Tectonic accretion and underplating of mafic terranes in the Late Eocene intraoceanic fore-arc of New Caledonia (Southwest Pacific): geodynamic implications

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Dominique Cluzel[a], Jonathan C. Aitchison[b] and Christian Picard[c]

[a] Institut des Sciences de la Terre d'Orléans, University of Orleans, CNRS UMR 6113, BP 6759, 45067 Orleans Cedex 2, France
[b] Department of Earth Sciences, University of Hong Kong, Pokfulam Road, Hong Kong, People's Republic of China
[c] University of New Caledonia, LGPMC, BP 4477 Noumea, New Caledonia

Abstract

This paper deals with the tectonic events that result in the accretion of mafic terranes in the fore-arc region and a close juxtaposition of ultramafic rocks, low grade and high-grade mafic terranes in many collisional orogens. The example is taken from New Caledonia where tectonic accretion, subduction, underplating and obduction of mafic terranes took place during the late Eocene in an intra-oceanic forearc setting. The late Eocene tectonic complex comprised three major terranes: an overlying ultramafic, mainly harzburgitic allochthon named the Ophiolitic Nappe, an intermediate mafic, mainly basaltic off-scraped melange, composed of kilometre-scale slices of oceanic upper crust, called the Poya Terrane, parts of which have been metamorphosed into an eclogite/blueschist facies complex, the Pouebo Terrane; and a lower, continental basement formed by the Norkolk Ridge terranes. Based upon exhaustive sampling of the mafic terranes and field surveys, our tectonic, micropaleontologic and geochemical data reveal that Poya and Pouebo terranes rocks originally formed within one single Campanian to late Paleocene oceanic basin, floored by tholeiitic basalt associated with some minor seamount-related intraplate alkali basalt. The tholeiitic basalt displays a continuous range of compositions spanning between “undepleted” and “depleted” end-members; the former being volumetrically predominant. The overall geochemical and isotopic features indicate an origin from a prominently heterogeneous mantle source during the opening of a marginal basin, the South Loyalty Basin, which almost completely disappeared during Eocene convergence. The opening of this basin originally located to the east of the Norfolk Ridge was synchronous with that of Tasman Sea basin as a consequence of oceanward migration of the west-dipping Pacific subduction zone. Establishing the origin of the ultramafic Ophiolitic Nappe is beyond the scope of this paper; however, it appears to be genetically unrelated to the mafic Poya and Pouebo terranes. Although it was located in the Late Eocene fore-arc, the Ophiolitic Nappe and the corresponding oceanic lithosphere originated before the Late Cretaceous, to the east of the South Loyalty Basin in a back-arc setting; or alternatively in a much older, trapped basin.

For reasons that remain unclear, a new east-dipping subduction started in the Eocene and consumed most of the South Loyalty Basin, forming the intra-oceanic Loyalty Arc. Due to a changing subduction regime (underplating of the Diahot Terrane?), the mafic slices that now form the Poya Terrane were tectonically accreted in the Loyalty fore-arc region and remained under low pressure-low temperature conditions (possibly at the subsurface) until the Norfolk Ridge reached the subduction zone diachronously. This resulted in the final obduction of the fore-arc area. The two-step obduction involved first the mafic complex forming the Poya

Terrane and thereafter the lithospheric mantle that now forms the Ophiolitic Nappe. In contrast, pieces of the accretionnary complex were dragged down into the subduction zone, underplated at depth ca. 70 km and metamorphosed into high-temperature eclogite to form the Pouebo Terrane metamorphics that display the same geochemical features as the Poya Terrane basalt. A mid-to-late Eocene syntectonic piggy-back sedimentary basin (the Nepoui flysch basin) mainly filled with mafic clastic material and shallow water carbonates that record the progressive uplift of the fore-arc region due to the accretion and underplating of mafic ocean-related and other material. In contrast, a slightly younger foreland basin located upon the Norfolk Ridge (the Priabonian Bourail Flysch basin) received a massive input of detrital material derived from the Norfolk Ridge itself and a time-increasing amount of mafic, Poya-derived material that recorded the first step of obduction. Thereafter, the Bourail Flysch was overthrust by the Poya Terrane and finally by the Ophiolitic Nappe.

At the same time, buoyancy-driven uplift and exhumation of the high-pressure metamorphic complex occurred as a consequence of the diachronous blocking of the subduction zone. Finally, for a short time a new subduction started along the west coast of New Caledonia and generated small amounts of active margin-related magmas. These events have resulted in the close juxtaposition of unmetamorphosed and highly metamorphosed mafic and ultramafic terranes that may represent a good pre-collision analogue of mafic/ultramafic belts found in many collisional mountain ranges.

and subduction-related (Kroenke, 1984). A recent reappraisal of the structural, paleontological and geochemical features of New Caledonian terranes (Cluzel; Aitchison and Aitchison) allows correlation of the Formation des basaltes with widespread discrete mafic schuppen (tectonic slices) pinched beneath the ultramafic terrane. This whole set has been renamed the Poya (allochthonous) Terrane (Cluzel and Cluzel) from the name of the type locality (Fig. 1). Geochemical features (Eissen et al., 1998; this paper) and paleomagnetic data (Ali and Aitchison, 2000) of the Poya Terrane basalt allow its interpretation as a marginal basin originally located to the east or northeast of New Caledonia. Furthermore, the Poya Terrane may be tentatively correlated with the Pouebo Terrane, a mafic eclogite/blueschist complex that crops out at the northern tip of the island (Cluzel et al., 1994). The predominantly mafic Poya and Pouebo terranes have previously been regarded as of uncertain origin. According to recently obtained data, they display some geochemical similarities, and are therefore suspected to be related. The question of whether or not Pouebo and Poya terranes were generated in the same basin has been discussed (Cluzel and Meffre) although no compelling evidence was presented. In order to check this hypothesis, we have considered three kinds of evidence: (1) the age of these terranes based on fossil content or radiometric data; (2) their magmatic affinities; and (3) their tectono-metamorphic features. Further, the relationship between the Poya and Pouebo terranes, and the ultramafic Nappe Ophiolitique will be discussed in the light of syntectonic sedimentation.

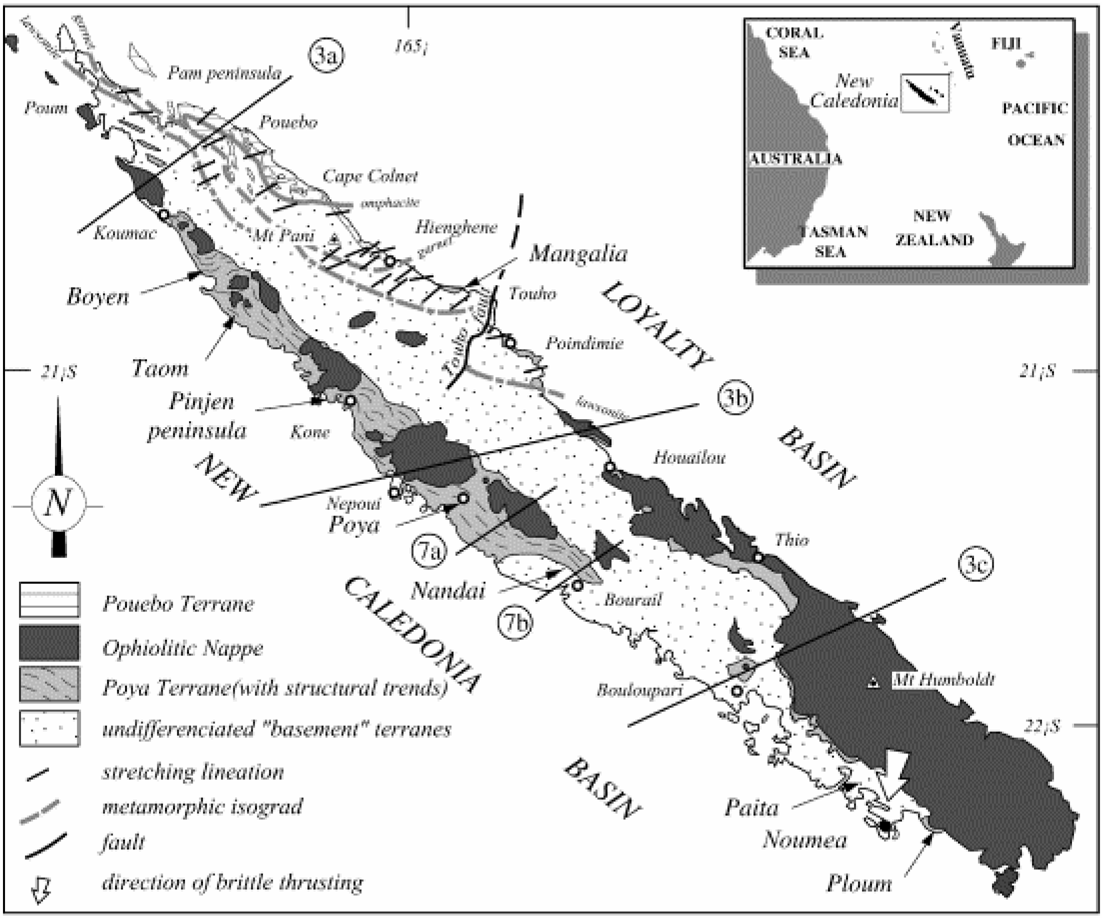

Fig. 1. Structural map of the “Grande Terre” of New Caledonia with emphasis given to the late Eocene tectono-metamorphic features. (location of Fig. 6 and Fig. 7 cross-sections). The white arrow indicates the late emplacement direction of the Nappe Ophiolitique determined from fracture analysis in the Noumea peninsula.

Finally, all available data will be synthesized in a geological model which highlights: (1) the role of oblique subduction and arc migration due to slab roll-back; (2) the relationship between obduction of mantle ophiolites and LT-HP metamorphic belts; (3) the role of microplates in the accretion of large mafic terranes in fore-arcs, and (4) its consequences on continentward obduction of large mafic-ultramafic belts in collisional orogens.

The data and interpretation presented here are based upon new field evidence and exhaustive sampling by the authors since 1992.

2. Outline of the geology of New Caledonia

New Caledonia is located in the Southwest Pacific, within a complex set of marginal basins and “continental” or volcanic-arc ridges (Fig. 2). It is composed of several islands that are part of the parallel Norfolk and Loyalty ridges. The main island (or “Grande Terre”) belongs to the Norfolk Ridge, which is connected southward to the large continental plateau that also bears New Zealand. The Belep Islands to the north, and Ile des Pins to the south of the main island also belong to the Norfolk Ridge. In contrast, the Loyalty Islands represent the emerged part of a sinuous submarine ridge (the Loyalty Ridge) that runs more or less continuously parallel to the Norfolk Ridge over more than 1500 km from the d'Entrecasteaux Ridge (west of Espiritu Santo, Vanuatu) in the north, to the Cook Fracture Zone in the south (Fig. 2). The geology of the Loyalty Ridge is still poorly known due to the lack of basement outcrops and a thick carbonate cover. Upper Miocene basalts/dolerites that locally crop out on Mare Island display typical intraplate alkaline (OIB) affinities and are likely to be part of a north–south-trending hot spot trail and younger than the ridge itself (Meffre, 1995). Geophysical and swath bathymetry data (Bitoun and Lafoy) indicate that this ridge appears to be composed of a series of regularly spaced seamounts the sizes and shapes of which are very similar to those present in most island arcs. The northernmost seamount of the ridge (the Bougainville Seamount) has been drilled to the west of Espritu Santo Island, where it is being subducted into the New Hebrides trench, and Eocene andesite was found underneath 700-m-thick Oligocene to Holocene limestone (Dubois; Collot and Greene). In the North-Loyalty Basin, at the DSDP 286 site, Middle to Late Eocene andesitic volcanoclastic turbidites have been drilled (Andrews et al., 1975). Furthermore, the Three Kings Ridge which is the extension of the Loyalty Ridge south of the Cook Fracture Zone is an Oligocene–Miocene island arc that is most likely based upon an older, possibly Eocene island arc (Cluzel et al., 1999). Because of these features, we consider the Loyalty Ridge to be an Eocene island-arc, and the North Loyalty Basin the associated back-arc basin. This interpretation has a major bearing upon our understanding of the Eocene subduction/obduction events.

Between the Loyalty and Norfolk ridges, the Loyalty Basin is a narrow, 1500–3000-m-deep, oceanic basin partly filled with 1000–3000-m-thick post-Eocene (?) sediments. The oceanic lithosphere of the Loyalty Basin is more or less in continuity with the Eocene Nappe Ophiolitique resulting from obduction of the basin (Collot et al., 1987). The fan-like shape of seismic reflectors within the sedimentary filling of the basin (Bitoun and Récy, 1982) and occurrence of disorganised sedimentary bodies due to large-scale slumping along its southwestern boundary (Vially and Bénard, Institut Français du Pétrole, pers. commun., 1999) are evidence of its syntectonic character (i.e. syn-obduction). To the southwest of the Grande Terre, the New Caledonia Basin, which is located between the Lord Howe and Norfolk ridges, is of uncertain origin, floored by either oceanic, or thinned continental crust, or both (Mignot; Symonds; Symonds and Auzende).

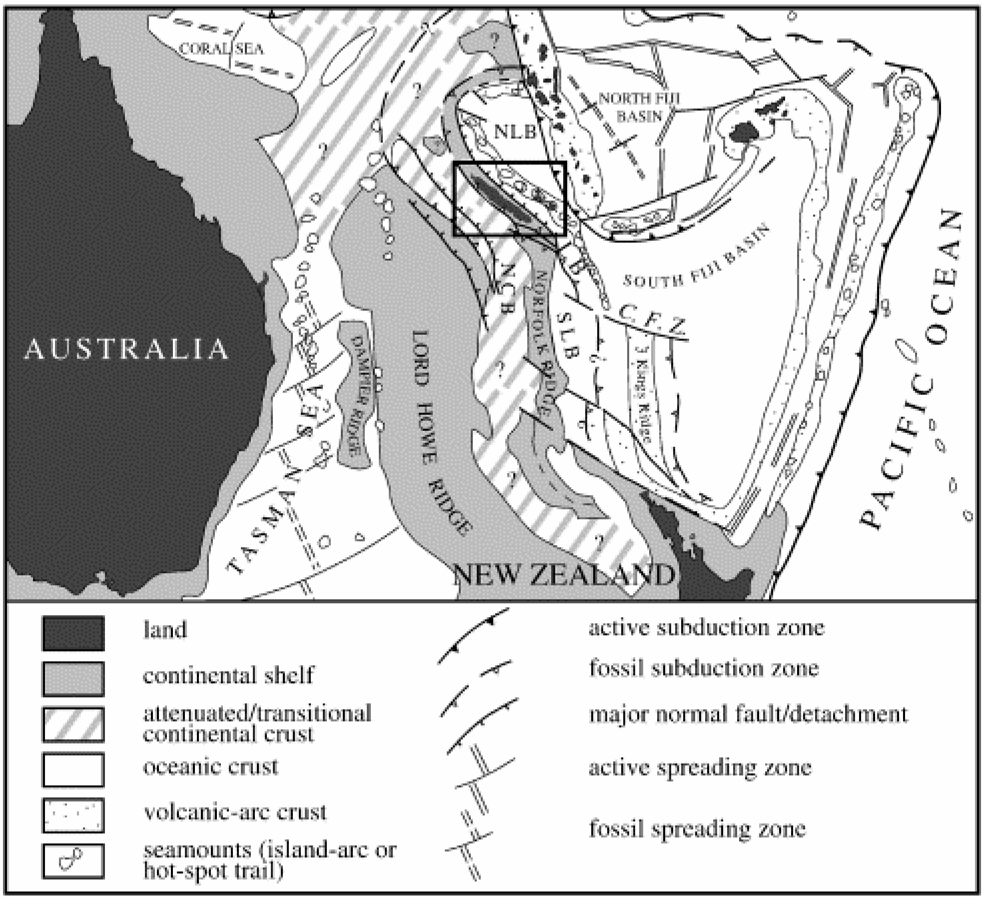

Fig. 2. Structural sketch map of the Southwest Pacific (LB: Loyalty Basin; NLB: North Loyalty Basin; NCB: New Caledonia Basin; SLB: South Loyalty Basin; CFZ: Cook Fracture Zone).

Similar to New Zealand, the Grande Terre of New Caledonia is a complex mosaic of volcanic, sedimentary and metamorphic terranes. New Caledonian terranes were assembled during two major tectonic episodes; a Late Jurassic to Early Cretaceous tectonic collage the assembly of which may be correlated with the Rangitata Orogeny of New Zealand (Bradshaw and Paris); and a late Eocene obduction/collision (Avias; Paris; Collot; Aitchison and Aitchison). Both events included periods of high-pressure metamorphism and are therefore thought to have occurred in connection with plate convergence in subduction zones.

Pre-Cretaceous terranes comprise: (1) the disrupted Late Carboniferous (Aitchison et al., 1998) fore-arc Koh ophiolite (Meffre et al., 1996); (2) the mid-Triassic to late Jurassic Central Chain volcano-sedimentary arc terrane (Gu and Meffre); (3) the mid-Permian to Late Jurassic Teremba volcano-sedimentary arc terrane (Campbell and Campbell); and (4) the undated Boghen Terrane which is composed of metamorphosed intra-oceanic pillow lavas and mixed terrigenous and volcanosedimentary deep fan sediments of uncertain origin (Cluzel, 1996). Except the latter, all pre-Cretaceous terranes developed in arc-related settings offshore the eastern Gondwana active margin during periods of marginal basin development (Meffre, 1995). They have been tectonically amalgamated along the edge of the Australian continent and eroded before the deposition of unconformably overlapping Upper Cretaceous terrigenous rocks; as attested to by the occurrence of Precambrian (ca. 1 Ga) zircons reworked in marine deltaic sandstone (Aronson and Aitchison). The overall weakness of tectonic deformation and occurrence of Late Jurassic high-pressure metamorphism in the Boghen Terrane only (Blake et al., 1977) favour the hypothesis that these terranes represent a tectonic collage that developed along the Late Jurassic to Early Cretaceous east-Australian convergent margin.

Post-Early Cretaceous terranes include five units:

(1) The New Caledonia block (or super-terrane) which was formed through the amalgamation of the four previously mentioned terranes, and subsequently rifted away from Australia during the Late Mesozoic Gondwana break-up. These pre-Cretaceous terranes are unconformably covered by a late Cenomanian to late Eocene sedimentary pile.

(2) The Senonian to Eocene (?) Diahot Terrane (Cluzel et al., 1994) is composed of metamorphosed arc-related volcanic and volcanosedimentary rocks, terrigenous argillite, sandstone and conglomerate, and pelagic siliceous argillite.

