Ma (Coudray and Gonord, 1967; Carroué, 1972). However, most cherts clearly interbedded with the basalts have yielded Radiolaria indicating a Campanian (73–83 Ma) age (two new ages from Meffre, 1995 and ten from Aitchison et al., 1995b). Several fossiliferous Senonian intercalations (66–88 Ma) in the Koné facies were also described by Espirat (1963) and Carroué (1972), and Paris (1981) associated a Turonian to Santonian (83–90 Ma) *Inoceramus* macrofauna with the same facies.

Most palaeontologically determined Eocene ages for sequences containing basalts come from the area

Published K–Ar ages for PTB are 38.5 ± 1.5 Ma, 42 ± 2 Ma, 51 ± 7 Ma and 59 ± 6 Ma (Guillon and Gonord, 1972). We have obtained 9 new K–Ar age determinations for low-grade metamorphosed PTB and these range from 39.7 ± 2.1 Ma to 61.6 ± 2.8 Ma (Table 2). These new K–Ar age determinations were performed at the University of Brest by H. Bellon and J.C. Philippet following a procedure described by Bellon and Rangin (1991). Three groups can be recognized: (1) between 39.7 and 43.8 Ma; (2) between 47 and 49 Ma; and (3) with a single isotopic age of 61.6 Ma. The latter age represents the average of two dates determined on separate granulometric fractions of the same sample leached by acetic acid to eliminate the alteration phases (an age of 56.0 Ma was obtained on the same sample without the leaching procedure). All these K–Ar ages are anomalously low compared to the palaeontological ages of the few sedimentary rocks closely associated with the volcanic rocks, presumably reflecting varying extents of resetting, in part probably related to emplacement and subsequent metamorphism beneath the over-riding harzburgite nappe.

The geodynamic significance of the PTB gave rise to various interpretations. The first of these (e.g. Avias, 1967; Challis and Guillon, 1971; Cameron, 1989) argues that the PTB are the volcanic carapace of the main New Caledonia ophiolite. The second (Espirat, 1966, 1971; Guillon, 1972; Guillon and Gonord, 1972; Avias and Coudray, 1975; Gonord, 1977; Aitchison et al., 1995a; Meffre et al., 1996) proposes that PTB represent allochthonous crust of a Late Cretaceous backarc basin. Black (1993) referred to the PTB as ‘N-MORB basalts with some arc affinities’. The autochthonous or subautochthonous origin proposed by other authors (Routhier, 1953; Paris, 1981; Kroenke, 1984; Maurizot et al., 1985) is now known to be wrong. Therefore, the allochthonous models will be evaluated following presentations of our new geochemical data.

3. Associated sediments

Relatively few sedimentary rock types have been identified in close association with the PTB, most being inter-pillow sediments or local thin intercalations of pelagic or hemi-pelagic origin (Paris, 1981; Cameron, 1989; Meffre, 1995). They include (Table 1): (1) brown and dark red siliceous rocks (cherts, jaspers) occasionally containing Radiolaria, which, although generally poorly preserved, have yielded Campanian (73–83 Ma) age determinations (Meffre, 1995; Aitchison et al., 1995b); (2) white or pink micritic limestone, containing some Globigerina and rarely some Radiolaria; (3) altered hyaloclastite with green jasper fragments; (4) clast-supported breccias with dominant volcanic fragments (autobrecciated pillow lavas during of shortly after their formation or ocean floor fault scree) and siliceous or calcareous cement (calcarenite); (5) detrital rocks with clay–siliceous cement (distal turbidite type or greywacke); and (6) tuffaceous mudstone and siltstone.

The non-volcanic detrital fraction of the coarser-grained breccias appears to be derived from Palaeozoic formations of the New Caledonia basement. Despite the different types defined, no clear variations are observed within the whole formation or between the different petro-geochemical groups defined below.

4. Petrography

Our new petrographic and geochemical data (Tables 2 and 3) show that two discrete magmatic groups constitute our PTB: a MORB-like tholeiitic suite, and a more alkaline intra-plate basaltic suite. The very poor outcrop has prevented determination of mutual contact relationships between these groups.