(3) The Pouebo Terrane (Cluzel et al., 1994) which is mainly composed of eclogite-facies tholeiitic metabasic rocks, metasediments and talc-schist.

(4) The Poya Terrane (Cluzel et al., 1994) is composed of tholeiitic basalt with subordinate bathyal red chert and argillite.

(5) The ultramafic “Nappe Ophiolitique” (Avias, 1967) represents the oceanic mantle lithosphere of the Loyalty Basin which was obducted in the Late Eocene (Paris; Prinzhofer; Prinzhofer and Collot). All these terranes are fault-bounded and thus bear no stratigraphic relationship to each other.

The post-Jurassic sedimentary pile in New Caledonia contains information crucial to constraining the pre-obduction period in New Caledonia; it may be divided in two parts. The lower megasequence (Fig. 3) is formed of a late Cretaceous volcanosedimentary and terrigenous pre-and syn-rift sedimentary pile overlain by Maastrichtian/Paleocene to mid-Eocene hemipelagic or pelagic post-rift sediments. The upper megasequence (Late Eocene, see below, Fig. 4) has recorded the pre-obduction events related to the arrival of the Norfolk Ridge in a newly formed east-dipping subduction zone that also generated the Loyalty Arc (Fig. 5).

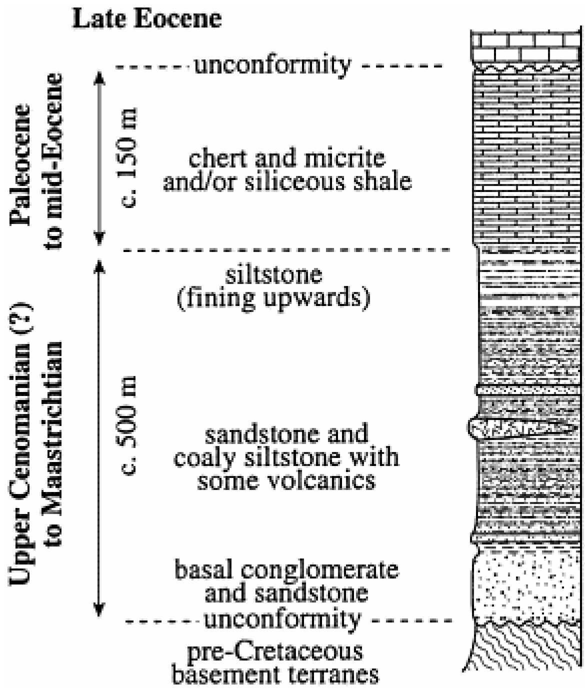

Fig. 3. Sketch columnar section of the Late Cretaceous to mid-Eocene sedimentary megasequence of New Caledonia (Noumea region) showing the late Cretaceous fining upwards clastic sequence and the Paleocene to mid Eocene siliceous and carbonated post-rift sequence.

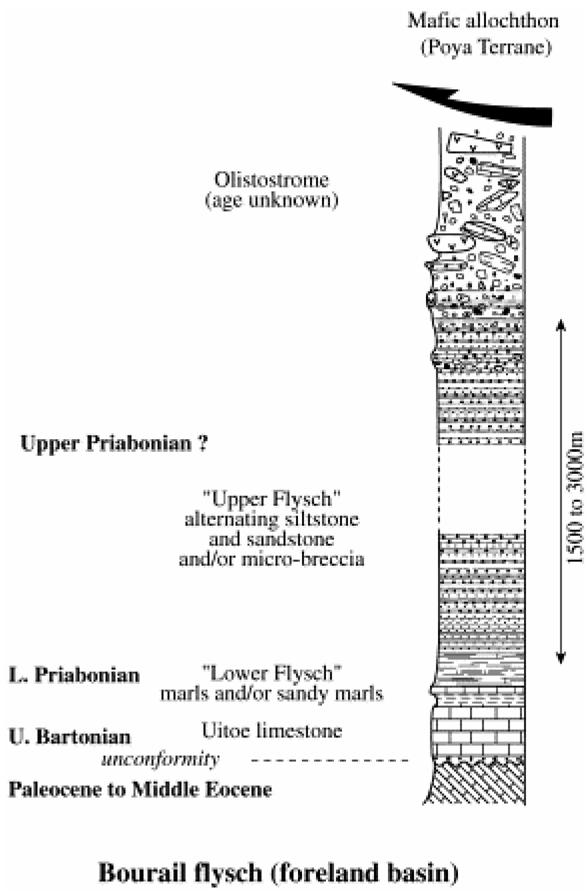

Fig. 4. The Late Eocene pre-obduction megasequence of New Caledonia (Noumea-Bourail area) showing the basal unconformity due to the fore-arc bulge emersion and the upwards coarsening sequence of the Bourail flysch topped by an olistostrome.

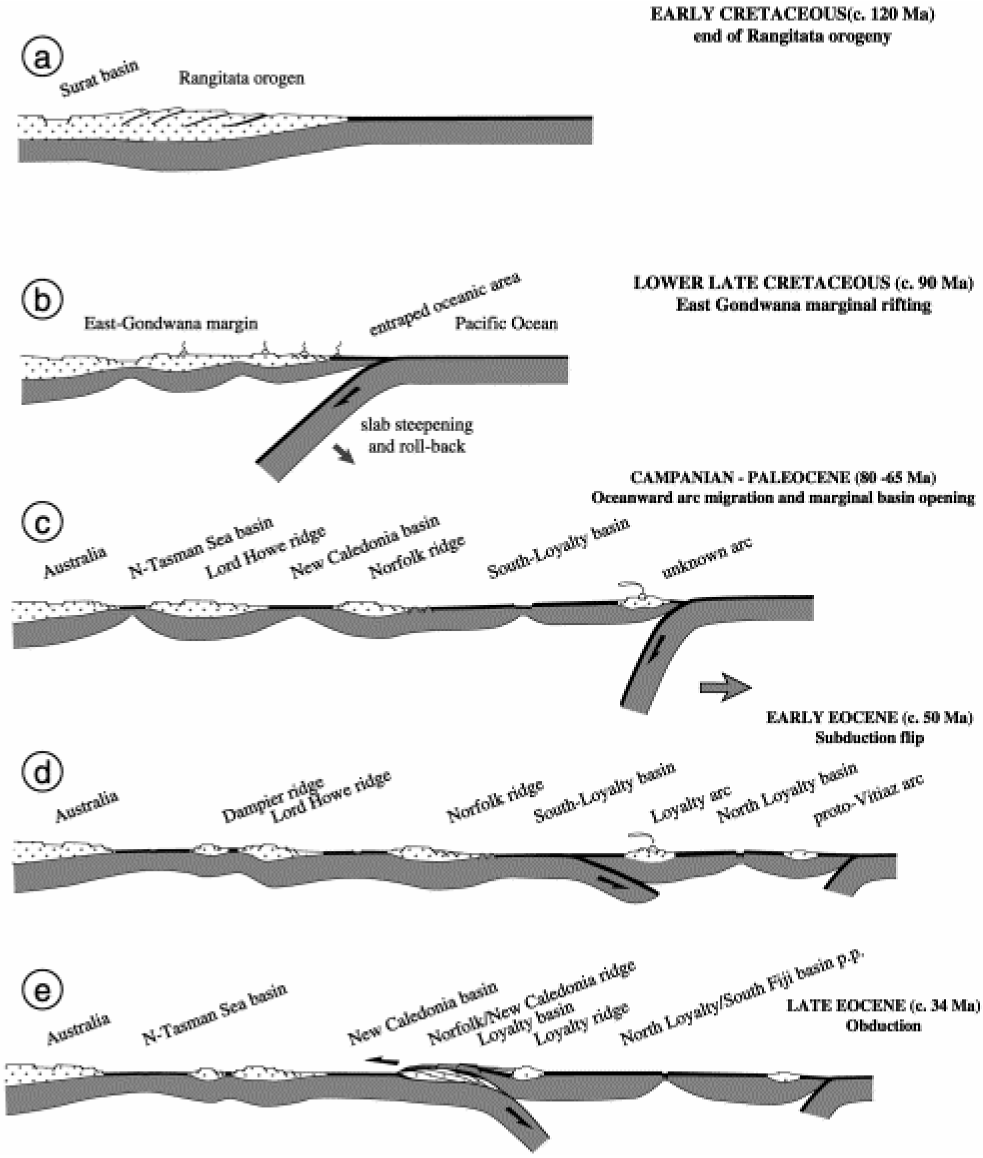

Fig. 5. Late Cretaceous to Late Eocene geodynamic evolution of the Southwest Pacific depicting the lithospheric boudinage and marginal basin opening (late Cretaceous to mid-Eocene), and the Late Eocene convergence following the subduction reversal and North-Loyalty back-arc basin opening.

Following the late Jurassic–early Cretaceous collage, marine late Cenomanian to Campanian (93–72 Ma) terrigenous deposits (commonly designed as "Senonian"), partly derived from the Australian continent (Aronson and Aitchison), accumulated unconformably upon pre-Cretaceous rocks. The coal-bearing clastic rocks locally contain calc-alkaline and alkaline volcanic rocks (Black; Picard; Gros and Debeuf) that recorded the post-collage active margin magmatic activity and arc-to-rift transition within the eastern Gondwana margin. The volcanic arc activity was likely due to a west-dipping subduction zone that also triggered and/or controlled the opening of Tasman Sea and New Caledonia marginal basins due to the oceanward migration of the active arc (Fig. 5a and b). During the Coniacian to Campanian (88–72 Ma), terrigenous deposits are progressively fining upwards and change in the Maastrichtian–Palaeocene (70–60 Ma) into a pelagic chert-and-micrite association that deposited onto the mid-Eocene (ca. 42 Ma; Gonord and Maurizot) (Fig. 3). The end of terrigenous imput and the thermal subsidence are due to the marginal rifting of the Tasman

Sea and New Caledonia basins that opened between Australia and the Norfolk and Lord Howe ridges, respectively (Cluzel; Aitchison and Aitchison) (Fig. 2 and Fig. 5). Magnetic anomaly indicate an early Campanian (83–79 Ma) opening age for the southern part of the New Caledonia basin (Sutherland, 1999); but a disconformity in the basin infill has been interpreted to post-date the rifting at ca. 70 Ma (Maastrichtian) in the northern part of the basin (Mignot, 1984) infering a diachronous nothward opening of the basin. Similarly, the Tasman basin opened diachronously from south to north during the Campanian–Paleocene (Weissel and Gaina); therefore, landbridges between Australia and New Caledonia were likely to still exist during the deposition of the "Senonian" (Fig. 5b). A complete opening of the Tasman Sea was effective in the Upper Paleocene (ca. 53 Ma) that is well before the onset of obduction-related convergence; therefore, the latter is likely due to an independent process, e.g. the mid-Eocene acceleration of Antartica–Australia northward drift and changing vergence of the Pacific plate at ca. 43 Ma; and/or the westward migration of the Loyalty arc due to slab roll-back and north-Loyalty/south-Fiji Basin opening (Fig. 5e).

The Upper Eocene pre-obduction sequence developed after a short period of emersion and fault/block tectonism marked by angular unconformity and erosion (Coudray and Cluzel). Upper Bartonian–lower Priabonian (ca. 40 Ma) carbonate rocks lie unconformably upon eroded pre-Cretaceous to mid-Eocene terranes. The sequence started with pebbly shallow water limestone; thereafter, the basin depth increased abruptly and marls or calcareous sandy marls accumulated upon the basal limestone. The marls progressively change into alternating sandy marls or calcarenite and marls. In turn, the "lower flysch" is overlain by the "upper flysch" (Tissot and Noesmoen, 1958) with local disconformity (Gonord, 1977). The "upper flysch" is an upward coarsening sequence toped by an olistostrome (Fig. 4) that reworks all the components of the sedimentary cover including clasts of the Upper Eocene bioclastic limestone and the flysch itself (Gonord, 1970). The olistostrome is in turn tectonically overlain by the Poya Terrane and finally the Ophiolitic Nappe.

3. The ultramafic allochthon

The "Nappe Ophiolitique" is the dominant geologic feature of New Caledonia. It is mainly composed of harzburgite and dunite forming a slice of obducted sub-oceanic mantle lithosphere. The mafic and ultramafic cumulates (the paleo-Moho) that crop out in the Massif du Sud (Dupuy et al., 1981) formed the bottom of magma chambers that still represent 20% of the surface of the ophiolite. It most probably had a complete crustal sequence that was deeply eroded or tectonically detached before, during and/or after obduction. From the N–S orientation of the high-temperature stretching lineation in the harzburgite, and wrench shear zones interpreted as transform faults (Prinzhofer and Prinzhofer), a genesis from an east–west-trending ridge is expected. This has been taken as an evidence for the north-over-south overthrusting direction of the ophiolite that was thought to have resulted from the synchronous opening of the Loyalty Basin (Collot et al., 1987). However, no direct evidence of the age of the ultramafic rocks is available. The few existing data for the ophiolite are not consistent with this interpretation. Crosscutting mafic dykes display two age groups at 80–100 and ca. 50 Ma (K–Ar, Prinzhofer, 1981). As the older mafic dykes crosscut the harzburgite and transition zone, they must postdate the pre-Late Cretaceous formation of the ophiolite itself. Therefore, the orientation of the ductile stretching lineation bears no relationship to the Late Eocene obduction. The younger mafic dykes, which mainly display island-arc tholeiite (IAT) affinity (in progress), likely record the pre-obduction subduction of the South Loyalty basin below the Loyalty arc.

Eissen et al. (1998) pointed out that there is no way for the “Nappe Ophiolitique” to be genetically related to the Poya Terrane basalt. Indeed, residual harzburgites as well as the mafic cumulates are extremely depleted and cannot have been in equilibrum with a melt of MORB composition. Therefore, we consider that the Nappe Ophiolitique only represents the older basement upon which the Loyalty Arc was established.

4. New Caledonian mafic terranes

The post-early Cretaceous mafic terranes of New Caledonia have major importance with respect to our knowledge of the subduction–accretion–obduction process. They consist of the relatively low-grade Poya Terrane and the high-grade Pouebo Terrane.

The Poya Terrane (Cluzel and Cluzel), previously named “Formation des basaltes” (Routhier, 1953), is a regionally extensive 200-km-long unit exposed along the west coast of New Caledonia from Bourail to Koumac, and at a number of smaller outcrops on the east coast and in the Noumea area (Fig. 1). A 200-km-long magnetic anomaly extending north-westwards off the axis of the island may be interpreted as the offshore trace of the terrane (Collot et al., 1987). This terrane, which is always located underneath the ultramafic allochthon (the “Nappe Ophiolitique”), is a tectonically complex set of pillowed and massive basalts, dolerites and fine grained gabbros interbedded with, and covered by, bathyal black, green or red argillite and chert. The magmatic rocks are locally associated with some mineralization (Mn, Cu, Hg and W) and Au occurrences associated with Hg and Cu have been discovered locally. No economic-size resources have been proven, but polymetallic mineralization of this type is a common feature of marginal basins. Due to the lack of reliable age data, this mafic terrane was controversially interpreted either to be allochthonous (Espirat; Guillon; Avias; Gonord; Black; Cluzel; Cluzel and Eissen); the detached cover of the tectonically overlying ultramafic allochthon (Espirat, 1963); or autochthonous (Routhier; Paris; Kroenke; Maurizot and Maurizot), and thought to have been erupted in a continental active margin setting as a result of Neogene east-dipping subduction along the west coast of New Caledonia (Paris, 1981) the volcanic activity being controlled by the so-called “west-Caledonian fault” (Maurizot and Maurizot); or alternatively, in connection with west-dipping subduction along the east coast (Kroenke, 1984). These interpretations do not fit the magmatic affinities of the Poya Terrane basalts which are predominantly MORB-like tholeiites erupted in a marginal basin (Cluzel and Eissen; this paper), nor the age of the terrane as determined from radiolarian faunas (Aitchison and Aitchison; this paper).

The Poya Terrane overlies all pre-Neogene terranes of New Caledonia and is overthrust by the ultramafic allochthon (Fig. 1 and Fig. 6). Without exception, the boundaries of the Poya Terrane are faulted. This has, however, been interpreted by previous authors as a consequence of limited reverse faulting and the allochthonous character of the unit has not been hitherto accepted. If we consider that the basalts which are located on both coasts and in the Central Range are of correlative ages and have similar geochemical features (this paper), then it is likely that they belong to a single unit that not only covered two thirds of the island, but may also extend below the ultramafic nappe. Where observed the basal thrust along the northeastern boundary of the terrane, dips gently to the southwest and is potentially mistaken for an extensional normal fault; in any case, this rules out the possibility of a late reverse faulting (see above). In most “dolerite” or massive basalt outcrops, the original attitude of the rocks is impossible to decipher and except for a great number of brittle and semi-brittle shears and tension cracks filled with calcite, quartz and other low grade minerals, it is quite difficult to evaluate the strain intensity. In contrast, where abyssal argillite and chert are interbedded with mafic rocks, or in pillow basalt outcrops, it is clear that the terrane has been severely folded, faulted and disrupted. Indeed, except in a few localities along the west coast (Pinjen, Foué, Voh), most sedimentary inliers are fault-bounded, dip steeply (60–90°, generally to the north–east or to the north) and in several places overturned pillows may be observed (Fig. 7a and c).