The Poya Terrane MORB-like basalts range from subophitic-textured, almost aphyric massive flows and microgabbro plugs and dykes, through intergranular- and intersertal-textured basalts with 2–10 modal% of plagioclase and augite phenocrysts, to occasional olivine + plagioclase-phryic or plagioclase + augite-phryic pillow basalts with largely devitrified glassy rims. Except in a few unaltered glassy pillow rims (Table 4) in which fresh olivine is preserved...
### Table 1

**Summary of the lithology and origin of the studied samples of the Poya Terrane collected along the New Caledonia west coast between Bourail and Koumac**

<table>
<thead>
<tr>
<th>Sample number</th>
<th>Lithology</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bo-5</td>
<td>Inter-pillow greywacke</td>
<td>Seashore near Pouaco</td>
</tr>
<tr>
<td>Bo-7</td>
<td>Inter-pillow greywacke</td>
<td>Seashore near Pouaco</td>
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<tr>
<td>Bo-9</td>
<td>Inter-pillow limestone</td>
<td>Seashore near Pouaco</td>
</tr>
<tr>
<td>Bo-13</td>
<td>Gabbro</td>
<td>Boyen river</td>
</tr>
<tr>
<td>Bo-16</td>
<td>Dolerite</td>
<td>Boyen river</td>
</tr>
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<td>Bo-33</td>
<td>Sediment breccia</td>
<td>Boyen river</td>
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<tr>
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<td>Boyen river</td>
</tr>
<tr>
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<td>Boyen river</td>
</tr>
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<td>Bo-40</td>
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<td>Boyen river</td>
</tr>
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<td>G-1</td>
<td>Basaltic pillow</td>
<td>Gomen Point</td>
</tr>
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<td>G-2</td>
<td>Inter-pillow jasp with lamination</td>
<td>Gomen Point</td>
</tr>
<tr>
<td>G-4</td>
<td>Basaltic pillow</td>
<td>Gomen Point</td>
</tr>
<tr>
<td>G-5</td>
<td>Basaltic pillow</td>
<td>Gomen Point</td>
</tr>
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<td>Basaltic pillow</td>
<td>Foué Point Quarry</td>
</tr>
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<td>K-3</td>
<td>Pink siliceous limestone</td>
<td>Foué Point Quarry</td>
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<td>K-7</td>
<td>Hyaloclastite</td>
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<td>Foué Point Quarry</td>
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<tr>
<td>K-9</td>
<td>Pink micritic limestone with glass fragments</td>
<td>Foué Point Quarry</td>
</tr>
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<td>K-10</td>
<td>Pink micritic limestone</td>
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<td>Pink micritic limestone</td>
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<td>K-13</td>
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<td>Seashore south of Fousou</td>
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<td>Inter-pillow micritic limestone</td>
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<td>Greywacke</td>
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<tr>
<td>O-2</td>
<td>Basaltic pillow</td>
<td>Quarry bet. RT 1 – coast</td>
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<td>O-6</td>
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<td>Quarry bet. RT 1 – coast</td>
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<td>O-7</td>
<td>Basaltic pillow</td>
<td>Quarry bet. RT 1 – coast</td>
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<tr>
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<td>Inter-pillow calcarenite</td>
<td>Quarry bet. RT 1 – coast</td>
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<td>Inter-pillow calcarenite</td>
<td>Quarry bet. RT 1 – coast</td>
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<td>RT1 near Poya</td>
</tr>
<tr>
<td>P-3</td>
<td>Green/grey jasp leuse in basalt</td>
<td>RT1 near Arangus</td>
</tr>
<tr>
<td>P-5</td>
<td>Basaltic pillow</td>
<td>RT1 near Arangus</td>
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<td>P-7</td>
<td>Yellow siliceous sediment</td>
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</tr>
<tr>
<td>P-8</td>
<td>Red jasp</td>
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<td>Greywacke breccia</td>
<td>Pinjen peninsula</td>
</tr>
<tr>
<td>Pi-5</td>
<td>Inter-pillow pink bio-micrite</td>
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### Table 1 (continued)