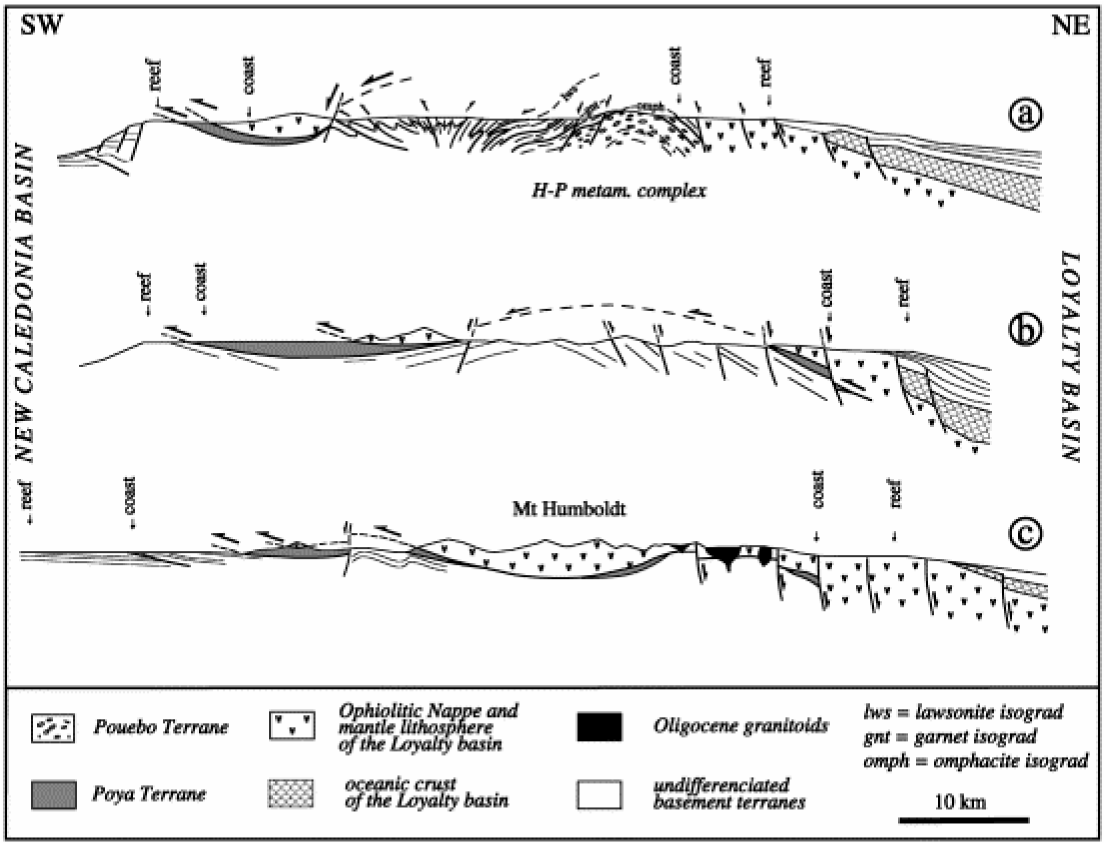

Fig. 6. General cross sections of the “Grande Terre” showing the relationship between the “Nappe Ophiolitique”, the Poya Terrane and the late Eocene high-pressure metamorphic complex.

The main unit of the Poya Terrane may therefore be interpreted as a duplex (i.e. imbricate thrust complex) consisting of a pile of dominantly basaltic slices. The absence of “continent-related” lithologies within the complex is a major feature from which it may be inferred that duplexing occurred in an intra-oceanic environment. Along the west coast, the overall shape of the terrane is that of a broad synform. In the Bourail area, however, the southeastern pericline displays steeper boundaries indicating a late stage compression of the terrane (the Tene synform, Fig. 7b). The average strike of steep sedimentary inliers, major faults and serpentinite lenses is N50°W. This is 10° oblique to the strike of the island, and more or less perpendicular to the strike of the stretching lineation of the high-pressure metamorphic complex (Fig. 1, see below). This average strike is disturbed by kilometre-scale folds with steeply dipping axes (Fig. 1) that also effect the whole island, including the Late Eocene metamorphic complex, with the exception of the “Nappe Ophiolitique”. The asymmetry of these broad folds is consistent with a dextral transpression that obviously occurred during or immediately after eclogite exhumation. The absence of such a deformation in the “Nappe Ophiolitique” may infer either that the latter was emplaced after this tectonic episode. An alternative explanation more consistent with the regional geologic features is that the tectonic sole of the ultramafic allochthon played the role of a major detachment that allowed decoupling between the deforming complex and the overriding plate (e.g. the oceanic lithosphere of the Loyalty Basin).

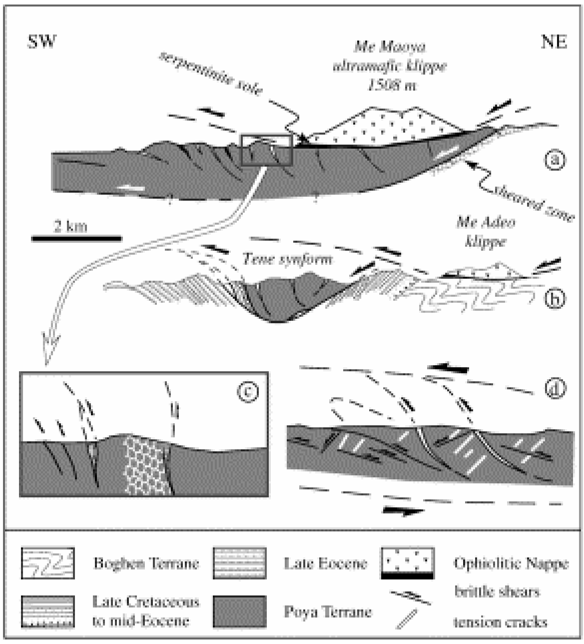

Fig. 7. Tectonic structure of the Poya Terrane; (a) general cross-section of the Poya Terrane near the type-locality (loc., see Fig. 1); (b) cross section of the southern end of the main body of the Poya Terrane to show the basal parautochthonous schuppen and late shrinking of the terrane (loc., see Fig. 1); (c) a detail of the Me Maoya section (a) to show the upright attitude of the mafic slices marked by pillow structure and sedimentary inliers; (d) cartoon showing the geometry of minor structures indicating south to southwest-directed shearing of the terrane (no scale).

Considering the brittle or semi-brittle character of faults and the kink style of the associated folds, duplexing may have occurred at a moderate depth. At the base of the terrane (i.e. around the New Zealand WWII Cemetery, along the main road, 4–7 km southeast of Bourail, Fig. 7b), some 100 m scale tectonic slices of the autochthonous terranes may be found, they comprise “Senonian” sandstone and siltstone, Bartonian calcarenite, and Boghen Terrane basalt, that have been off scraped and pushed in front of the Poya Terrane. The lack of slivers of these terranes within the duplex itself infers that the latter has been first duplexed and finally thrust “en bloc” over the autochthonous (i.e. the Norfolk Ridge). Brittle features such as tension cracks, low-angle shears and south- or southwest-facing normal faults are consistent with southwest- or south-directed shearing of the terrane (Fig. 7d). Shearing occurred either during duplexing, or more likely, overthrusting by the Nappe Ophiolitique.

In contrast to the west coast, along the east coast and in the central range, thinner and tectonically disrupted mafic slices characterise the Poya Terrane which is severely pinched between the autochthonous and the ultramafic allochthon. It is not certain that all the mafic bodies located beneath the peridotites actually belong to the Poya Terrane as shown on the 1:200,000 geologic map of Paris (herein named the “west coast basalts”; Paris, 1981). Indeed, some of these basalts (e.g. along the lower reaches of Tiwaka river near Poindimié) are boudinaged and metamorphosed arc-tholeiites that may be correlated with the Central Chain Terrane (Meffre, 1995); others are E-MORB basalts similar to those of the Poya Terrane but display a higher metamorphic grade (lower amphibolite facies), are closely associated with

Boghen Terrane schist (e.g. between Thio and Canala at the Petchecara Pass and along the upper reaches of the Dothio River) and could be correlated with the latter. Nevertheless, most of mafic tectonic bodies found between Canala and Hienghene may be correlated with the Poya Terrane on the basis of their lithologic, tectonic and geochemical features.

The Poya Terrane in general (except alkaline basalt found in Foué and Pinjen peninsulas, Fig. 1) has been affected to various degrees by a low pressure, low-temperature metamorphic event (M1), characterized by greenschist facies and lower grade mineral assemblages (chlorite–smectite±prehnite±pumpellyite±actinolite–green hornblende, calcite). M1 is absent in the underlying rocks, but present in the mafic clasts of the upper Eocene breccia; therefore, it may be distinguished from the late Eocene high-pressure tectono-metamorphic event (M2). The M1 metamorphic mineral development is characterised by random orientation and thus is likely to result from "ocean floor" metamorphism related to zones of hot rock and sea water interaction. From the mineral assemblages present, maximum temperatures may be estimated at ca. 300 °C. This indicates that these rocks have recrystallised at depth not exceeding 2–3 km and thus represent the upper levels of an "oceanic" crust. There is no systematic evolution in the average metamorphic grade within the main unit of the Poya Terrane along the west coast. Instead, very low grade rocks may coexist with greenschist facies rocks as a result of contrasting rock/water ratios and subsequent tectonism. After M1 development, the terrane was severely disrupted by faulting, such that no isograd can be traced in the field. This is, nevertheless, evidence which corroborates interpretation of the Poya Terrane as an allochthonous accretion complex (see below) and clearly accounts for the inconsistency of radiometric data.

The M1 event was followed by syntectonic high-pressure metamorphism of probable latest Eocene age (M2), which is only present along the northern part of the east coast. The metamorphic grade increases northwestward together with ductile deformation, and a penetrative schistosity is first noticed within red argillite beds in the Touho area and appears further north (Hienghene region) in mafic rocks as well. Blueschist facies minerals are mainly sodic amphiboles (crossite-glaucophane) developing against clinopyroxene. From the polarity of the M2 event, we infer the southward or south-southwestward vergence of the Poya Nappe which is confirmed by kinematics of low angle shears and normal faults. From the occurrence of high pressure minerals, we also infer the infra-subduction zone character of the M2 metamorphism experienced by parts of the Poya Terrane. Thus, the Poya Terrane cannot realistically be considered to be the detached cover of the supra-subduction "Nappe Ophiolitique" as postulated by Espirat (1963).

The Pouebo Terrane (Cluzel and Cluzel) mainly crops out at the northern end of the island, on the Pam Peninsula and Balabio Island, and discontinuously along the east coast from Balade to Poindimié (Fig. 1). It is mainly composed of metamorphosed mafic rocks which were described as metadolerites by early authors (Lacroix, 1941). These high-pressure metamorphic rocks have been extensively studied from a mineralogical point of view in order to decipher their metamorphic evolution (i.e., Bell; Yokoyama; Black; Black; Clarke; Carson and Carson), but very little has been known of their tectonic development (Lillie; Briggs and Clarke) and nothing about their geochemical features. On the occasion of remapping the Pam-Ouegoa sheet by BRGM, these rocks were interpreted to be a metamorphosed ophiolitic melange, originally composed of mafic rocks, serpentinite and bathyal sediments (red chert) (Maurizot et al., 1989). Indeed, some outcrops (Amos Pass, Balade, Koe Beach, Anse Ponandou) display unsorted, melange-like cyclopean conglomerates composed of 0.1–10 m scale mafic clasts or blocks. However, in other places, hundreds of metre-to-kilometre scale mafic slices/blocks are bounded by anastomosing retrograde hydrous shear zones related to the tectonic uplift and unroofing of the eclogite complex, so that it is difficult to distinguish primary (block) and tectonic boundaries. The Pouebo Terrane was interpreted as the pre-Senonian basement of the Diahot Terrane, and therefore correlated with the Triassic–Jurassic Central Chain Terrane by Maurizot et al. (1989). However, this interpretation does not account for the metamorphic gap that exists between the Diahot and Pouebo terranes; nor does it account for magmatic affinities of Pouebo Terrane mafic rocks (mainly E-MORB tholeiites; this paper) that sharply contrast with those of the Central Chain Terrane (arc-tholeiites; Meffre, 1995). In addition, together with a number of uncorrelatable mafic or felsic metasediments now metamorphosed into gneiss or blueschist/greenschist rocks (Ghent and Ghent), the Pouebo Terrane contains some blocks/boulders that may be correlated with the Diahot Terrane. Some felsic gneisses (Pam Peninsula) may represent metamorphosed calc-alkaline volcaniclastic or volcano-sedimentary rocks and blocks of quartz-bearing metaconglomerate found at Anse Ponandou (Touho) are similar to those of the Diahot Terrane (e.g. to the northwest of Koumac, at Tiabet). Like the Poya and Diahot terranes, the Pouebo Terrane displays some base metal and gold mineralization which has been tectonically reworked and is now contained in late quartz veins. Therefore, the Pouebo Terrane clearly appears to be a metamorphosed melange that reworks lithologies similar to those of the Diahot and Poya terranes (see also correlation of mafic rocks below). As with the other "basement" terranes of New Caledonia that were metamorphosed in the late Eocene, the Pouebo and Diahot terranes contain numerous slivers of serpentinite (0.1–10 km long, 0.01–2 km thick) metamorphosed to various degrees. These serpentinite bodies are interleaved with high-pressure metamorphic rocks and likely result from the tectonic erosion of the hanging wall in the subduction zone (i.e. they represent fragments of the mantle lithosphere of the Loyalty Basin).

A recent tectonic and metamorphic reappraisal of this terrane has shown that it forms the bulk of the uplifted eclogitic core of a metamorphic complex bounded by detachment faults and surrounded by the blueschist facies Diahot Terrane (Cluzel and Clarke). The Pouebo Terrane roughly forms the core of a large elongated foliation anticlinorium along the strike of which the metamorphic grade increases progressively but discontinuously from southeast to northwest, and steeply from southwest to northeast across the strike of foliation. The metamorphic isograds run roughly parallel to the strike of the island and are apparently parallel to lithologic boundaries, but the metamorphic zones are in fact tectonically juxtaposed along exhumation-related detachments. For example, eclogites of the Pouebo Terrane are in sharp contact with blueschist facies rocks of the Diahot Terrane along the Ouegoa–Amos transect (Fig. 6a), and thus appear there to have been severely uplifted with respect to the Diahot Terrane which is wrapped around it, as a result of late-exhumation shrinking.

The foliation anticlinorium bears a penetrative stretching lineation striking N60°E on average (Fig. 1). This regional lineation is composed of elongated pre-tectonic objects (i.e. pebbles, siliceous nodules), syntectonic minerals (albite, glaucophane), sheath fold axes, and quartz rods. Strain intensity progressively decreases upwards in the Diahot Terrane and abruptly downwards in the Pouebo Terrane; therefore, this stretching lineation may be interpreted as a result of differential motion between the two terranes during blueschist/greenschist facies metamorphism prior to final exhumation. Shear strain features (e.g. asymmetric pressure shadows, rotated albite porphyroblasts, S/C structures) systematically indicate a northeast over southwest relative motion along the strike of the lineation on both sides of the anticlinorium (Cluzel et al., 1995). Post-kinematic mineral recrystallisation (e.g. sheaf-like stilpnomelane aggregates) that crosscut the foliation are symptomatic of the post-kinematic thermal equilibration of the complex. Following the doming of the foliation limited right-lateral motion occurred along the regional southwest dipping foliation of the southwestern limb, overprinting and in places, erasing the downdip stretching lineation. The latter tectonic event may be correlated with the late open folds with steeply dipping axes that effect the regional foliation and isograds as well. The asymmetry of these folds is consistent with a regional-scale right-lateral motion. Such folds are also well developed in the Poya Terrane (see above) and to a lesser extent, in the rest of the island.

Along the east coast, swath bathymetry and geophysical data indicate the presence of a 250-km long linear tectonic boundary which may have facilitated the huge vertical differential motion between the metamorphic complex and the Loyalty Basin (Bitoun and Récy, 1982). In contrast with a common view (as in most metamorphic maps of New Caledonia), the metamorphic complex may be extended to the Southeast of Touho fault with an apparent dextral offset of ca. 6 km (Maurizot and Maurizot) (Fig. 1), onto the Houaïlou region where the gently dipping regional foliation passes below the Nappe Ophiolitique. Therefore, in contrast to some recent interpretations (Baldwin et al., 1999), there is apparently a direct relationship between subduction beneath the Loyalty Basin, now represented by the "Nappe Ophiolitique", and development of high-pressure metamorphism. The high-pressure metamorphic complex of northern New Caledonia has not been tectonically juxtaposed to the rest of New Caledonian lower grade terranes during a major post-metamorphic dextral displacement along a lineament striking NW–SE. There is no evidence to suggest that the so-called west-Caledonian fault exists. As the overriding plate (the Nappe Ophiolitique) has not been affected by this tectonic event, right-lateral transpression is likely to have occurred in the lower plate only, during the late stages of obduction.

On the Pam Peninsula, peak metamorphic conditions in barroisitic eclogites may be estimated at 19±1 kbar at 650±30 °C during prograde rehydratation of the eclogite, while pressures may be estimated at 9.0±1.2 kbar at 400±50 °C in the chloritoid-bearing schist of the Diahot unit near Ouégoa (Clarke and Carson). This pressure gap of minimum 10 kbar is a common feature of high-pressure complexes and had already been noticed but under-evaluated at ca. 6–8 kbar (Brothers and Maurizot). It gives an estimate of the huge differential motion that occurred between the two terranes after eclogite facies development. The age of peak metamorphic conditions is still unknown. ^{39}Ar–^{40}Ar data on phengites, however, provide similar cooling ages in both terranes at ca. 40–34 Ma (Baldwin et al., 1999). The more scattered and older ages previously obtained on the lower grade Diahot lawsonite schist (ca. 44–51 Ma) (Ghent et al., 1994) are likely due to incomplete resetting of an inherited sedimentary phase. Final exhumation of the terrane to shallow crustal levels is likely to have occurred shortly after, as shown by K-feldspar ^{39}Ar–^{40}Ar (37–34 Ma) and apatite fission track data (34±4Ma) (Baldwin et al., 1999). Notably the exhumation of the metamorphic complex was synchronous with the obduction of the Nappe Ophiolitique at the Eocene–Oligocene boundary (see stratigraphic constraints below).

5. Geochemistry

An active margin setting for Poya Terrane basalts was proposed by previous workers (Rodgers; Cameron and Paris) on the basis of major elements geochemistry and only a few trace elements. Although the tholeiitic character of Poya Terrane basalts was recognized, some geochemical features such as LILE enrichment were wrongly taken as evidence for a subduction-related origin. Numerous dykes which are clearly of volcanic-arc affinity were also mistaken as an integral part of the Poya Terrane. These dykes are obviously younger as they crosscut both the Poya Terrane and the overlying Nappe Ophiolitique.

From a limited number of samples collected along the west coast, Eissen et al. (1998) recently distinguished four main lava types within the Poya Terrane basalts: (1) "normal" MORB (N-MORB); (2) MORB-dolerite; (3) back-arc basin basalts (MORB-BABB) thought to have erupted within a marginal basin; and (4) alkaline basalt interpreted as remnant intraplate seamounts.

Based upon a much larger sample set, our data indicate that fields for the first three types overlap and reveal a transitional evolution and origin from mixed or heterogeneous sources (see below). Eissen et al. (1998) also included Ca-rich boninite dykes within the Poya Terrane and inferred an overall supra-subduction character for the Poya Terrane basalt. However, the boninite dykes are not affected by the widespread seafloor metamorphic event (M1, see above), and still contain a significant amount of fresh volcanic glass and therefore contrast sharply with their highly altered serpentinite wallrock. As mentioned above these dykes may be related to a younger, possibly Miocene (Black et al., 1994), magmatic episode and therefore, they have to be considered separately. Until now no geochemical data for the Pouebo Terrane mafic rocks were available.