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<th>Sample number</th>
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<tr>
<td>Fi-8</td>
<td>Inter-pillow red jasp</td>
<td>Pinjen peninsula</td>
</tr>
<tr>
<td>Fi-9</td>
<td>Olivine phycric altered alkali basalt</td>
<td>Pinjen peninsula</td>
</tr>
<tr>
<td>Fi-11</td>
<td>Jasp</td>
<td>Pinjen peninsula</td>
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<tr>
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<td>Basaltic pillow</td>
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<td>Honfleur Quarry drill core</td>
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<td>G = Gomen; O = Ouaco; Ta = Taom; Bo = Boyen; V = Voh; Te = Temala; Fi = Pinjen peninsula; K = Kone; P = Poya; FT1 = core from the drill done by the BRGM in the Honfleur sulphide deposit near Poya. RT1 = Territorial Road number 1.</td>
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<td>a Fresh glass used for microprobe analyses.</td>
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(For3–85), sparse olivine phenocrystals are replaced by chlorite, or chlorite and calcite. Plagioclase is almost always albitized. Metamorphic assemblages are either prehnite ± pumpellyite facies or chlorite ± actinolite assemblages indicating lowest greenschist-grade burial degradation of ocean-floor type. It is noteworthy to remark that the underlying Mesozoic–Paleocene sedimentary rocks have not been affected by this metamorphic episode (Gonord, 1977; Paris, 1981; Cluzel et al., 1994; Meffre, 1995). The alkaline lavas are volumetrically much less abundant, are always notably more altered than the MORB suite, and occur in localized outcrop areas rarely larger than 2 km in diameter. They are mainly strongly vesicu-
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Table 2
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Note: The table includes various chemical compositions of volcanic rocks, with columns for different elements and their respective concentrations. The data are presented in a tabular format, showing the percentage composition for each chemical element across different samples and lava types.
Table 3
Bulk rock analyses of the studied volcanic samples (continued)

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Total
100.02 | 100.03 | 100.02 | 100.09 | 100.01 | 100.02 | 100.03 | 100.04 | 100.08 | 100.09 | 99.95 | 99.95 | 99.86 | 99.45 | 99.07 | 100.07 |

Analysts: Phil Robinson at the University of Tasmania, Hobart, Australia (major, traces, and REE by XRF). * Joseph Cotten at University of Bretagne Occidentale, Brest, France (major, traces and REE by ICP-AES), and # Helen Waldeon at Becquaret Laboratories, Menai, Australia (REE, Hf, Th, and Tm by NAA). BD = below detection limit. MORB1 = N-MORB; MORB2 = Boyen doleritic MORB suite; MORB3 = MORB with BABB affinity; alkali = alkali suite (see text for explanations). 49K/40Ar whole-rock dates: (a) average of two dates; (b) average of two dates done on two separate granulometric fractions leached by acetic acid to eliminate alteration phases. All analyses (isotopic composition of argon by mass spectrometry using isotopic dilution and potassium by atomic absorption spectrophotometry) were performed at the University of Bretagne Occidentale by Hervé Bellon and Jean-Claude Philippe.
lar pillow basalts with common olivine phenocrysts replaced by hematite and calcite, and occasional albitized plagioclase phenocrysts. Abundant vesicles are filled by calcite.

5. Analytical methods

The samples analysed were finely powdered in an agate mill. Major and trace elements analysed in Brest, except for Rb, were measured by ICP-AES with an ISA Jobin-Yvon® JY 70 Plus apparatus. Rb was measured by flame atomic emission using a Perkin-Elmer® 5000 spectrometer. The procedure of solution preparation was presented by Cotten et al. (1995). All the elements were determined from one solution without selective extraction. Calibrations were made using international standards (JB2, BEN, ACE, Mica-Fe) as well as specific references samples. Relative standard deviations are ≤ 2% for major elements and ≤ 5% for trace elements. Detection limits are Rb 0.5 ppm, Sr 0.5 ppm, Ba 2 ppm, Sc 0.2 ppm, V 3 ppm, Cr 2 ppm, Co 2 ppm, Ni 2 ppm, Y 0.5 ppm, Zr 2 ppm, Nb 1 ppm, La 0.8 ppm, Ce 2 ppm, Nd 2 ppm, Eu 0.2 ppm, Dy 0.3 ppm, Er 1 ppm and Yb 0.1 ppm. XRF major and trace elements analyses were performed at the University of Tasmania using an automated Philips® PW 1410 spectrometer. Major elements were measured with Rh tube, Sc, V, Cr, Sr, Zr, Nb and Ba using a Au tube and Ni, Rb and Y a Mo tube. Trace elements were determined using mass absorption coefficients calculated from major element analyses. Major elements were analysed using fused discs, following the method of Norrish and Hutton (1969). Loss on ignition was measured as weight percent loss of 1 g of powdered sample heated to 1000°C for 12 h, followed by 5 h at 400°C. Trace elements were analysed on 6-g pressed powder pellets coated with boric acid, using the method of Norrish and Chappell (1977). The REE data were measured using the ion separation XRF technique described by Robinson et al. (1986). INAA data were obtained at Becquerel Laboratories, Lucas Heights Research Laboratories, New South Wales (analyst Helen Waldron). Detection limits are Hf 0.5 ppm, Ta 0.2 ppm and Th 0.2 ppm.

6. Geochemistry

We have analyzed 36 basalts from outcrops along the length of the Poya Terrane between Bourail and Koumac on the west coast of New Caledonia (Fig. 1). Also available for consideration are eleven analyses of MORB-like basalts from this same area, reported by Cameron (1989). All analyzed PTB are basaltic compositions, and despite careful sample selection and analytical procedures, loss on ignition values are typically 1–4% for the MORB-like suite, and usually 5–8% for the amygdaloidal alkaline suite. Thus we consider it unlikely that the measured alkali- and K-group element abundances are pristine, and we rely mainly on immobile HFSE and REE elements for determining the affinities of these basalts.

Table 4
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<td>15.42</td>
<td>14.63</td>
<td>14.58</td>
<td>15.04</td>
</tr>
<tr>
<td>FeO*</td>
<td>8.86</td>
<td>9.15</td>
<td>11.08</td>
<td>11.01</td>
<td>10.02</td>
</tr>
<tr>
<td>MnO</td>
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<td>0.16</td>
<td>0.15</td>
<td>0.17</td>
<td>0.16</td>
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<tr>
<td>MgO</td>
<td>9.331</td>
<td>8.97</td>
<td>7.74</td>
<td>7.77</td>
<td>8.45</td>
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<tr>
<td>CaO</td>
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<td>12.15</td>
<td>11.82</td>
<td>11.75</td>
<td>12.07</td>
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<tr>
<td>Na₂O</td>
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<td>2.50</td>
<td>2.72</td>
<td>2.55</td>
<td>2.54</td>
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<tr>
<td>K₂O</td>
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<td>0.05</td>
<td>0.15</td>
<td>0.15</td>
<td>0.09</td>
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<tr>
<td>P₂O₅</td>
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<td>0.00</td>
<td>0.17</td>
<td>0.00</td>
<td>0.04</td>
</tr>
<tr>
<td>Total</td>
<td>99.92</td>
<td>99.65</td>
<td>99.68</td>
<td>99.19</td>
<td>99.59</td>
</tr>
</tbody>
</table>

Analysts: Jean-Philippe Eissen and Marcel Bohn. The entire set of glass microprobe analyses is available upon request from the senior author of this paper.
The alkaline basalts have Nb contents at least twice those of the MORB-like suite at similar stages of fractionation, and all have Zr/Nb values between 5 and 6 (Fig. 2; ‘alkali’ samples of Tables 2 and 3). N-MORB-normalized element variation patterns for immobile elements (Fig. 3) for this suite show significant enrichment in the more incompatible elements, and fall between those of enriched ridge-generated MORB (E-MORB) and ocean island basalt (OIB). We interpret these basalts to have derived from seamounts formed on ocean crust. They may be essentially in situ on MORB-like PTB, or they may have been scraped off their original oceanic substrate during subduction, and emplaced onto PTB MORB-like basalts during emplacement of the PTB as an allochthonous slice beneath the main harzburgitic ophiolite.