We have analysed more than 110 samples from Poya and Pouebo terranes (77 and 41, respectively) for major and trace elements. Some of the samples analysed were collected previously by P. Maurizot (B.R.G.M.) and R.N. Brothers (1975–1985) and come from the Auckland University collection. The other samples have been collected by the authors since 1992. They have been analysed by ICPMS for trace and rare earth elements in the LGCA (Grenoble University). A detailed analysis of the geochemistry of these rocks is beyond the scope of this paper and the data presented hereafter include a selection of representative samples. Twenty-two samples that are representative of the overall variation of geochemical features in Pouebo and Poya terranes have been selected for Nd and Sr isotope analyses. Isotope analysis has been carried out on a multicollector Finigan MAT 261 spectrometer at the Isotope Geochemistry Laboratory (University of Toulouse) after Nd and Sr separation on cationic and anionic "columns" in Grenoble University.

6. Whole-rock geochemistry

With the exception of alkaline pillow-basalts of the Poya Terrane that crop out in the Pinjen and Foué peninsulas near Poya, most volcanic rocks of the Poya and Pouebo terranes display similar tholeiitic features, i.e.: low K_2O contents ($0.1<K_2O<0.5\%$), flat chondrite-normalised REE patterns with relatively low (La/Yb)n ratios (0.41<(La/Yb)n<1.93), and Th/Ta ratios. Tholeiites in both terranes display prominent variability between "LREE-undepleted" and "LREE-depleted" end-members (Fig. 8).

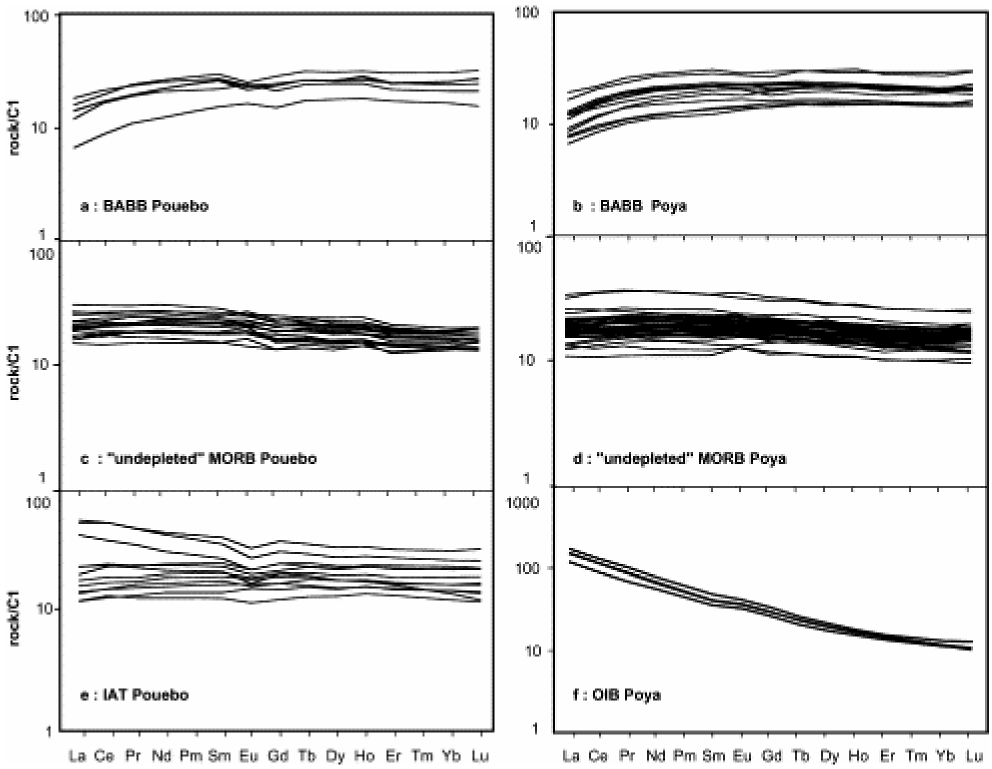

Fig. 8. REE chondrite-normalised spidergram (Evensen et al., 1978) to show the geochemical similitude of Poya and Pouebo terranes mafic rocks, coexistence in both terranes of the two end-members, i.e. "undepleted" and "depleted" MORB (here called BABB, see Fig. 9), and predominance of the "undepleted" end-member. (e) depicts the island arc tholeiites that are only found in the Pouebo Terrane and (f) the alkali basalts locally found in the Poya Terrane.

The "undepleted type" represents at least 80% of samples analysed. This type has been described as MORB 1 and MORB 2 by Eissen et al. (1998); it is distributed throughout the Poya Terrane on both east and west coasts, as well as in the Pouebo mélange. In the Poya Terrane, the undepleted type comprises basalts, dolerites and gabbros bearing low-Ti and Al augite and labrador-andesine phenocrysts in a microlitic groundmass. These mafic magmatic rocks (SiO_2=46–49 wt.%; MgO=6–10 wt.%, TiO_2=0.87–2.35 wt.%) display slightly enriched to slightly depleted, relatively flat chondrite-normalised LREE patterns (0.84<La/Sm)n<1.17; 0.96<(La/Yb)n<1.43 (Fig. 8c and d). On the expanded REE and trace spider diagrams, they also display enrichment in HFS elements (Th, Ta, Nb) (Fig. 9c and d), and moderately high Nb/Y and Th/Nb ratios (0.14<Nb/Y<0.23; 0.05<Th/Nb<0.13) (Table 1) typical of "enriched-MORB" lavas (E-MORB); LILE elements similarly display enrichment with respect to MORB values, in spite of erratic variations due to hydrothermal/metamorphic alteration and weathering. Similarly, "undepleted" basalts of the Pouebo Terrane (SiO_2=48–52 wt.%; MgO=5.9–8.8 wt.%, TiO_2=1.0–1.8 wt.%) display similar REE patterns (0.89<La/Sm)n<1.52; 1.11<(La/Yb)n<1.83) (Fig. 7c) and trace element ratios (0.13<Nb/Y<0.26; 0.08<Th/Nb<0.11) (Table 2). In the Th/Yb vs. Ta/Yb diagram (Fig. 10) these basalts plot in the undepleted MORB field.

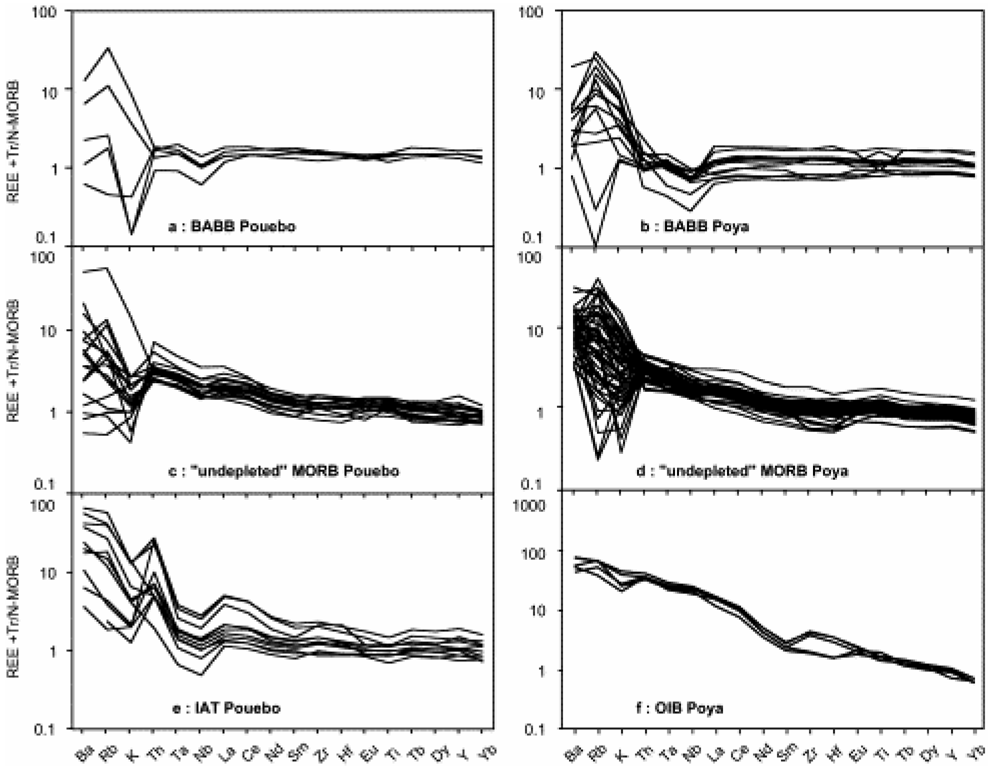

Fig. 9. Expanded-REE spidergrams normalised to the average MORB (Sun and McDonough, 1989) to confirm the geochemical similitude of Poya and Pouebo mafic rocks with the exception of island arc tholeiite (Pouebo Terrane) which display distinctive Nb–Ta negative anomaly and variable Th enrichment (see also Fig. 10).

In both Pouebo and Poya terranes, εNd values back-calculated at 80 Ma (Campanian, see micropaleontological data) vary continuously from 2.77 to 6.72 (Table 3; Fig. 11) within the range of typical E-MORB lavas (Sun and McDonough, 1989). Their $^{87}Sr/^{86}Sr$ ratios ($0.70386<(^{87}Sr/^{86}Sr)_i<0.70599$; $0<\varepsilon Sr<22$) are also consistent with E-MORB affinities, but reveal some Sr mobility due to hydrothermal alteration and weathering (Fig. 12). Such variability indicates prominent mantle source diversity/heterogeneity and is discussed below.

Table 1. Major, trace and rare earth elements geochemistry of selected samples of the Poya Terrane volcanic rocks

Sample		N-MORB			P-MORB								OIB
Location		PO 11	BCE 6	BCE 7	PO 10	BCE 5	BP 84	BP 93	BCE 1	BCE 4	BP 86	BP 95	PO 7
		Voh	Thio	Thio	Voh	Thio	Poindimié	Poindimié	Canala	Thio	Poindimié	Poindimié	Poya
SiO_2	wt.%	46.11	47.88	50.90	47.59	45.95	46.77	47.42	46.17	48.77	46.75	46.99	38.21
TiO_2	wt.%	1.17	1.25	1.33	1.49	1.40	1.27	1.65	1.42	1.39	1.37	0.83	2.17
Al_2O_3	wt.%	16.06	14.55	13.83	14.19	13.90	15.22	12.89	13.10	14.16	14.12	13.58	12.07
Fe_2O_3	wt.%	10.18	10.13	10.43	11.64	12.22	11.38	13.90	11.98	12.53	11.83	9.85	10.71
MnO	wt.%	0.17	0.18	0.19	0.19	0.20	0.17	0.23	0.18	0.24	0.25	0.22	0.19
MgO	wt.%	7.78	7.42	7.76	6.93	7.87	8.48	6.18	6.98	7.98	8.77	9.30	6.38
CaO	wt.%	13.17	11.65	8.48	11.11	9.83	10.72	11.12	11.41	8.96	9.12	12.94	14.23
Na_2O	wt.%	3.04	3.93	4.79	3.90	3.85	3.04	2.84	3.14	3.74	4.31	2.46	2.13
K_2O	wt.%	0.40	1.80	0.37	0.27	0.15	0.31	0.59	0.19	0.88	0.60	0.15	1.74
P_2O_5	wt.%	0.12	0.12	0.14	0.14	0.12	0.12	0.12	0.13	0.13	0.12	0.07	0.62
LOI	wt.%	1.90	1.47	3.10	2.57	4.09	3.85	2.94	3.90	2.95	3.81	3.44	9.60
Total	wt.%	100.10	100.38	101.32	100.02	99.58	101.33	99.89	98.61	101.73	101.04	99.83	98.05
Ba	ppm	26	66	32	46	24	38	77	46	78	71	49	271
Rb	ppm	5	14	6	3	0	4	9	1	24	16	2	29
Sr	ppm	261	127	59	233	29	190	63	62	206	26	435	527
Ta	ppm	0.27	0.38	0.20	0.39	0.39	0.32	0.52	0.39	0.32	0.30	0.27	2.88
Th	ppm	?	0.15	0.18	0.42	0.44	0.40	0.56	0.49	0.40	0.37	0.27	4.44
Zr	ppm	11	57	94	91	92	80	80	88	86	78	41	149
Nb	ppm	1.64	2.55	2.30	4.09	4.22	4.04	6.09	4.40	3.92	4.19	3.26	44.32
Y	ppm	25	27	35	26	28	22	29	28	26	26	17	21
Hf	ppm	1.63	1.80	2.62	2.57	2.59	2.25	2.30	2.44	2.32	2.22	1.08	3.26
V	ppm	265	213	291	327	295	266	365	325	297	305	245	195
Cr	ppm	324	336	191	51	305	248	12	292	301	404	284	324
Ni	ppm	109	128	78	89	93	134	57	100	94	120	123	143
Co	ppm	44	46	43	42	49	46	46	47	46	45	47	41
U	ppm	0.02	0.13	0.05	0.11	0.12	0.11	0.15	0.13	0.10	0.10	0.06	0.99
Pb	ppm	0.11	0.27	0.34	0.71	2.02	0.04	0.26	0.00	0.70	2.94	2.48	0.35
Cs	ppm	4.71	0.28	0.60	0.16	0.18	0.17	0.68	0.39	0.90	0.36	0.40	0.81
La	ppm	2.23	3.09	2.08	4.31	4.06	3.94	4.97	4.69	4.17	4.06	2.62	38.30
Ce	ppm	7.57	9.52	7.32	12.25	11.19	10.72	12.82	12.34	11.20	10.72	6.73	73.40
Pr	ppm	1.34	1.65	1.37	1.95	1.78	1.69	1.96	1.91	1.77	1.72	1.04	8.36
Nd	ppm	7.12	8.51	7.68	9.73	8.95	8.52	9.60	9.49	8.93	8.52	5.17	31.44
Sm	ppm	2.50	2.88	2.89	3.08	2.91	2.66	3.09	3.03	2.85	2.73	1.68	6.35
Eu	ppm	0.97	1.14	1.05	1.17	1.09	1.01	1.22	1.16	1.07	1.07	0.73	2.13
Gd	ppm	3.28	3.64	4.02	3.80	3.66	3.52	4.31	3.77	3.85	3.72	2.39	6.04
Tb	ppm	0.62	0.67	0.78	0.69	0.68	0.60	0.76	0.69	0.65	0.66	0.42	0.89
Dy	ppm	4.19	4.41	5.44	4.49	4.51	3.75	4.86	4.59	4.16	4.26	2.68	4.97
Ho	ppm	0.93	0.96	1.23	0.97	0.99	0.81	1.06	0.99	0.90	0.92	0.60	0.95
Er	ppm	2.64	2.68	3.54	2.70	2.78	2.24	3.01	2.76	2.52	2.59	1.68	2.40
Yb	ppm	2.49	2.49	3.45	2.54	2.72	2.11	2.90	2.63	2.37	2.39	1.61	1.93
Lu	ppm	0.41	0.40	0.58	0.41	0.43	0.32	0.44	0.43	0.37	0.37	0.24	0.26
(La/Sm)n		0.56	0.68	0.45	0.88	0.88	0.93	1.01	0.97	0.92	0.94	0.98	3.80
(La/Yb)n		0.60	0.84	0.41	1.15	1.01	1.26	1.16	1.20	1.19	1.15	1.10	13.39
(Ce/Yb)n		0.79	0.99	0.55	1.25	1.06	1.31	1.14	1.21	1.22	1.16	1.08	9.84
(Gd/Yb)n		1.06	1.18	0.94	1.21	1.09	1.35	1.20	1.16	1.31	1.26	1.20	2.53
Nb/Y		0.06	0.10	0.07	0.15	0.15	0.18	0.21	0.16	0.15	0.16	0.19	2.12
Zr/Y		0.43	2.16	2.67	3.46	3.31	3.62	2.73	3.12	3.26	3.04	2.42	7.16
Th/Yb			0.06	0.05	0.17	0.16	0.19	0.19	0.19	0.17	0.15	0.17	2.30
Ta/Yb		0.06	0.09	0.06	0.15	0.14	0.15	0.18	0.15	0.14	0.13	0.17	1.49

Sample	N-MORB			P-MORB								OIB
Location	PO 11	BCE 6	BCE 7	PO 10	BCE 5	BP 84	BP 93	BCE 1	BCE 4	BP 86	BP 95	PO 7
	Voh	Thio	Thio	Voh	Thio	Poindimié	Poindimié	Canala	Thio	Poindimié	Poindimié	Poya
Zr/Nb	6.71	22.49	40.72	22.36	21.80	19.89	13.13	19.95	21.85	18.70	12.54	3.37
MgV		0.63	0.64	0.58	0.60	0.64	0.51		0.60	0.64	0.69	
Nb/Ta	6.07	6.71	11.50	10.49	10.82	12.63	11.71	11.28	12.25	13.97	12.07	15.39

Table 2. Major trace and rare earth element geochemistry of selected Pouebo volcanic rocks

		N-MORB		P-MORB		
Sample		BP 26	BP 72	BP 27	BP 50	NCH 1
Location		Pouébo	Pam	Pouébo	Pouébo	Pouébo
SiO_2	wt.%	51.09	50.19	48.27	47.91	46.26
TiO_2	wt.%	1.69	1.77	1.52	1.22	1.07
Al_2O_3	wt.%	13.73	13.46	14	14.52	14.27
Fe_2O_3	wt.%	12.25	13.63	13.06	12.03	11.45
MnO	wt.%	0.17	0.18	0.18	0.22	0.18
MgO	wt.%	6.88	6.92	7.37	7.79	8.08
CaO	wt.%	10.13	9.86	9.89	11.09	12
Na_2O	wt.%	2.87	2.96	3.68	4.22	2.74
K_2O	wt.%	0.01	0.01	0.15	0.1	0.19
P_2O_5	wt.%	0.16	0.2	0.15	0.15	0.12
LOI	wt.%	0.85	0.7	1.55	0.58	3.47
Total	wt.%	99.83	99.88	99.82	99.83	99.83
Ba	ppm	14.2	6.94	31.9	60.5	100
Rb	ppm	1.44	0.99	6.54	2.58	4.13
Sr	ppm	47	84	88	142	108
Ta	ppm	0.12	0.20	0.33	0.27	0.27
Th	ppm	0.11	0.20	0.34	0.30	0.31
Zr	ppm	109	115	89	69	59
Nb	ppm	1.42	2.53	4.11	3.36	3.54
Y	ppm	41.7	42.0	27.5	21.2	22.3
Hf	ppm	2.92	3.07	2.33	1.82	1.53
V	ppm	388	380	335	287	325
Cr	ppm	244	138	323	344	362
Ni	ppm	120	105	123	103	147
Co	ppm	37.5	49.8	42.2	42.7	43.7
U	ppm	0.06	0.09	0.11	0.11	0.11
Pb	ppm	1.18	1.17	1.99	5.73	0.94
Cs	ppm	nd	nd	nd	nd	nd
La	ppm	2.96	3.49	4.80	4.51	3.60
Ce	ppm	10.68	11.00	13.20	11.12	9.20
Pr	ppm	1.86	1.91	2.05	1.71	1.42
Nd	ppm	10.57	10.56	10.58	8.70	7.06
Sm	ppm	4.04	4.11	3.32	2.79	2.29
Eu	ppm	1.27	1.34	1.22	1.15	0.93
Gd	ppm	4.84	4.87	3.69	3.06	2.69
Tb	ppm	1.00	1.00	0.73	0.57	0.54
Dy	ppm	6.72	6.64	4.67	3.69	3.52
Ho	ppm	1.60	1.64	1.00	0.85	0.80
Er	ppm	4.12	4.17	2.54	2.05	2.10
Yb	ppm	4.22	4.26	2.50	2.16	2.19
Lu	ppm	0.70	0.69	0.41	0.34	0.34
(La/Sm)n		0.46	0.53	0.91	1.02	0.99
(La/Yb)n		0.47	0.55	1.30	1.41	1.11
(Ce/Yb)n		0.66	0.67	1.37	1.33	1.09
(Gd/Yb)n		0.93	0.92	1.19	1.14	0.99
Nb/Y		0.03	0.06	0.15	0.16	0.16
Zr/Y		2.61	2.74	3.23	3.23	2.64
Th/Yb		0.026	0.047	0.136	0.139	0.142
Ta/Yb		0.028	0.047	0.132	0.125	0.123