The MORB-like basaltic suite shows a limited fractionation range (36–103 ppm Zr), and has TiO₂ contents from 1.0–1.8%; ferrobasalts (FeO²⁺>14%) are unrecorded. The majority of these basalts have of Ti/Zr (87–137) and La/Nb (1.09 ± 0.12), values characteristic of N-MORB. However, Zr/Nb values (20 ± 4) are slightly lower than for N-MORB (32–40), due to very slightly higher Nb values of the PTB compared to N-MORB at similar stages of fractionation (Fig. 2). Some dolerites which outcrop mainly along the Boyen River or near Koné, even have slightly lower Zr/Nb values (12 ± 2), but their spidergrams are very similar to those of the main basaltic lava group (MORB 2 of Table 2). Despite this, N-MORB-normalized element variation patterns show weak negative Nb anomalies and have Th/Nb values (0.09 and 0.12 for the two samples analyzed) significantly higher than values for N-MORB (0.03–0.08: our average of 51 samples from the literature is 0.066 ± 0.015). Chondrite-normalized REE patterns show only slight LREE-enrichment, and flat HREE (Fig. 4). However, at least two basalts of the MORB-like suite show more depleted REE patterns and a slight but significant negative Nb anomaly (Fig. 3) relative to adjacent La as observed for BABB (MORB 3 of Tables 2 and 3); however, these more depleted basalts have Zr/Nb~50, values characteristic of LREE-depleted N-MORB.

Cameron (1989) reported Nd–Sr isotopic data for three PTB with a Zr/Nb of 18–22, for which initial εNd values (recalculated to 80 Ma) are +3.1, +3.5 and +4.6, significantly below values for N-MORB of this age (+8). We suggest that the trend to low εNd and high Th/Nb relative to MORB, and the weak negative Nb anomalies, reflect limited assimilation
II I Il II
Sea of Japan
ODP 797 units 1-5
ODP 797 units 10-14
ODP 794 units 1-9
PTB Averages
N-MORB suite
MORB dolerites suite
MORB/BABB suite

Fig. 3. Variation diagrams of N-MORB-normalized elements showing representative patterns for: (a) averages of the PTB N-MORB suite, the PTB Boyen dolerites and the PTB MORB/BABB in comparison with selected patterns from three different units of ODP Site 797 (Sea of Japan; data from Allan and Gorton, 1992 and Pouclet and Bellon, 1992); (b) three lavas for the Poya Terrane basalts alkaline suite. N-MORB, E-MORB and OIB values after Sun and McDonough (1989).

of older (Mesozoic or Palaeozoic?) continental crust during eruption of these otherwise MORB-like rift tholeiites. We note that backarc basin basalts from stable intra-oceanic backarc basins such as the southern and central parts of the North Fiji Basin and the northern Lau Basin have $\varepsilon_{Nd}$ values normally $>7$ and Th/Nb $<0.08$ (Jenner et al., 1987; Volpe et al., 1988; Auzende et al., 1990). In contrast, BABB generated during rifting of older continental margin arc crust (e.g. Sea of Japan) have $\varepsilon_{Nd}$ values between 3 and 7, and Th/Nb values averaging $0.120 \pm 0.04$ for Sea of Japan ODP Site 794 (sixteen samples) and $0.273 \pm 0.06$ for the lower basalts in ODP Site 797 (eight samples) (Pouclet and Bellon, 1992; Nohda et al., 1992; Cousens and Allan, 1992). Basalts in the Okinawa Trough, a backarc rift in the thinned continental margin crust of eastern China, have $\varepsilon_{Nd}$ values from $+2.3$ to $+4.7$ (Chen et al., 1995), but Th/Nb values are unavailable.

7. Discussion

Our new data for the PTB indicate that they are an allochthonous slice of tholeiites produced during rifting at $\sim 70$–$80$ Ma along the eastern margin of the Norfolk–New Caledonia Ridge. The PTB cannot be related to the main New Caledonian harzburgitic ophiolite because of the very refractory nature of the latter (see below). The PTB are interpreted instead as an allochthonous slice below the ophiolite, picked up by the ophiolite and transported shortly south-westward along the base of the ophiolite during its emplacement in the Upper Eocene.

The major tectono-stratigraphic unit in New Caledonia is the massive harzburgitic ophiolite, for which the cumulate carapace is volumetrically insignificant and directly associated lavas remain unknown. The very refractory nature of the dominant harzburgitic tectonites (Prinzhofer et al., 1980) indicates that equilibrium melts must have been depleted, second-stage melts, not MORB-type tholeiites such as the PTB, which would have been in equilibrium with lherzolitic residues (Falloon et al., 1989). Lavas chemically appropriate for equilibrium melts with the harzburgitic tectonites are the low-Ca boninites that crop out in several restricted areas, probably as blocks within a serpentinite melange, immediately beneath the ophiolitic harzburgites in the Nepoui area of western New Caledonia (Fig. 1b). These boninites contain Fo$_{94}$ olivine phenocrysts and extremely refractory Cr-rich chromites ($Cr/(Cr+Al) = 0.84$ to 0.88; Cameron, 1989). We suggest that these boninites are part of the liquid complement of the ophiolite represented by the harzburgites, and that slices of the carapace of the ophiolite were over-ridden by massive harzburgite allochthons during emplacement.

Analogous to the Bonin–Mariana forearc, the New Caledonian boninites and their overlying
harzburgitic residues may represent forearc basement to an oceanic island arc represented now by the basement of the Loyalty Islands, presently capped by Quaternary coral reefs (Maillet et al., 1983; Collet et al., 1987; Aitchison et al., 1995b). Strongly altered lavas of intraplate affinity (K/Ar age of 9–11 Ma) outcrop in a restricted area of the island Mare (Baudron et al., 1976; Monzier, 1993) and OIB and comendites (K/Ar age of 30–33 Ma) have been described in the submarine basement of the island Mare between water depths of 4000 and 5200 m associated with backarc basin basalts (Monzier et al., 1989; Monzier, 1993).

However, the Loyalty Islands are the emergent part of an 1100-km-long ridge that includes 12–15 seamounts; the islands are ~150 km east of the leading edge of the main harzburgitic ophiolite nappe in New Caledonia, and are spaced ~70 km apart, similar to the spacing for major volcanoes in the Mariana arc (50 km; Bloomer et al., 1989) and the southern part of the New Hebrides arc (90 km; Macfarlane et al., 1988). Where the Loyalty Ridge swings northeasterward into the d’Entrecasteaux Ridge, seamounts on the latter are known to be composed of primitive island arc tholeiites (Coltorti et al., 1994; Baker et al., 1994), strongly supporting the suggestion that the Loyalty Ridge marks an intra-oceanic arc. Arc-derived volcaniclastics of Mid-Eocene age drilled in...
the North Loyalty Basin at DSDP Site 486 (Andrews et al., 1975) can be traced seismically to lap onto the northern Loyalty Ridge seamounts (Meffre, 1995).

The cessation of magmatism on the arc volcano seamounts of the South d’Entrecasteaux Chain has been dated at around 38 Ma (Quinn et al., 1994; Baker et al., 1994), and in Hole 486, a marked decrease in volcaniclastic sedimentation in the Late Eocene suggests that arc magmatism may have ceased on the Loyalty arc around 38 Ma (Andrews et al., 1975). The North Loyalty Basin contains magnetic lineations, oriented N70° to N80°, which have been interpreted as anomalies 18 to 23 (42 to 55 Ma; Lapouille, 1982; Weissel et al., 1982).

We suggest that the eastern margin of the Australian section of Gondwana was fragmented by progressive, subparallel rifting episodes around 50-75 Ma, with the creation of true ocean crust and magnetic anomalies around 73 Ma in the Tasman Sea and around 75 Ma in the New Caledonia Basin. This rifting created, from west to east, the Tasman Sea, the Lord Howe Rise, the New Caledonia Basin, and the Norfolk–New Caledonia Ridge. That ribbon of continental crust represented by the basement of the Norfolk Ridge (Green, 1978) is now only exposed in the core of New Caledonia, where it is composed of a Permo–Triassic arc-related collage (Paris, 1981). The nature of the freeboard east of the Norfolk Ridge at this time is unknown, but we note that no Paleocene or late Mesozoic subduction-related volcanics are known along the SW Pacific margin, implying that the eastern margin of the pre-Eocene Australian plate was not an active subduction margin.

We have argued above that the PTB represent rift magmatism produced during the breakup stage of development of a small ocean basin. In our preferred tectonic model, the PTB represent a slice of the near-breakup western passive margin of a small ocean crust-floored basin, here called the East New Caledonia Basin, that began to form around 85 Ma (Fig. 5a). Later east-directed subduction of the ocean crust of the East New Caledonia Basin led first to the closure of the latter followed by the collision of the western passive margin of the ERast Caledonia Basin with the forearc of the Loyalty oceanic arc (see later) which emplaced the main New Caledonian ophiolite sheet from the north-northeast back onto this passive margin (Fig. 5b). If this is correct, there must be a conjugate passive margin representing the eastern side of the East New Caledonia Basin. We suggest that the remnants of this easternmost ribbon of continental crust that rifted eastward from the Norfolk–New Caledonia Ridge during opening of the East New Caledonia Basin, and later of the South Fiji Basin, is now located in the basement of the Western Belt of the New Hebrides (Vanuatu) island arc (Fig. 1), and possibly also in the basement of the Fiji platform, both of which evolved into the South Fiji Basin (Fig. 5b).