	N-MORB		P-MORB		
Sample	BP 26	BP 72	BP 27	BP 50	NCH 1
Location	Pouébo	Pam	Pouébo	Pouébo	Pouébo
Zr/Nb	76.76	45.45	21.63	20.39	16.61
MgV	11.83	12.65	12.45	12.44	13.11
Nb/Ta	11.83	12.65	12.45	12.44	13.11

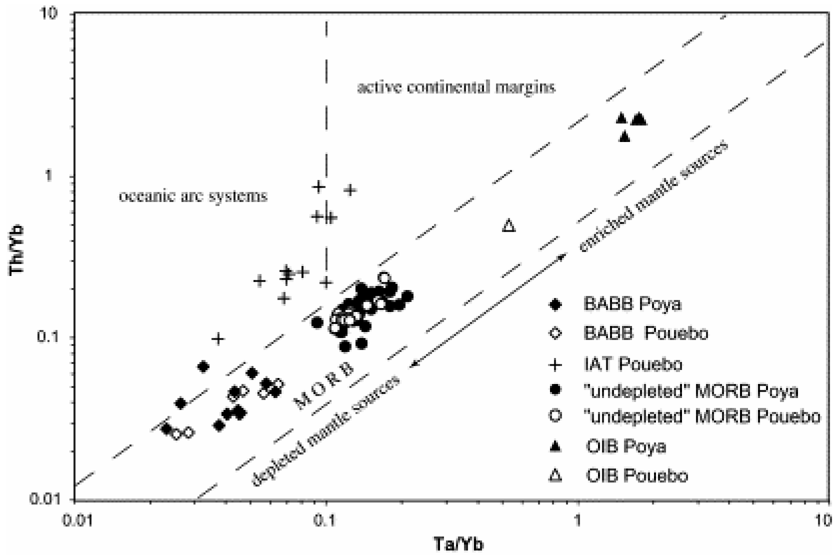

Fig. 10. Th/Yb vs. Ta/Yb diagram showing the location of most mafic rocks of the Pouebo and Poya terranes in the mantle array, with the exception of those derived from the Diahot Terrane which distinctively plot in the arc-related field.

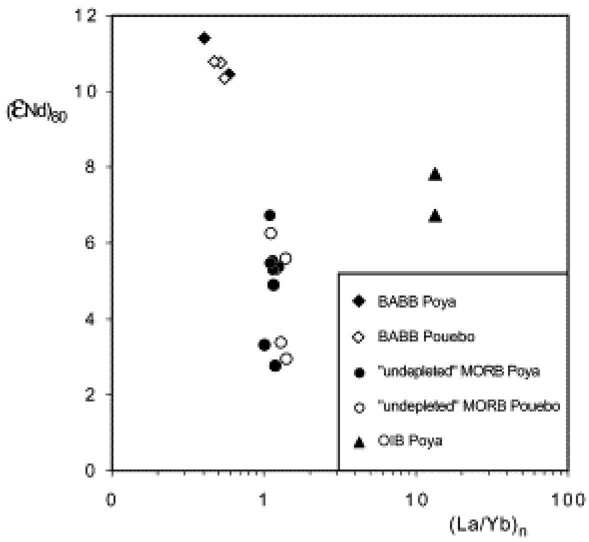

Fig. 11. εNd_{80} vs. (La/Yb)n plot to show the close similitude of Poya and Pouebo mafic rocks and the wide range of Nd isotopic ratios especially in the "undepleted" group, most likely depicting a mixed source origin. Due to the limited number of samples analysed, it is not certain that Nd isotopic compositions form distinct groups or one single array as it may be suspected from the REE patterns (see Fig. 8).

Table 3. Isotope geochemistry of Poebo and Poya terranes basalt

	Poya terrane											
Rock type	BABB	BABB	BABB	P-MORB	P-MORB	P-MORB	P-MORB	P-MORB	P-MORB	P-MORB	OIB	OIB
Sample	BCE 7	PO 11	BCE 6	BCE 4	BCE 5	BP 93	BCE 1	BP 84	BP 95	PO 10	PO7	PO 21
Age (Ma)	80	80	80	80	80	80	80	80	80	80	80	80.00
Sm	2.89	2.50	2.88	2.85	2.91	3.09	3.03	2.66	1.68	3.08	6.35	6.30
Nd	7.68	7.12	8.51	8.93	8.95	9.60	9.49	8.52	5.17	9.73	31.44	32.00
Rb	5.60	4.81	13.78	24.27	0.15	8.60	1.30	4.14	1.87	3.09	28.80	
Sr	59.00	261.00	126.77	205.94	28.66	62.72	62.00	190.00	435.19	232.79	526.70	
$(^{143}Nd/^{144}Nd)_{mes}$	0.51291	0.51301	0.51307	0.51278	0.51281	0.51289	0.51291	0.51291	0.51292	0.51292	0.51300	0.51294
$(^{87}Sr/^{86}Sr)_{mes}$	0.70391	0.7046	0.70453	0.70498	0.70486	0.7043	0.70515	0.70477	0.70442	0.70428	0.70314	nd
$^{147}Sm/^{144}Nd$	0.23151	0.21603	0.20822	0.19634	0.20003	0.19803	0.19643	0.19208	0.19992	0.19475	0.12211	0.12112
$(^{143}Nd/^{144}Nd)_i$	0.51279	0.51290	0.51296	0.51268	0.51271	0.51279	0.51281	0.51281	0.51282	0.51282	0.51294	0.51254
Eps (Nd)	4.95	7.06	8.31	2.77	3.32	4.90	5.31	5.35	5.47	5.52	7.84	6.72
$^{87}Rb/^{86}Sr$	0.26454	0.0533	0.3144	0.32850	0.01514	0.39658	0.05845	0.06302	0.01198	0.03839	0.15813	
$(^{87}Sr/^{86}Sr)_i$	0.70361	0.70454	0.70417	0.70461	0.70484	0.70385	0.70508	0.7047	0.70441	0.70423	0.70296	
$^{87}Sr/^{86}Sr$	0.70441	0.70441	0.70441	0.70441	0.70441	0.70441	0.70441	0.70441	0.70441	0.70441	0.70441	
Eps (Sr)	− 11.31	1.87	− 3.31	2.85	6.13	− 7.89	9.62	4.11	0.01	− 2.48	− 20.55	

	Pouebo terrane								
Rock type	OIB	IAT	BABB	BABB	BABB	P-MORB	P-MORB	P-MORB	P-MORB
Sample	PO 21	NBP 1	BP 72	NBP 8	BP 26	BP 50	BP 27	NCH 1	NCH 2
Age (Ma)	80.00	80.00	80	80.00	80	80	80	80	80
Sm	6.30	2.59	4.11	4.64	4.04	2.79	3.32	2.29	2.68
Nd	32.00	7.59	10.56	12.61	10.57	8.70	10.58	7.06	8.37
Rb			0.99		1.44	2.58	6.54	4.13	
Sr			84.00		47.00	142.00	88.00	108.00	
$(^{143}Nd/^{144}Nd)_{mes}$	0.51294	0.51308	0.51319	0.5132	0.51321	0.51279	0.51281	0.51296	0.51293
$(^{87}Sr/^{86}Sr)_{mes}$	nd	nd	nd	nd	nd	nd	nd	nd	nd
$^{147}Sm/^{144}Nd$	0.12112	0.21019	0.23947	0.22659	0.23517	0.19729	0.19305	0.19956	0.19699
$(^{143}Nd/^{144}Nd)_i$	0.51254	0.51296	0.51306	0.51309	0.51309	0.51269	0.51271	0.51286	0.51282
$^{143}Nd/^{144}Nd$	0.12112	0.21019	0.51254	0.22659	0.51254	0.51254	0.51254	0.51254	
Eps (Nd)	6.72	8.39	10.33	10.74	10.77	2.96	3.39	6.25	5.60
$^{87}Rb/^{86}Sr$									
$(^{87}Sr/^{86}Sr)_i$									
$^{87}Sr/^{86}Sr$									
Eps (Sr)									

In the metamorphosed Pouebo melange, some samples of garnet "glaucophanite" or eclogite differ significantly from the dominant E-MORB type. They display undepleted to slightly enriched LREE patterns (Fig. 8e; $0.79<(La/Sm)_n<1.56$; $0.82<(La/Yb)_n<2.12$), prominent Nb–Ta depletion and Th-LILE enrichments (Fig. 9e; $0.05<Nb/Y<0.17$; $0.17<Th/Nb<0.69$). These features are similar to that of island-arc tholeiite and suggest a volcanic arc origin (see also Fig. 10). Such geochemical features that are never found in the Poya Terrane basalt are similar to that of some metavolcanic rocks of the Diahot Terrane (Picard; Gros and Debeuf) and their origin will be discussed below.

The "depleted type" ($0.43<(La/Yb)_n<0.69$), which is similar to the MORB type 3 of Eissen et al. (1998), is only found in three areas of the Poya Terrane, in the Gatope peninsula (near Voh) at Kone (west coast) and in the Thio area (east coast). It is also found as metamorphosed boulders in the Pouébo mélange at Anse Ponandou, Colnett, Balade and Pam Peninsula (Fig. 1) where it coexists with the "undepleted" type. In the Poya Terrane, lavas contain augite microphenocrysts with higher Al and Ti contents than those of enriched type lavas. In both terranes, the depleted-type rocks (SiO_2=46–50 wt.%; MgO=6–8 wt.%; TiO_2=0.96–1.57 wt.%) display N-MORB-like chondrite-normalised REE patterns with marked LREE depletion. The "depleted" type is clearly distinctive of the "undepleted" one ($0.45<(La/Sm)_n<0.69$;

0.41<(La/Yb)n<0.73, in Poya Terrane rocks; 0.43<(La/Sm)n<0.68; 0.4<(La/Yb)n<0.76 in Pouebo Terrane rocks) (Fig. 8a and b). Expanded REE diagrams in both terranes display typical N-MORB-like patterns with respect to HFS elements (Fig. 9d), with the exception of a slight Nb–Ta depletion inferring lower Nb/Y ratios (0.03<Nb/Y<0.08) (Table 1 and Table 2). Save for K that may have been leached during weathering, LILE elements are enriched with respect to MORB values. Nb depletion and LILE enrichment are characteristic features of a magma source that has undergone hydration and metasomatism in a supra-subduction setting. The features of the “depleted type” are transitional between island-arc tholeiite and N-MORB and are similar to those of lavas erupted in a back-arc environment (BABB). εNd values for the “depleted type” range from 10.33 to 10.77 in the Pouebo Terrane, and from 6.98 to 11.58 in the Poya Terrane (Table 3). Values ca. 10 are consistent with a N-MORB-type source for the Pouebo rocks (Fig. 11).

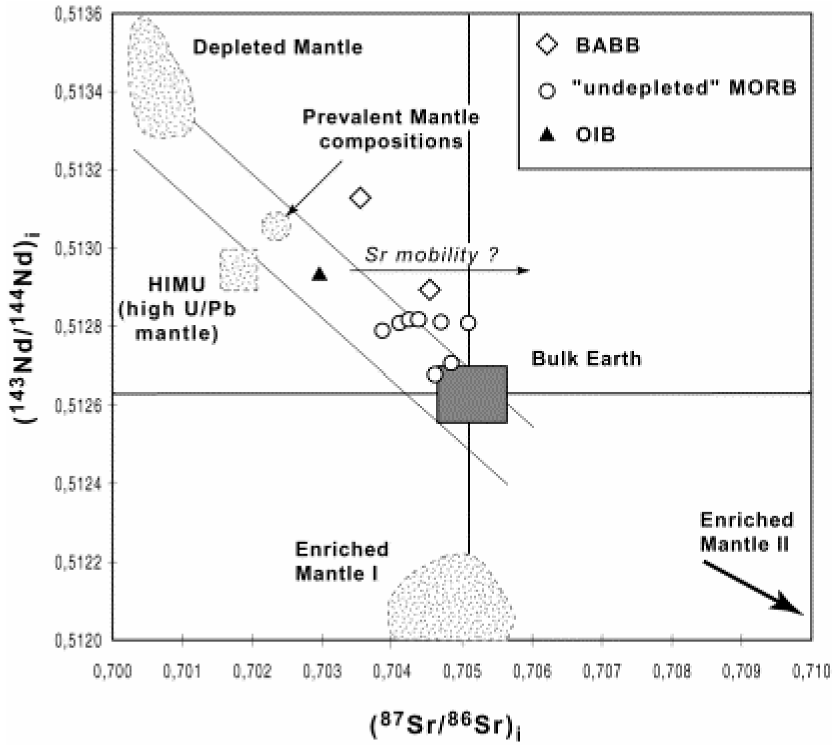

Fig. 12. On the $(^{143}Nd/^{144}Nd)_i$ vs. $(^{87}Sr/^{86}Sr)_i$ diagram (Zindler and Hart, 1986), the basalts of the Poya Terrane display high $^{143}Nd/^{144}Nd_i$ and low $^{143}Nd/^{144}Nd_i$ ratios indicative of a mantle origin in spite of some Sr mobility due to hydrothermal alteration (owing to their high metamorphic grade and prominent Sr mobility, Sr isotopic compositions of the Pouebo mafic rocks have not been determined).

Alkaline volcanic rocks mainly occur at one locality on the west coast (Foué and Pinjen peninsulas, Fig. 1). They include highly vesicular pillow-basalts and pillow breccia with pinkish Al and Ti-rich diopside; andesine-labrador and Fe–Ti oxides microphenocrysts in a devitrified groundmass. They have suffered strong weathering and rubefaction. Major element distribution has been greatly disturbed and should be assessed with care. However, mineral chemistry, high TiO_2 contents (1.9–2.4 wt.%), HREE depleted and LREE enriched patterns (3.27<(La/Sm)n<3.8; 10.7<(La/Yb)n<13.9) (Fig. 8f) demonstrate their alkaline character and origin from a garnet-bearing peridotite deep source. These basalts that also display prominent enrichment in Ta, Nb and Th (Fig. 9 and Fig. 10) and εNd=6.72–7.82 (Fig. 11). They most probably represent OIB-type lavas erupted in an intraplate environment. These pillow-basalts display micritic interpillow material that contrasts with the bathyal siliceous argillite and red chert that are usually found with Poya basalts, and was deposited at a shallower depth. These volcanic rocks are thus likely to have erupted as part of an intraplate seamount. A single mafic rock sample with similar affinities has been found in the Pouebo metamorphosed melange (Pam Peninsula).

7. Magma sources and geodynamic setting

The coexistence in both terranes of "depleted" and "undepleted" type lavas suggests that they originated from at least two different mantle sources. More specifically, within each rock-type, the REE patterns are crosscutting, suggesting either different degrees of partial melting of the same source, or genesis from slightly different sources. One of the most striking features of the tholeiitic rocks is the great diversity in Nd isotopic signatures.

In the "undepleted" type, εNd varies from 2.77 to 6.66 whereas in the "depleted type", εNd varies from 6.98 to 11.58. On the εNd vs. (La/Yb)n diagram (Fig. 11), representative points of both Poya and Pouebo tholeiites are spread along one single trend which infers a progressive evolution between an enriched E-MORB-type end-member (BCE 4; εNd=2.77) and a depleted N-MORB-like end-member (BCE 7; εNd=11.58). Such a trend may result from mixing of different mantle sources. In spite of Sr mobility, the distribution for the Poya lavas on the ($^{143}Nd/^{144}Nd$)i vs. ($^{87}Sr/^{86}Sr$)i diagram (Fig. 12) infers mixing between a superficial depleted mantle and a deeper Enriched Mantle II (EMII) as characterized by Zindler and Hart (1986) with possible contribution of HIMU (high U/Pb) reservoir. A contamination by the continental crust has been suggested by Eissen et al. (1998) to account of the Nb–Ta depletion in BABB (high Th/Nb ratio relative to MORB) and low εNd values ($3.1<\varepsilon Nd<4.6$, in Cameron, 1989). However, MORB-like REE patterns; low Th (Th/Nb=0.02–0.15; (Th/Ta)n=1) and our εNd data ($6.98<\varepsilon Nd<11.58$ for the BABB-like basalt) are not consistent with such a possibility; therefore, lower crust contamination appears to be extremely weak not to say unnoticeable.

Hence, considering the data presented above, it appears certain that Poya Terrane basalts and Pouebo mafic rocks erupted either within one single basin, or at least, in basins formed in identical settings. Indeed, save for IAT, these terranes (1) contain rocks with similar geochemical and isotopic features and (2) display the same diversity.

From the occurrence of arc-related basalt (IAT) in the Pouebo melange only, we cannot infer the geodynamic setting of the basin in which the Poya Terrane originated. Indeed, parts of the Pouebo melange may derive from other terranes (e.g. the Diahot Terrane, see discussion below). Therefore, we only consider the Poya Terrane basalt features in order to decipher the geodynamic setting of the South Loyalty basin from which it originated.