In this model, there are two conceivable possibilities for the origin of the North Loyalty Basin. It may simply be a remnant of the youngest part of the East New Caledonia Basin, which continued to spread until at least 42 Ma following cessation of rifting in the Tasman Sea and New Caledonia Basin at ~56 Ma (Fig. 5b). But this hypothesis is in contradiction with the NE-SW orientation of the magnetic lineations (Lapouille, 1982; Weissel et al., 1982). Alternatively, the North Loyalty Basin may have been a backarc basin of the Loyalty–d’Entrecasteaux arc which evolved eventually later into the South Fiji Basin after the formation of the Vitiaz–Lau–Colville arc (proto-New Hebrides–Tonga–Kermadec

Fig. 5. Schematic geodynamic reconstruction of New Caledonia and the SW Pacific around ~70 Ma (a), near 56 Ma (b), between 50 and 45 Ma (c), and around 40–38 Ma (d). A = Australia; ENCB = East New Caledonia Basin; LHR = Lord Howe Rise; NC = New Caledonia; NCB = New Caledonia Basin; NLB = North Loyalty Basin; NR = Norfolk Ridge; NZ = New Zealand; PVTS = Proto Vitiaz Tonga Subduction; SFB = South Fiji Basin; TS = Tasman Sea. See text for explanations.
O Loyalty arc volcano  
\[ \Rightarrow \] NC peridotites obduction (~38 Ma)  
D Fracture zone  
\[ \rightarrow \] Active spreading centre  
\[ \] Submerged continental crust or ridge  
\[ \] Continental crust  
\[ \] Location of future PTB  
\[ \] Location of future NW New Zealand nappe  
\[ \] Extinct spreading centre  
\[ \] Continental rifting  
\[ \] Active subduction  
\[ \] Extinct/aborted subduction  
\[ \] Location of future NC peridotites nappe  
\[ \] And its associated boninites  
73 Age (Ma) of start/end of spreading  
\[ \] Direction of plate motion
arc). However, this demands the existence of an early subduction and arc formation at least around 55 Ma, for which there is no evidence. DSDP drilling in the North Loyalty Basin penetrated only open ocean, pelagic sediments of Eocene age, despite the proximity to the Loyalty–d’Entrecasteaux arc volcanoes.

Following cessation of rifting in the Tasman Sea, New Caledonia Basin and Eastern New Caledonia Basin at ~56 Ma (Fig. 5b), continued extension at the eastern margin of the Australian plate was taken up by opening of the North Loyalty Basin, in which the oldest magnetic anomalies are ~56 Ma, which evolved later into the South Fiji Basin. This opening combined with the general northward displacement of the Australian plate, presumably generated regional more or less NE–SW compression, which we believed initiated subduction along the spreading ridge in the East New Caledonia Basin spreading centre (Fig. 5c). This slow subduction, NE deepening, generated boninitic lavas from the hot young lithospheric wedge of the just extinct spreading centre and explains the very refractory nature of the associated harzburgitic crust in the forearc position of the nascent Loyalty–d’Entrecasteaux arc, with the mantle now exposed in the New Caledonia ophiolite. Boninite generation requires abnormally high temperatures at relatively shallow levels in the upper mantle (Crawford et al., 1989), and subduction initiation may well have been focussed on the thermally weakened, only recently extinct, spreading centre in the East New Caledonia Basin. Continued subduction, albeit relatively short-lived, produced the primitive arc volcanoes of the Loyalty–d’Entrecasteaux arc.