All these rocks originally formed the bottom of a marginal basin. The occurrence of minor (ca. 5%) "BABB-like" tholeiite with slight Nb depletion may infer formation in a back-arc setting as suggested by Eissen et al. (1998); or alternatively, from a mantle source that had been previously affected by suprasubduction processes (e.g. the Paleozoic–Mesozoic East-Gondwana active margin). The occurrence of slightly older volcanic arc rocks in the basement terranes is a rather sound evidence for the existence of a volcanic arc immediatly prior to marginal basin opening. However, the predominance of basalt with E-MORB affinities generated from to a relatively deep mantle source suggests that this basin did not evolve in response to a simple back-arc opening process. In the absence of arc-related volcanosedimentary rocks in the cover of the Poya Terrane, it is not possible to solve this address.

Similar features may be found in other recent back-arc basins of the Southwest Pacific, i.e. the North Fiji basin (Nohara and Eissen) where the oceanic crust is accreted at RRR triple junctions where MORB-like and enriched lavas coexist due to the interference of normal shallow and deeper, plume-like, mantle sources. Alkaline (OIB) lavas are likely to have erupted independantly from the basin formation and may represent hot spot-related seamounts younger than the tholeiite that formed their basement.

8. Time constraints

8.1. Poya Terrane

The boundaries of the Poya Terrane are generally faulted or concealed; therefore, due to a lack of stratigraphic evidence, the age and tectonic setting of this terrane have long been a matter of debate. Although K–Ar whole-rock ages determined for altered and/or metamorphosed basalts suggest Paleocene to mid-Eocene apparent ages, a late Eocene isochron age of 38.5±1.5 Ma has been proposed (Guillon and Gonord, 1972). This result has been questioned (Rodgers, 1975), but it has been widely accepted as an evidence for the eruption age of the basalts (Paris and Kroenke). A late Eocene age is untenable for reasons outlined below, and the Paleocene–early Eocene dates must be taken to represent minimum ages that only record an incomplete resetting of the K–Ar system due to metamorphic and/or alteration events (Eissen et al., 1998).

Minerals and rock clasts characteristic of rocks found within the Poya Terrane are widespread in the upward-coarsening late Eocene foreland flysch, and monogenetic volcanoclastic breccia (Routhier; Gonord; Gonord and Espirat). Mafic olistoliths, within these units, have been misinterpreted as evidence for late Eocene volcanic activity (“Volcanic Flysch” of Routhier, 1953). However, a lack of evidence for volcanic activity and feeding dykes in the underlying Paleocene–Eocene pelagic sediments contradicts this interpretation. In addition, most of the volcanic clasts have been affected by a greenschist facies metamorphism (similar to that of the Poya Terrane rocks) which is absent in the matrix. The alternative model of a Late Cretaceous in situ volcanic activity (Maurizot and Maurizot) also does not fit the drastically different geochemical features (volcanic arc) of the Cretaceous volcanic rocks in the Noumea area (calc-alkaline to alkaline; Black; Gros and Debeuf), and those of the Diahot Terrane (arc tholeiites to alkaline; Picard and Debeuf).

From the discussion above, it clearly appears that the Poya Terrane rocks may not have been extruded in situ and an allochthonous character may be considered as a reasonable hypothesis which is to be confirmed by paleontological and petrologic evidence. In order to better constrain the age of the Poya Terrane, over 200 samples (approximtely 300 g each) of pelagic red chert and siliceous siltstone interbedded with, and overlying the basalts have been collected from throughout New Caledonia and processed for radiolarians. Recovery of identifiable material was generally very poor. Nevertheless, we are able to report well preserved Upper Cretaceous and Paleogene radiolarian faunas from several localities (Fig. 6). Sample localities are reported using the Institut Géographique National 1:50,000 map series. Map sheet numbers and names together with grid references (GR easting/northing) are given. New Caledonian faunas were compared with other known radiolarian faunas from ocean drilling and on-land sequences of similar age (Pessagno; Pessagno; Pessagno; Pessagno;

Foreman; Foreman; Foreman; Foreman; Foreman; Riedel; Dumitrica; Nakaseko; Nakaseko; Mizutani; Sanfilippo; Hollis; Nishimura; Okamura; Hollis; Sporli; Urquhart; Urquhart and Strong).

Upper Cretaceous (Campanian) faunas are the most common and were mostly recovered from along the west coast. From a small quarry on the eastern side of RM20 red ribbon-bedded cherty sediments at Tribu de Boyen [4810 Voh GR 4594/76956] north–west of Voh yielded the following taxa: *Amphipyndax stockii* (Campbell and Clark), *Amphipyndax pseudoconulus* (Pessagno), *Theocampe urna* (Foreman), *Archaeodictyomitra lamellicostata* (Foreman), *Phaseliformas carinata* Pessagno, *Alievium superbum* (Squinabol), *Dictyomitra andersoni* (Campbell and Clark), and *Sciadiocapsa* (?) *rumseyensis* Pessagno. The presence of these taxa suggests assignment to the lower to middle Campanian *A. pseudoconulus* zone of Sanfilippo and Riedel (1985). A less diverse fauna of similar age was recovered from the northern outskirts of Noumea in samples of red ribbon bedded cherts collected from a road cutting at the Ondémia junction on the Noumea Paige near Paita [4833 Nouméa GR6377/75518] and includes: *A. stockii* (Campbell and Clark), *A. lamellicostata* (Foreman) and *Crucella espartoensis* Pessagno. Red cherts and other siliceous sediments of the Poya Terrane are widespread along the main highway (RT1) northwest of Bourail towards Poya. The best Cretaceous faunas were recovered from outcrops of steeply dipping red chert and siliceous shale located approximately 3 km west of Nandai military base [4820 Mé Maoya GR5450/76224] and include: *T. urna* (Foreman), *Cryptamphorella conara* (Foreman), *A. stockii* (Campbell and Clark), *Archaeospongoprunum nishiyamae* Nakaseko and Nishimura, *Pseudoaulophacus florensis* Pessagno, *A. lamellicostata* (Foreman), *D. andersoni* (Campbell and Clark), *Dictyomitra koslovae* (Foreman), *Amphisphaera privus* (Foreman), *Alievium gallowayi* Pessagno, *Amphipyndax tylotus* Foreman, and *Lithomellisa* (?) cf. *Lithomellisa* (?) of Pessagno (1975). The taxa present suggest assignment to the (uppermost Cretaceous) upper Campanian to Maestrichtian *Amphipyndax tylotus* zone of Sanfilippo and Riedel (1985). Poorly preserved material of Cretaceous affinity was also extracted from blueschist facies rocks along the northeast coast of the island at Mangalia and supports correlation of basalts exposed on the east coast of New Caledonia with similar lithologies more widely exposed on the west coast.

Paleogene faunas have been recovered from three localities at Mt. Dore, Kone and Taom. *Bekoma* sp., *Phormocyrtis striata exquisita* (Kozlova), *Lychnocanoma* sp., and *Buryella tetradica* Foreman are present in samples collected from east of Noumea in red siliceous shale and chert located in a small quarry which truncates the Baie de Plum east of Mont Dore along RT2 bis [4836 Mt. Dore GR6668/75343]. *P. striata striata* (Kozlova), and *B. tetradica* Foreman occur in samples from new road cuttings between Pouembout and Kone [4814 Pouembout GR4884/76671] located approximately 50 m to the east of the start of the new RM4 Kone to Tiwaka highway in a west-facing faulted sequence of red chert, siliceous siltstone, brown tuffaceous chert and manganese oxide stained sediments. Samples collected from along RT1 approximately 2 km east of Taom in red chert and siltstone tectonically interleaved with basalt [4810 Voh GR4551/76947] contain: *Bekoma bidartensis*. Riedel and Sanfilippo, *Rhapalocanium* sp., *Phormocyrtis striata striata* (Kozlova), *Lychnocanoma* sp., *B. tetradica* Foreman, ?*Microsciadiocapsa* sp., and *Lamptonium* cf. *L. pennatum* Foreman (Plate 1 and Plate 2).

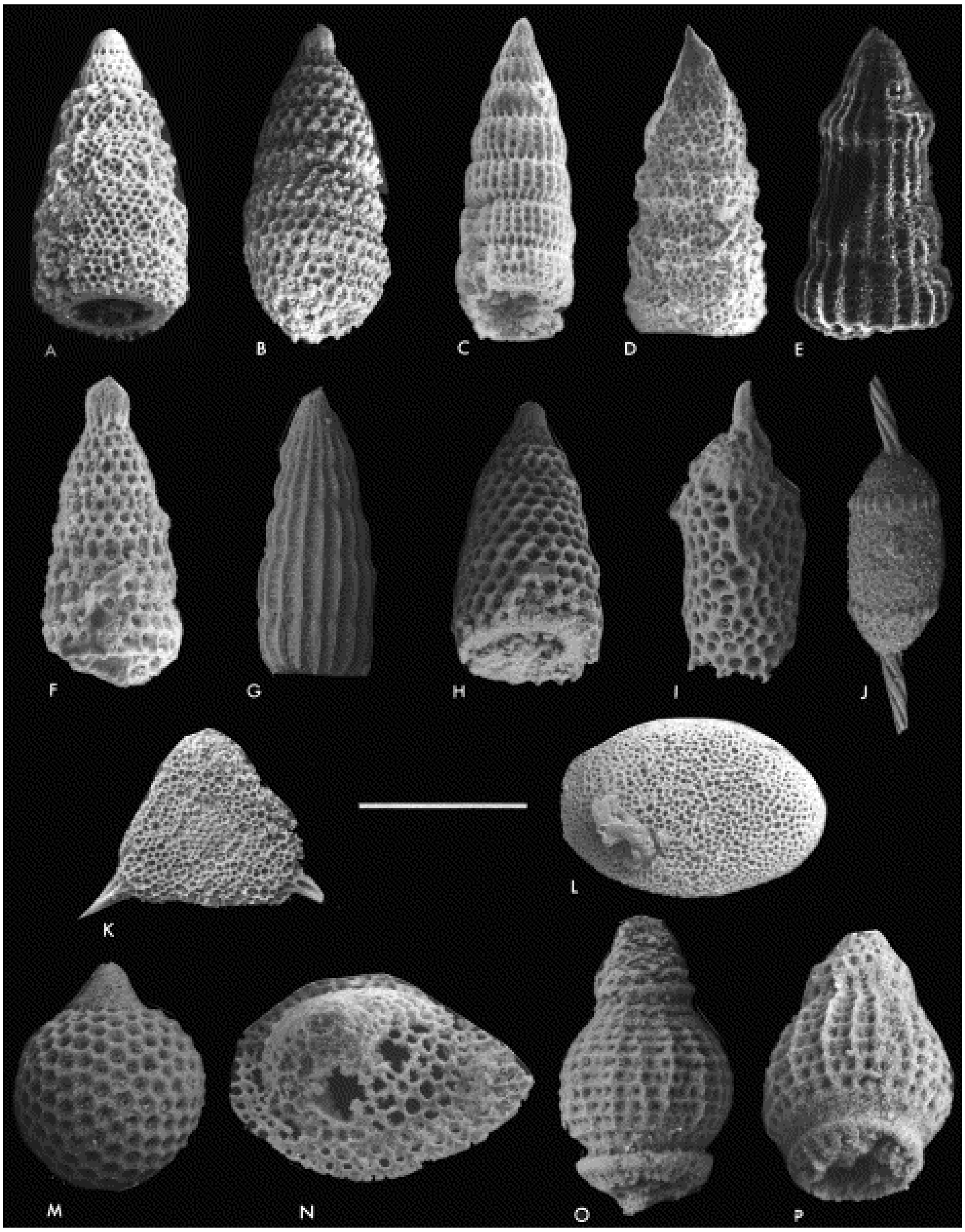

Plate 1. Upper Cretaceous (Campanian) radiolarians from siliceous sediments collected from the Poya Terrane, New Caledonia. (sample location and length of scale bar indicated in parentheses). (A) *Amphipyndax tylotus* Foreman (Bourail) (120 µm); (B) *Amphipyndax stockii* (Campbell and Clark) (Paita) (135 µm); (C) *Dictyomitra andersoni* (Campbell and Clark) (Boyen) (120 µm); (D) *Amphipyndax pseudoconulus* (Pessagno) (Boyen) (135 µm); (E) *Dictyomitra koslovae* (Foreman) (Bourail) (85 µm); (F) *Amphipyndax stockii* (Campbell and Clark), (Boyen) (100 µm); (G) *Archaeodictyomitra lamellicostata* (Foreman) (Bourail) (135 µm); (H) *Amphipyndax stockii* (Campbell and Clark) (Bourail) (100 µm); (I) *Lithomellisa* (?) cf. *Lithomellisa* (?) of Pessagno 1975 (Bourail) (100 µm); (J) *Archaeospongoprunum nishiyamae* Nakaseko and Nishimura (Bourail) (150 µm); (K) *Alievium superbum* (Squinabol) (Boyen) (155 µm); (L) *Phaseliformas carinata* Pessagno (Boyen) (165 µm); (M) *Cryptamphorella conara* (Foreman) (Bourail) (135 µm); (N) *Sciadiocapsa* (?) *rumseyensis* Pessagno (Boyen) (105 µm); (O) *Theocampe urna* (Foreman) (Bourail) (90 µm); (P) *Theocampe urna* (Foreman) (Boyen) (70 µm).

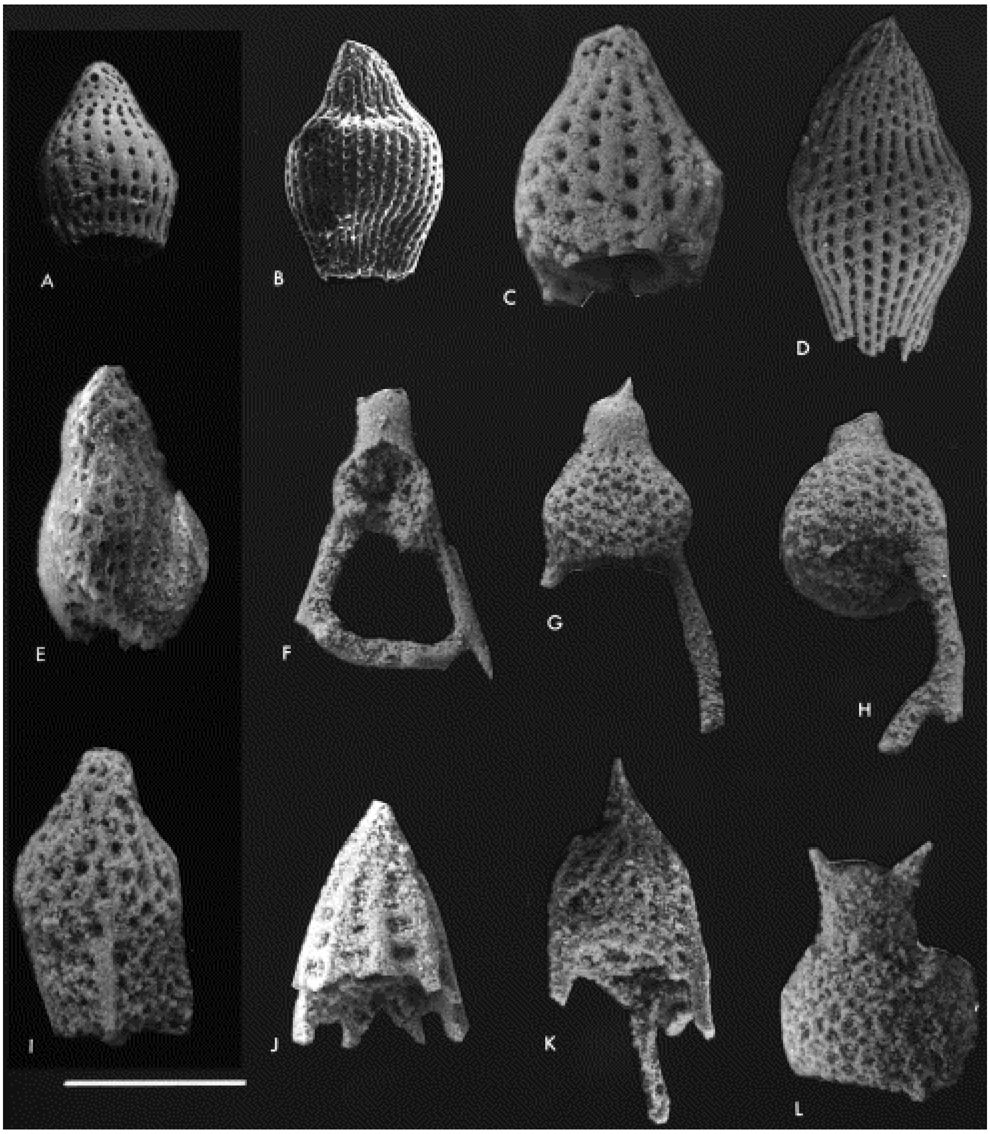

Plate 2. Paleogene (uppermost Paleocene to lower Eocene) radiolarians from siliceous sediments collected from the Poya Terrane, New Caledonia (sample location and length of scale bar indicated in parentheses). (A) *Buryella tetradica* Foreman (Mt. Dore) (85 µm); (B) *Phormocyrtis striata exquisita* (Kozlova), (Mt. Dore) (175 µm); (C) *Buryella tetradica* Foreman (Kone) (60 µm); (D) *Phormocyrtis striata striata* (Kozlova), (Kone) (120 µm); (E) *Phormocyrtis striata striata* (Kozlova), (Taom) (100 µm); (F) *Bekoma bidartensis*. Riedel and Sanfilippo, (Taom) (160 µm); (G) *Bekoma bidartensis*. Riedel and Sanfilippo, (Taom) (130 µm); (H) *Bekoma bidartensis*. Riedel and Sanfilippo, (Taom) (120 µm); (I) *Lamptonium* cf. *L. pennatum* Foreman, (Taom) (80 µm); (J) *Rhapalocanium* sp., (Taom) (120 µm); (K) unidentified nassellarian, (Taom) (100 µm); (L) ?*Microsciadiocapsa* sp. (Taom) (100 µm).

The Paleogene faunas bear greatest similarity with those reported from the lower Eocene of New Zealand (Strong et al., 1995). Some samples maybe as old as the uppermost Paleocene *Buryella tetradica* (RP5) zone of Hollis (1993) but are possibly better correlated with the basal Eocene *Buryella clinata* (RP8) zone of Foreman (1973b). Uncertainty in the precise assignment of these faunas is related in part to their low diversity and in part to the fact that radiolarian faunas of this age are less well studied than those of the Upper Cretaceous or Eocene and younger.

The pinkish micritic interpillow material found associated with OIB lavas at Pinjen peninsula contains *Planorotalites (Globorotalia) pseudomenardii* indicative of the Upper Paleocene

(P4, 56–59 Ma) (Coudray and Gonord, 1966) thus, representing possibly the youngest rocks of the terrane. This is consistent with an intraplate origin for these alkaline lavas that were erupted upon older tholeiites forming their basement.

Both Cretaceous and Paleogene faunas of the Poya Terrane are found in rocks associated with basaltic lithologies. This could be interpreted to indicate eruptive activity over a considerable time. However, we note that many of the localities are severely disrupted by faulting and the original relationships are almost impossible to determine. Nevertheless, radiolarians can be used to constrain the duration of sedimentation within the basin in which the basalts were erupted. The basin existed from the Late Cretaceous (Campanian) until, at least, the latest Paleocene or earliest Eocene. The fact that metamorphosed Late Cretaceous to early Eocene "oceanic" rocks are found overlying unmetamorphosed sediments of the same age range, is evidence which further supports interpretation of allochthonous nature of the Poya Terrane.

8.2. Pouebo Terrane

Because of the high average metamorphic grade, micropaleontologic evidence for the age of the Pouebo Terrane is difficult to obtain. Globigerine "ghosts" have been reported in a metamorphosed limestone lens (or boulder) north of Hienghene (Carroué, 1971), but the sampling locality has never been found again (Maurizot and Maurizot) and this discovery has not been confirmed. Besides the fact that lithologies typical of both Diahot and Poya terranes are reworked in the Pouebo Terrane, indicating a post-Late Paleocene/Early Eocene age, the only indirect age evidence comes from Ar–Ar geochronological data from phengite that indicates a minimum Eocene age (51–40 Ma) for the unroofing of eclogite (Ghent and Baldwin). Thus, a direct stratigraphic correlation of the mafic rocks of the two terranes remains impossible and geochemistry provides the only means to test possible correlation.

9. The sedimentary record of obduction

9.1. Foredeep deposits

The pre-obduction sequence of New Caledonia rests unconformably upon pre-Upper Eocene terranes (Fig. 4), it records the emersion of the Norfolk Ridge in the fore-arc bulge, its progressive subduction below the Loyalty arc and the subsequent obduction. Upper Eocene carbonate rocks lie unconformably upon eroded pre-Cretaceous to Middle Eocene rocks. At the base of the sequence are 20–50 m of neritic foraminiferal limestone locally contains centimetre-size pebbles of reworked underlying basement rocks. Together with large benthic foraminifers, these rocks contain glauconite grains, crinoid and red algae bioclasts consistent with high-energy shallow water deposition that records a drastic change from the pelagic environment that prevailed until the Middle Eocene. In places, neritic limestone laterally changes into and is interbeded with well-sorted deltaic calcarenite. Microfacies distribution allows an irregularly and slowly subsiding platform to be reconstructed (Chiron, 1996). In some places along the west coast (St. Vincent Bay) emersion-related cut-and-fill structures, intraformational conglomerate and rubefaction reveal some tectonic instability within the basin (Gonord, 1977). In contrast, locally within the Central Chain and along the east coast, occurrence of darker clayey facies that bear pelagic foraminifers (globigerinidae and globorotalidae) indicate deeper water deposition and therefore a gentle eastward or north-eastward dip of the platform. A reappraisal of the benthic microfaunas (Chiron, 1996) provides consistent Priabonian ages for the basal limestone, except in one location of the west coast (Uitoë) where the base of the sequence may be correlated with the Ta3–Tb boundary (Upper Bartonian, ca. 37.5±1 Ma) (Odin, 1994).

Following this short-lived neritic episode, hemipelagic sedimentation resumed with marls and calcareous sandy marls directly overlying the basal limestone (Fig. 4). The marls change progressively into alternating sandy marls or calcarenite and marls (the "lower flysch" of Tissot and Noesmoen, 1958), 50–300 m thick, containing Upper Priabonian pelagic foraminifers (Upper P16–P17, 33.7–35 Ma; Chiron, 1996). Clastic minerals in the "lower flysch" are still mainly derived from the basement rocks. In turn, the "lower flysch" is overlain by the "upper flysch" (Tissot and Noesmoen, 1958) with local disconformity (Gonord, 1977). The "upper flysch" is an upward-coarsening clastic sequence topped by an olistostrome that reworks both lithologies of the Poya Terrane and all the components of the sedimentary cover of the New Caledonia Block, including clasts of the Upper Eocene bioclastic limestone and the flysch itself (Fig. 4). The olistostrome is in turn tectonically overlain by the Poya Terrane and finally by the Ophiolitic Nappe. The age of the lower boundary of Noumea–Bourail flysch is constrained by the lower Priabonian (ca. 37 Ma) transgressive limestone at the base of the Upper Eocene sequence (Chiron, 1996). In addition, in the Noumea area, the Ophiolitic Nappe is thrust upon the "Formation de la Cathedrale" which has been correlated with the Upper Priabonian (upper P16–P17; 33.7–35 Ma) (Chiron, 1996). Therefore, the timing of obduction is accurately constrained (at least in the Noumea area) to be post-35, pre-34 Ma.

One of the most striking features of the Upper Flysch is the progressively increasing supply of mafic ophiolitic detritus recording the approach of a mafic allochthon. The typical "rusty" weathering colour of the Upper Flysch is due to the oxidization of basaltic clinopyroxene and titanomagnetite, that first appear as clastic minerals. Angular mafic clasts (i.e. basalt, dolerite and fine-grained gabbro) are widespread in breccias and their abundance and size increase upwards. Together with subordinate red chert, they may locally form the bulk of some hundreds of metre thick coarse breccia bodies (Espirat, 1963). These volcanic rocks display "undepleted" and "depleted" MORB signatures similar to those of Poya Terrane basalt that display the same greenschist-facies mineral assemblages, and are obviously reworked from the Poya Terrane (Cluzel et al., 1997). In the absence of Oligocene taxa, the Upper Flysch is considered to be Upper Eocene (Tissot; Gonord and Paris) although microfauna appears to have been reworked from the "Lower Flysch" and older formations. Thus, obduction of Poya Terrane is though to have occurred during the late Priabonian (ca. 33.7–35 Ma). In contrast, ultramafic minerals and clasts are absent in the Noumea–Bourail flysch (Bodorkos; Paul and Meffre). The obduction is therefore inferred to have occurred in two steps. The Poya Nappe was first emplaced during the late Priabonian and the Ophiolitic Nappe in turn was obducted after the late Priabonian pro parte and likely before the beginning of the Oligocene.

9.2. Piggy-back basin

According to Paris et al. (1979), the Tertiary obduction is post-dated by the Upper Eocene Nepoui Flysch (Fig. 4). Indeed, pelagic foraminifers of the Upper Bartonian Globigerinatheka semiivoluta zone (upper P14 to lower P16, 35–38 Ma) were found in limestone beds within the flysch. This has been confirmed recently by Chaproniere (in Meffre, 1995) who described an Upper Bartonian to Priabonian pelagic microfauna in a limestone olistolith found near the base of the sequence. This correlation poses a problem that cannot be solved without questioning the post-obduction character of the Nepoui flysch. A maximum late Priabonian age for the obduction is constrained by the parautochthonous "Formation de la cathedrale" (see above). Therefore, the obduction could be either post-late Priabonian (33.7–35 Ma) and/or pre-late Bartonian (35–38 Ma). This inconsistency infers that the post-obduction and autochthonous characters of the Nepoui flysch must be reconsidered.

The clastic and carbonate sediments that crop out in the Nepoui Peninsula (Fig. 13) can be divided into three main sequences (Coudray and Paris). In ascending order they are (Fig. 14): (1) The basal sequence rests on serpentinite or highly deformed basalt. Thin calcarenite or calcarenitic microconglomerate with reworked well-rounded serpentinite clasts is overlain by light brown to khaki argillite containing reworked (?) mid-Eocene radiolarians (Coudray, 1975). At its base, the argillite contains a few metre-scale foraminiferal calcarenite olistoliths derived from the basal limestone, and changes up-section into argillaceous arenite. (2) An alternation of argillite containing heulandite (zeolite), fine and coarse grained calcarenite, and carbonate rocks, is toped by an olistostrome reworking the underlying rocks. (3) The upper sequence is an alternation of arenite and dolomicrite lying disconformably upon the olistostrome. The whole succession is folded and faulted and covered unconformably by Miocene conglomerate (Fig. 13).

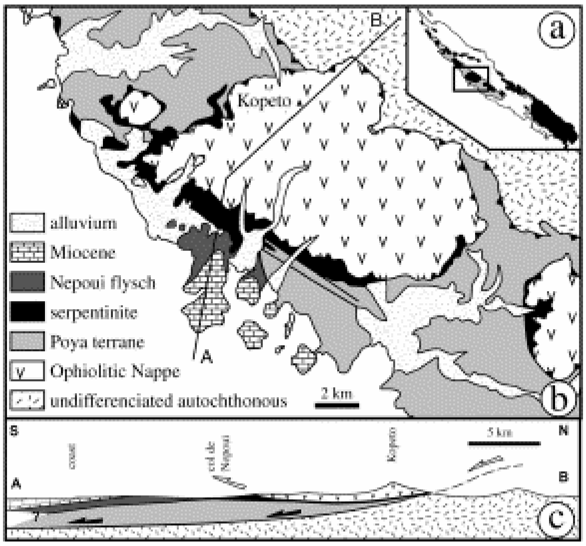

Fig. 13. The Bartonian Nepoui flysch; (a) location of the Nepoui area, (b) Geological sketch map of the Nepoui peninsula, (c) a cross-section of the Kopeto–Boulinda ultramafic klippe, Poya Terrane and Nepoui flysch showing the geometrical relationship between these terranes and the unconformably overlying Miocene conglomerate and limestone.

The Nepoui Flysch is composed of carbonate and mafic material; bioclastic carbonate rocks are indicative of a neritic environment (external platform) alternating with inner platform micrite and dolomicrite. Arenite, high-Ti argillite, and heulandite clearly indicate the alteration and erosion of a predominantly basaltic terrane (Coudray, 1975). Mafic clasts are always angular, precluding significant fluvial or coastal transport. Mid-Eocene radiolarians may have been reworked from associated bathial sediments. Ultramafic clasts are rare except in the basal calcarenite, limestone olistoliths, and at the top of the upper sequence were it includes chromite, and serpentinite clasts and their weathering products. Therefore, the Nepoui Flysch is mainly derived from basalt and minor altered serpentinite. Fresh peridotite clasts are notably absent. Thus, as serpentinite clasts may have been derived from serpentinite slivers of the Poya Terrane, the contribution of the Ophiolitic Nappe remains extremely modest if any. Scarce and small-sized ultramafic clasts within the Nepoui Flysch contrast sharply with the coarse peridotite conglomerate in the unconformably overlying Miocene.

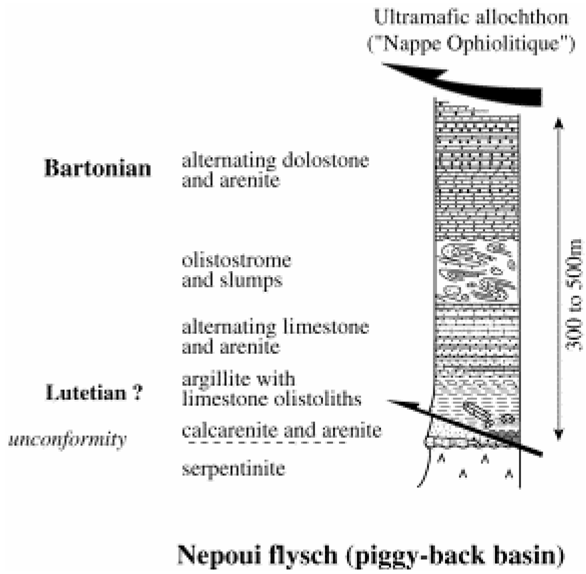

Fig. 14. Stratigraphic column for the Nepoui flysch according to Coudray (1975) to show the three sequences recording the late Eocene piggy-back evolution of the Loyalty fore-arc.

The overall features of the Nepoui Flysch are consistent with deposition in an unstable shallow to medium depth basin and derivation from a mafic terrane and subsidiary altered serpentinite, but do not fit its postulated post-obduction character. A coarser character and prominent derivation from peridotite might be expected. Although it clearly appears that the flysch locally rests unconformably upon serpentinite, it is not obvious that this serpentinite belongs to the tectonic sole of the Ophiolitic Nappe. Compared to the very simple structure of the latter, the Nepoui Flysch appears to be severely deformed; therefore, this tectonic event is unlikely to postdate the obduction. This and the inconsistency of micropaleontologic data suggest that the inferred post-obduction and autochthonous characters of the Nepoui Flysch are questionable. Instead, a model involving a transported ("piggy-back") basin may be proposed. Accordingly, the Nepoui Flysch may have accumulated in a fore-arc basin and thereafter have been transported en bloc upon the New Caledonian microcontinent during the final stage of obduction.

The argillite forming the base of the sequence could result from the erosion of basaltic laterite (Coudray, 1975) formed by the weathering of an already emerged unit; either in the fore-arc area, or in an accretionary complex. The limestone olistoliths that are found in the argillite (Fig. 14) may have been deposited originally in an outer shelf setting during a relatively quiet period that also resulted in the weathering of basalts and thereafter were reworked as olistoliths during the tectonic reactivation of the basin (Fig. 15c). The serpentinite clasts and chromite found in the limestone could have been derived from serpentinite bodies which may have been formed on the oceanic floor during accretion, i.e. serpentinite horsts formed during amagmatic seafloor spreading (Auzende et al., 1999) and/or transform faulting (Auzende et al., 1994). Alternatively, they may have been derived from serpentinite seamounts formed by stretching and mantle unroofing of the fore-arc lithosphere (the next Ophiolitic Nappe) (Taylor, 1992). The arenaceous flysch accumulated thereafter in response to an increasing substratum instability that culminated with olistostrome deposition (Fig. 14 and Fig. 15). The "in sequence" propagation of the duplex resulted in a relatively stable period in the flysch basin marked by the deposition of the uppermost arenite/dolomite sequence (Fig. 14). Finally, the progressive subduction of the margin of the New Caledonia block resulted in the underplating of the mafic duplex forming the Poya Nappe below an “out of sequence” thrust involving the fore-arc upper mantle forming the Ophiolitic Nappe (Fig. 15e). Then, the Nepoui Flysch basin was pushed in front of the Ophiolitic Nappe reaching its pre-Miocene location.

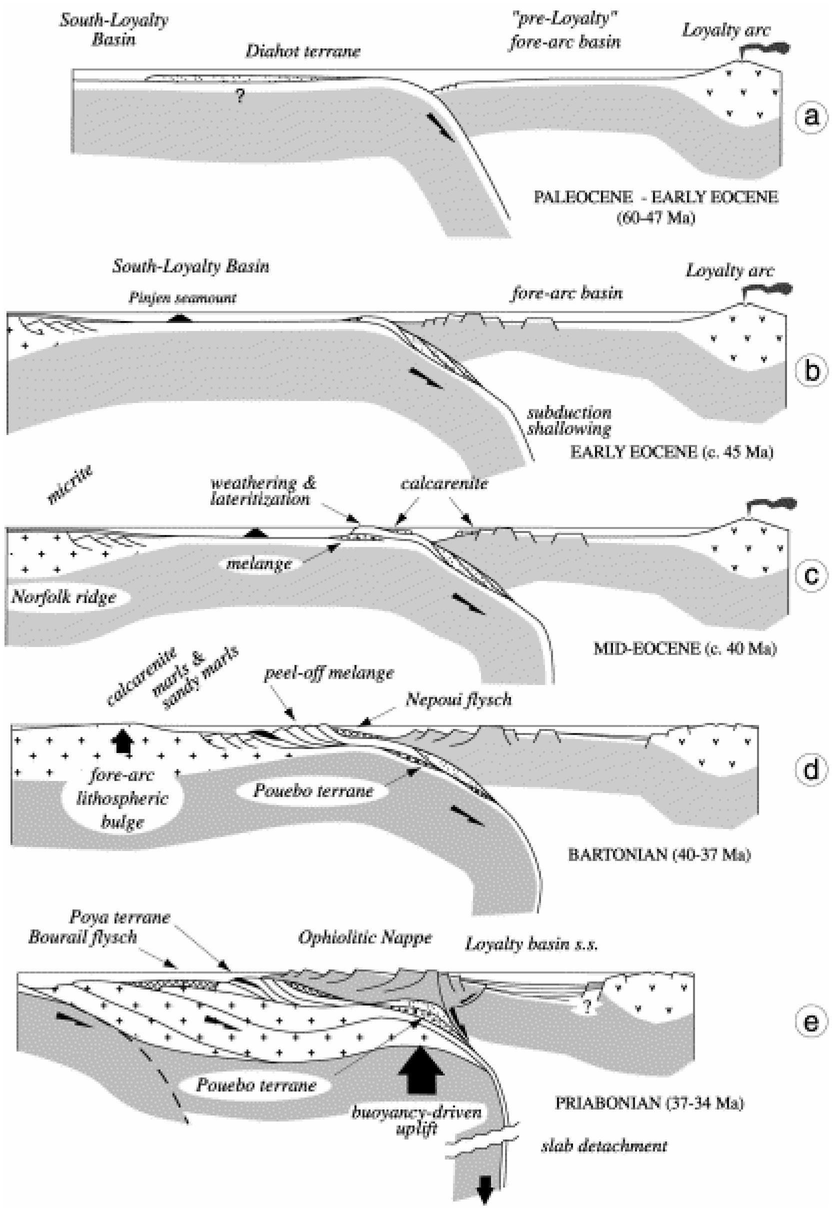

Fig. 15. A tectonic/geodynamic model for the evolution of the Eocene accretion/subduction complex of New Caledonia (not to scale). Owing to the oblique character of the subduction/obduction with respect to the trend of Norfolk Ridge, the tectonic features are subject to north-southward diachronous changes; therefore, the sections presented here are only intended to represent a synthetic view, not the actual structure in a precise location.

The post-obduction character that was previously suggested for the Nepoui Flysch is not consistent with its overall features that better fit with a model of transported or piggy-back basin. As, a pre-late Bartonian timing for the end of ophiolite obduction cannot be accepted, the timing of obduction is constrained stratigraphically by the parautochthonous “Formation de la Cathédrale” as post-late Priabonian (i.e. ca. 34 Ma).

10. Discussion and conclusion

10.1. Allochthonous character of the Poya Terrane

From the evidence presented here, we confirm the allochthonous character for the Poya Terrane for the following reasons: (a) siliceous argillite and hydrothermal/radiolarian red chert associated with the Poya Terrane are clearly related to the bathyal or abyssal zone. They are therefore in sharp contrast with underlying Cretaceous to Eocene shallow water sediments and were deposited in different remote places; (b) the Campanian to earliest Eocene age of the terrane given by the associated radiolarian-bearing sediments is older than that of the underlying rocks; (c) its metamorphic grade is appreciably higher; and (d) its tectonic structure is far more complicated than that of underlying terranes.