The major reorganization of plate motion in the Pacific region around 42 Ma, best shown by the bends in the intraplate island chains in the Pacific Ocean such as that in the Hawaiian–Emperor chain, led to a NW-directed motion for the Pacific plate. This reorganization also affected the Australian plate motion (Duncan and McDougall, 1989; Lanyon et al., 1993). At anomaly 19 (~43 Ma), spreading in the Southern Ocean between Australia and Antarctica accelerated to about 5 times its previous velocity, and imposed a major N–S-directed motion on the Australian plate (Veevers et al., 1991).

This reorganization eventually drew the PTB-bearing rifted eastern margin of the Norfolk–New Caledonia Ridge into the trench around 38 Ma, terminating arc magmatism on the Loyalty–d’Entrecasteaux arc and emplacing the forearc-derived ophiolite on New Caledonia and initiating accompanying foredeep sedimentation (Fig. 5d).

Middle Eocene foredeep sediments and olistostromes record the collision of the Norfolk–New Caledonia Ridge continental ribbon with the forearc region of the intra-oceanic Loyalty–d’Entrecasteaux arc (Aitchison et al., 1995a). We propose that the first nearby allochthon (almost parautochthon) to be detached and emplaced southwestward in this collision is that represented by the PTB. Subsequent allochthons were derived from the forearc region of the colliding arc system, and these piggy-backed over the PTB parautochthon. The mantle section of the forearc may eventually have over-ridden its own lava-cumulate carapace. This ophiolite emplacement could have been synchronous or preceded a similar collision observed in the NW New Zealand peninsula (Brothers and Delaloye, 1982). Post-collisional extension led to exhumation of the high-grade metamorphic basement of the Norfolk–New Caledonia Ridge in northern New Caledonia (Aitchison et al., 1995a; Cluzel et al., 1995).

In summary, we believe that the Poya Terrane basalts form a proximal allochthonous sheet below the main harzburgitic nappe constituting the New Caledonian ophiolite. They are genetically unrelated to the ophiolite, and are interpreted to be 70–80-Ma-old rift tholeiites formed during opening of the East New Caledonia Basin, when an unknown continental fragment rifted eastward from the Norfolk–New Caledonia Ridge. Later closure of the East New Caledonia Basin, by an east-directed subduction, led to the collision of the western passive margin of the latter ridge with the forearc region of the intra-oceanic Loyalty–d’Entrecasteaux arc around 40 Ma. This collision forced the westward translation of the PTB as a slice picked up from the upper crustal section of the colliding Norfolk–New Caledonia Ridge. Subsequent collisional tectonism led to the detachment of the main New Caledonian harzburgitic nappe and its associated boninites from their forearc location, and their southwestward emplacement over the PTB nearby allochthon, the boninites being emplaced first below the main harzburgitic nappe.
8. Implications for other ophiolites

It is noteworthy that, in at least two other arc-continent collision settings with which we are familiar, there are parautochthonous sheets of basalts unrelated to immediately overlying forearc-derived, boninite-bearing harzburgitic ophiolites. Worthing and Crawford (1996) have documented the geochemistry of the Emo-Kokoda metabasalts from beneath the harzburgite-dominated, Eocene, boninite-bearing (Cape Vogel) Papuan ophiolite. Like the PTB, the Emo-Kokoda greenschists and amphibolites are dominated by low-K tholeiites with Zr/Nb values mainly between 8 and 16, but with two samples having typical N-MORB Zr/Nb values (>30). N-MORB-normalized element variation patterns show weak negative Nb anomalies, and the two samples analyzed for Nd isotopes have initial εNd values of +3 to +4. In the extensive Early to Middle Cambrian ophiolites of the Lachlan Foldbelt of eastern Australia, latest Proterozoic rift tholeiites (Crimson Creek Formation and correlatives) underlie the boninite-bearing, harzburgite-dominated ophiolite sheet at numerous locations in Tasmania where sections are best exposed (Crawford and Berry, 1992). These basalts are compositionally identical to the PTB, despite being ~600 million years older.

We conclude that basaltic volcanics underlying major ophiolites are not necessarily genetically related to the adjacent supra-subduction zone, probably forearc-derived ophiolites. Rather, they probably represent tholeiites erupted during extension and rifting of continental crust to form a passive margin. Subsequent arc-continent collision may have emplaced forearc-derived ophiolite allochthons onto the tholeiite-bearing passive margin, and one or more parautochthonous slices of this passive margin basement may have been transported along it, attached to the base of the ophiolite nappe.

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