10.2. Geotectonic setting of the Poya Terrane

The dominant geochemical features of the Poya Terrane basalts (E-MORB) allow a model of plume-related magmatism to be suggested for the origin of the Poya Terrane. Such “enriched” basalts are commonly found in oceanic plateaus (i.e. Ontong Java Plateau). The regional geologic features of the terrane are, however, inconsistent with such an origin. The associated sediments are typical of a bathyal environment and with the exception of the late Paleocene micrite deposited upon alkaline seamounts, much of the sea bottom must have been below the CCD. Therefore, the basin in which Poya Terrane basalts erupted had a “normal” bathymetry of ca. 3500–4500 m. Nd isotopic variations suggest mixing of two mantle sources DM and EM II, and plume-enhanced back-arc marginal basin opening similar to that of the North Fiji Basin. A small amount of contamination by remnants of the stretched lower continental crust cannot be ruled our for some rocks belonging to the “depleted-MORB” end-member only. This could be consistent with an origin in a marginal basin setting, but it is to be confirmed by Pb isotope data (in progress). The rest of the terrane is formed of rocks free of any contamination by the continental crust.

In contrast to the interpretation forwarded by Eissen et al. (1998), the Poya Terrane cannot be considered as a coherent slice of oceanic crust picked up and pushed in front of the “Nappe Ophiolitique”. Instead, it appears to be formed of a number of hundreds of metres to kilometre-thick slices of upper oceanic crust which have been piled-up together at shallow crustal levels. Paleomagnetic evidence reveal that Poya Terrane basalts were generated at a low southern latitude, at ca. 200 km to the north of their present location (Ali and Aitchison, 2000). After dip correction, declination values display a dispersal of 140° or so, which is obviously due to local deformation. Such dispersal may be a result of folding around steeply dipping axes that regionally affect the Poya Terrane (Fig. 1). Strike correction however, does not result in a better declination fit. Thus local rotations are likely to have occurred before the late folding event, i.e. during tectonic scrap off and accretion in the forearc region. This and the absence of “continent-related” lithologies within the terrane, allow a model of intra-oceanic accretion to be forwarded for the secondary origin of the terrane. Sediment off-scrapping is a common feature of accretionnary prisms, and ophiolitic “coloured” melanges commonly occur in paleosubduction zones as disorganised sets of blocks of various size including mafic and ultramafic rocks. Accretion of kilometre-scale slices of coherent crustal material seems to be an unusual phenomenon in the present record. According to analog experiments for material transfer in accretionnary wedges (Gutscher et al., 1998), the peeling of the upper part of the down-going slab remains a possibility. This may result either from an extreme roughness of the down-going plate, or from a very shallow dip of the subduction zone, or both. Off-scrapping of the Poya Terrane in the Loyalty fore-arc probably took place because the Diahot Terrane had been previously subducted and underplated (Fig. 15a and b). Underplating of light crustal material at moderate depth (ca. 30 km) generally results in crustal thickening and uplift (Platt, 1993 and references herein). Underplating of the Diahot Terrane and incipient accretion of the Poya Terrane likely resulted first in thickening and subsequent uplift of the forearc crust (Fig. 15c). As a consequence, the upper part of the complex underwent tectonic thinning and local upper mantle unroofing. A temporary acceleration of the subsidence rate in the fore-arc basin due to tectonic thinning was marked by the formation of the Nepoui Flysch basin which is partly floored by serpentinite. A changing subduction regime has been recorded in the fore-arc region by facies changes within the Nepoui Flysch, basin instability and olistostrome development.

The mafic rocks that form the Poya Terrane have been scrapped off the ocean floor of a basin which was originally opened during the Campanian–Paleocene to the east of the Norfolk Ridge as a consequence of migration of the east-Gondwana margin (Fig. 5) towards the Pacific. It is notable that this basin opened synchronously with the Tasman Sea and that opening was closely followed by the end of volcanic-arc activity in New Caledonian autochthonous terranes. During the late Eocene convergence, this basin almost disappeared because of eastward subduction below the Loyalty Arc. Remnants of this basin are now likely to be located south of the Loyalty Islands, between the Loyalty and Norfolk ridges; therefore, we call this basin the South Loyalty Basin (Fig. 5, Fig. 15 and Fig. 16).

10.3. Correlation between Pouebo and Poya terranes

Although a precise stratigraphic correlation with the Pouebo Terrane remains impossible, it appears that its content is at least in part very similar to Diahot and Poya terranes rocks, especially the latter. Poya and Pouebo terranes contain mafic rocks that not only display the same geochemical features but also the same diversity. They are therefore likely to be genetically related.

10.4. Geotectonic setting of the Pouebo Terrane

The Pouebo Terrane is not simply a metamorphosed lateral equivalent of the Poya Terrane. Indeed, the diversity of rocks found in this terrane reveals a mixed origin from both “oceanic” and “continental” terranes. Metamorphosed kilometre-scale mafic slices and angular blocks, serpentinite slivers and other ocean-related lithologies may result from the tectonic erosion and subsequent underplating at ca. 30–60 km depths of fragments of the down-going South Loyalty slab that were already accreted to the fore-arc region. The metamorphosed coarse conglomerate composed of well rounded mafic clasts found at several places in the Pam Peninsula and along the east coast, likely represent remnants of cliff conglomerate that developed during periods of emersion (i.e. growing) of the accretionary complex and were subsequently underplated during periods of tectonic erosion (Fig. 15c and d). The serpentinite slivers may have originated either from the upper mantle of the overriding plate by tectonic erosion and down-scrapping, or alternatively, from serpentinite bodies that belong to the down-going plate and originated during the basin formation through amagmatic spreading and unroofing of the upper mantle (Auzende et al., 1999).

Due to the high metamorphic grade and multistage deformation, a precise identification of “continent-related” lithologies is not always possible. In a few cases, derivation from the Diahot Terrane is clear. This is sound evidence for the role the latter terrane played in the tectonic evolution of the area by entering the Loyalty subduction zone first. As this terrane includes deep-sea fan terrigenous sediments associated with volcanic-arc rocks that were accumulated upon an extremely thinned continental crust, or even directly upon an older oceanic crust (see regional setting above) it could have easily enter the subduction zone and been underplated. Some fragments may have been dragged to greater depths together with mafic material to form the Pouebo melange.

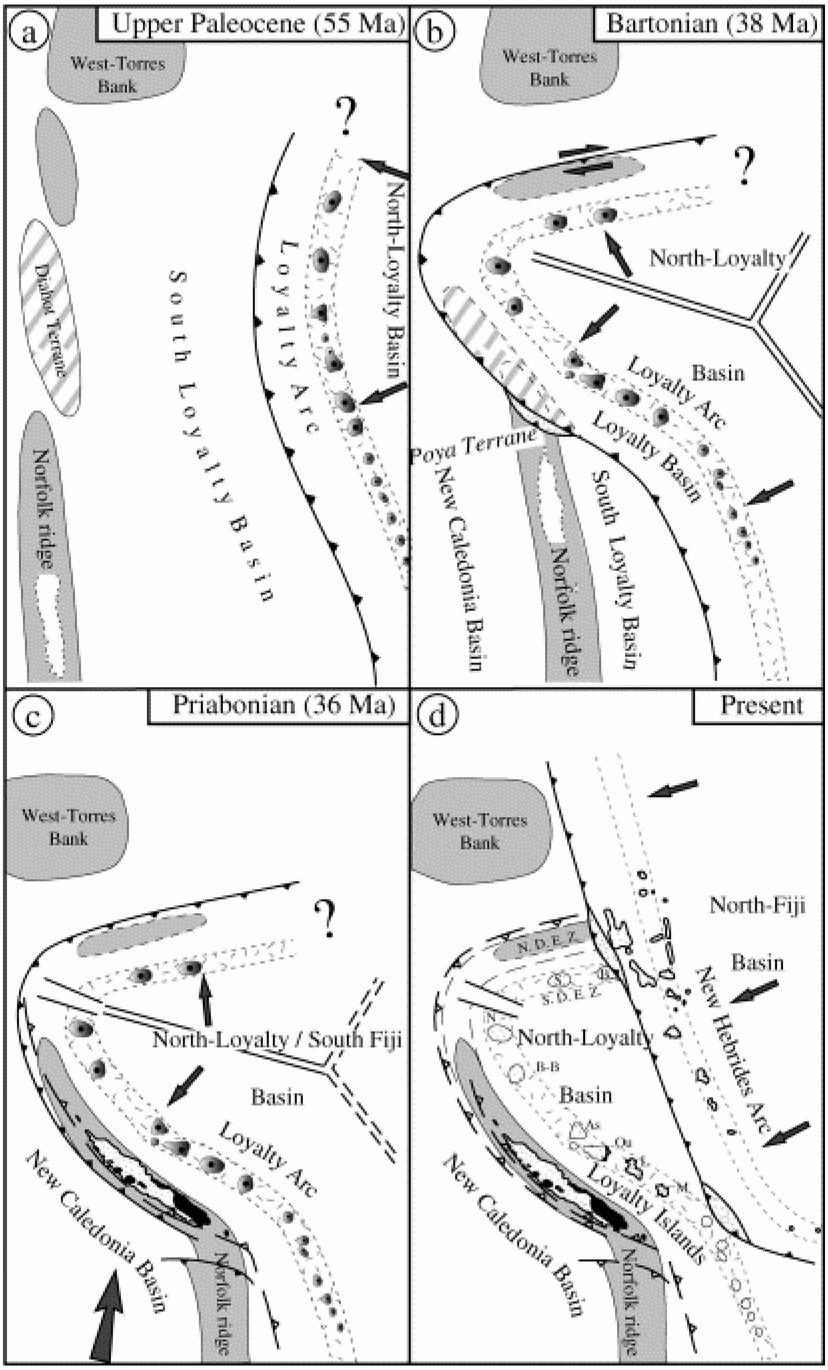

Fig. 16. Reconstruction of the Upper Eocene oblique convergence of the Norfolk Ridge and Loyalty arc. (a) In the initial stage of the east dipping subduction, the Loyalty arc had probably a more or less rectilinear shape. (b) A complex opening mechanism of the North Loyalty basin resulted in an arched shape of the arc, and in the diachronous subduction of various crustal slices derived from the east-Australian margin. (c) After blocking of the subduction, convergence continued shortly along the southwest coast of New Caledonia, roughly in a north–south direction, due to the east–west trend of the active ridge. In this model, the parallel trends of the Norfolk

and Loyalty ridges north of 23°S are directly related to the Eocene obduction/collision. (d) In the present situation, the North Loyalty basin as an integral part of the Australian plate is being subducted beneath the New Hebrides arc and incipient collision appears to the east of the d'Entrecasteaux Zone, and to the southeast of Mare Island. N.D.E.Z.: North d'Entrecasteaux Zone; S.D.E.Z.: South d'Entrecasteaux Zone; B: Bougainville eamount; S: Sabine bank; N: Noroit reefs; B-B: Beautemps and Beaupre reefs; As: Astrolabe reefs; Ou: Ouvea island; L: Lifou island; M: Mare island.

10.5. A tectonic/geodynamic model

Two main periods may be distinguished within the late Cretaceous to Eocene tectonic evolution of this area of the Southwest Pacific: (1) a Campanian to late Paleocene accretion or marginal basin opening period associated with the eastward migration of the Pacific margin; and (2) the marginal basin accretion was followed by a period of “convergence” and basin closure during the Eocene, due to the subduction flip and continent-ward migration of the Loyalty Arc (Fig. 5). During the former period, the Pacific margin migrated eastward and except in a few places, the possible remnants of the late Cretaceous arc/back-arc system are now largely hidden below sea level or beneath younger volcanic-arc rocks. Indeed, Late Cretaceous arc-related volcanic rocks are found in New Caledonia (Picard; Black; Gros and Debeuf), and Campanian volcanic-arc and back-arc basalts are known to exist in the Northland of New Zealand where they were obducted in the Oligocene (Nicholson and Nicholson). In addition, similar rocks form parts of the East Cape Allochthon (Matakaoa Volcanics, northeast New Zealand, in progress). It is worth noting that the resulting lithospheric boudinage was almost simultaneous over the entire area and therefore cannot be considered as a direct consequence of oceanward arc migration. The opening of marginal basins floored with “enriched” tholeiite of relatively deep mantle origin is consistent with a model of rapid stretching of the east-Gondwana continental margin.

During the Eocene, various slices of continental and oceanic material were accreted to the Loyalty fore-arc, until the subduction was blocked in the latest Eocene by the tip of the Norfolk Ridge. If we consider that the opening of Tasman Sea and New Caledonia basins was completed in the late Paleocene then, the driving mechanism of the accretion/obduction process could be either the northward drift of the Australian Plate due to the opening of the Southern Ocean, or alternatively the westward migration of the Loyalty Arc, or both. The arcuate shape of the northern part of the Loyalty Ridge is very similar to that of the New Hebrides Arc and may have similarly originated through a very complex back-arc opening process (Auzende et al., 1995). We propose that during a first step, the westward migrating and anticlockwise rotating Loyalty Arc subduction zone diachronously absorbed several ribbons of thinned continental crust that were previously rifted from Australia, including the Diahot Terrane which may be tentatively correlated to the Norfolk Ridge s.l. (Fig. 15a). The oblique motion of the arc with respect to the rifted fragments prevented them to blocking the subduction. At that step, a ENE–WSW directed convergence is consistent with the ductile strain features (i.e. regional stretching lineation, sheath folds, etc) found in the Diahot Terrane (Cluzel et al., 1995). Once the Diahot Terrane underplated in the Loyalty fore-arc at ca. 45–43 Ma, the subduction regime dramatically changed, either by subduction shallowing or by an increasing role of the buttress formed by the over-riding plate, and masses of mafic material were scrapped off the down-going plate and accreted to the fore-arc as it was emerging (Figs. 15a,c and 16b). Consequently, the Nepoui piggy-back basin developed in a fore-arc setting (Figs. 15c,d and 17). When the Norfolk Ridge s.s. entered the subduction zone at ca. 38 Ma, its thinned eastern margin first met the mafic accretionary complex. The Bourail foreland flysch then accumulated upon the Norfolk Ridge with an increasing amount of detrital mafic material derived from the approaching Poya Terrane (Fig. 15e). Subduction of a thicker part of the ridge provoked the uplift and unroofing of the fore-arc region now represented by the “Nappe Ophiolitique”. The prograde hydratation of the Pouebo Terrane eclogite may be a consequence of fluid input from the underplated terranes before the whole set was quickly uplifted and exhumed (Fig. 15e). Once the Loyalty subduction system was blocked by the Norfolk Ridge, a limited period of N–S directed convergence resulted in the sinistral transpression that affected the New Caledonia island as a whole and its offshore margins (Fig. 16c). The subduction then jumped diachronously back to the western margin of the Norfolk Ridge, and the New Caledonia Basin started subducting below the ridge (Fig. 15 and Fig. 17). Evidence for a limited period of subduction is provided by the development of a new sediment-dominated accretionary complex, the occurrence of an abandoned slab beneath New Caledonia (Regnier, 1988), and a short-lived active margin magmatic episode represented by mid-Oligocene (27 Ma) granodiorite intrusives (in progress). The arrival of the thinned margin of the Lord Howe Ridge during the latter subduction resulted in the overall compressional deformation that affected the entire area including the Lord Howe Ridge and the Loyalty Ridge itself (Symonds; Symonds and Auzende), and finally blocked the system before complete destruction of the New Caledonia Basin.

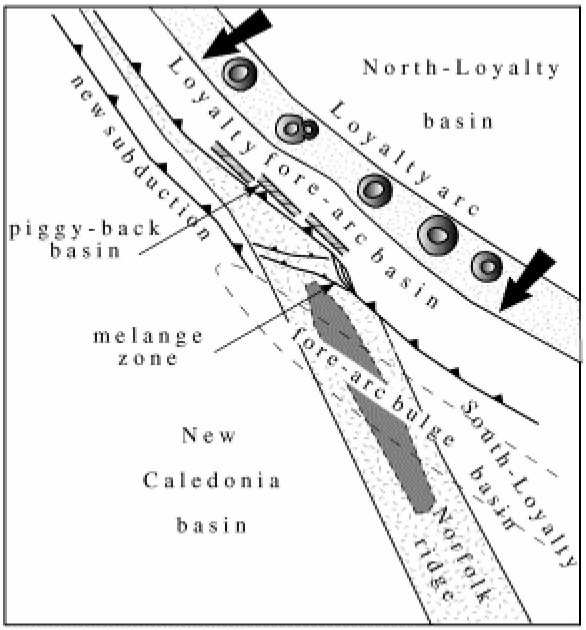

Fig. 17. Sketch map of the Norfolk Ridge–Loyalty arc system in the Bartonian (ca. 38 Ma). The black arrows represent the westward migration of the Loyalty arc relative to the Australian plate. Major features are the fore-arc bulge responsible for the pre-Priabonian unconformity, location of pyggy-back basins (Nepoui flysch) in the fore-arc, and the jump-back of the trench due to subduction blocking.

Our model for the obduction in New Caledonia illustrates the role of microplates (or isolated continental slices) in the continentward obduction of large ultramafic bodies that is still a controversial issue. It is notable that the fore-arc accretion closely followed by underplating and finally exhumation of New Caledonian terranes resulted in the close juxtaposition of unmetamorphosed and highly metamorphosed mafic/ultramafic terranes in a single narrow belt (less than 50 km) which very closely resembles that found in association with volcanic-arc terranes within many collisional orogens of Europe and Central Asia.

Considering the present evolution of the SW Pacific, a rapid convergence rate (ca. 20 cm/year) in the New Hebrides subduction zone will lead to a new collision between the Loyalty and New Hebrides arcs (Fig. 16d) and shortly thereafter between the Norfolk Ridge and the already collided arcs. Further convergence could result in the progressive amalgamation of SW Pacific terranes into a new orogenic belt along the east coast of Australia. In such an orogenic belt, the "Norfolk Terrane" composed of New Caledonia and New Zealand pro parte, will resemble the "exotic" terranes with prominent biologic endemicity that are found in cordilleran orogens.

Therefore, the Late Cretaceous to Recent evolution of the Southwest Pacific may be taken as an example of the pre-collisional events that are often erased during the final stages of evolution in large collisional orogens.

Acknowledgements

The analytical work has been supported by the French Ministry of Education through a special grant to D.C. during his stay in New Caledonia, the University of New Caledonia, and the University of Grenoble 1 (Laboratoire de Géodynamique des Chaînes Alpines, UMR 5025). Special thanks to Pr. P.M. Black for providing samples of the University of Auckland collection, and to the French Embassy in New Zealand for supporting French–New Zealand co-operation in Earth Sciences. A. Nicolas and J.-P. Eissen are thanked for throughout review and improving comments on the paper.

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