Geology of the New Caledonia region and its implications for the study of the New Caledonian biodiversity

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Introduction

The New Caledonian exclusive economic zone is located in the Southwest Pacific between Australia and the Vanuatu archipelago (formerly New Hebrides). It extends over 1200 km from north to south and 1800 km from west to east, from the Chesterfield Islands up to the Matthew and Hunter Islands at the southern tip of the Vanuatu archipelago (Figure 1). The main islands of New Caledonia are located on two parallel NW-SE trending ridges. The largest island (the 400 km-long and 50 km-wide Grande Terre) and subordinate islands including the Belep islands in the north and the Isle of Pines in the south are supported by the New Caledonia Ridge. The Loyalty Islands are supported by the Loyalty Ridge.

Geologically, the area is schematically composed of a series of NW-SE trending ridges and basins formed by i) stretching and spreading along the eastern margin of Australia during Late Cretaceous-Paleocene times and ii) Eocene convergence which was responsible for the emplacement of one of the largest ophiolitic complexes in the world: the New Caledonian ophiolitic nappe (a slice of oceanic lithospheric mantle) from which weathering formed one of the world's largest reservoirs of nickel. The ridges and basins are nowadays supported by the Australia plate that subducts beneath the Vanuatu active volcanic arc. The southernmost segment of the arc supports the active volcanic islands of Matthew and Hunter. These latter islands are thus on a plate (a micro-plate) that differs from the one supporting the main islands of New Caledonia territory.

After reviewing the different morphostructural units and discussing the geological history of the New Caledonia area, I will point out some facts that may be relevant in the study and understanding of New Caledonia biodiversity.

Geological of the morphostructural units of the New Caledonia region

From west to east, the main geological units between Australia and Vanuatu include the Tasman Sea Basin, the Lord How Rise with subordinate basins, ridges and chains, the New Caledonia Basin, the New Caledonia Ridge, the South Loyalty Basin, the Loyalty Ridge, the North Loyalty Basin and the Vanuatu Ridge (Missègue *et al.*, 1991; Smith and Sandwell, 1997; ZoNéCo 1998) (Fig. 1 plate 1/1 and Fig. 2 plate 2/1).

The Tasman Sea Basin

To the southeast of Australia the Tasman Sea Basin is a wide (up to 2000 km) and deep (4000 m) basin floored with oceanic crust formed by spreading from the Cretaceous (85 Ma: Santonian to Early Campanian) to Earliest Eocene (52 Ma) (Hayes et Ringis, 1973; Weissel and Hayes 1977; Gaina *et al.*, 1998). In the central part of the basin, the N-S trending Tasmantid volcanic chain is an Oligocene to Late Miocene hot spot chain showing the northward drift of the Australia plate (Vogt and Conolly, 1971; Mc Dougall and Duncan, 1988).

The Lord Howe Rise

The Lord Howe Rise is a main bathymetric feature which averages 400 km in width and extends over 1600 km from the Challenger Plateau off New Zealand to the Chesterfied area. Water depths on the crest are 1200 to 750 m. In the north, at the latitude of New Caledonia, the Lord Howe Rise includes several basins, ridges and chains. They are from west to east: the Dampier Ridge, the Middleton basin, the Chesterfiel/Bellona Plateau, the Faust basin, the Lord Howe Rise crest, the Fairway Basin and the Fairway Ridge. The Lord Howe Rise is interpreted as a thinned fragment of continental crust split from Gondwana by Cretaceous rifting and subsequent spreading in Tasman Sea Basin, and it is probably composed of Paleozoic basement overlain by rift basins with up to 4 km of Mesozoic and Cenozoic sediments (Willcox *et al.*, 2001; Van de Beuque *et al.*, 2003; Exon *et al.*, 2004). Stratigraphy also indicates an emergence and major unconformity from the Late Eocene to Early Oligocene (Burns *et al.*, 1973), which correlates with described compressive structures of that age (Lafoy *et al.*, 1994; Symonds *et al.*, 1999; Auzende *et al.*, 2000).

The Lord Howe rise has Cenozoic seamount chains as exemplified by N-S trending volcanic edifices on its crest (Van de Beuque *et al.*, 1998; Exon *et al.*, 2004) and the Oligocene-Miocene Lord Howe seamount chain on its western side. The Chesterfield/Bellona plateau is supported by five guyots that constitute the northern and oldest volcanoes (Late Oligocene ?) along the Lord Howe hotpsot chain (Missegue et Collot, 1987).

The 800 km long, 130 km wide and NW-SE to NNW-SSE trending and southward deepening (1000 to 3000 m) Fairway Basin, firstly interpreted as oceanic in nature (Ravenne *et al.*, 1977; Mignot, 1984; Eade, 1988; Uruski and Wood, 1991; Van de Beuque, 1999), is now considered to be floored by stretched-thinned continental crust based on the presence of salt diapirs derived from Cretaceous series (Auzende *et al.*, 2000), gravity modeling (Vially *et al.*, 2003) and extensional-type horst and graben structures of the crust (Lafoy *et al.*, 2005). It is proposed that the formation of the Fairway Basin took place during the Late Cretaceous (95-65 Ma) by continental stretching, at the same time as the stretching in the Middleton Basin (Lafoy *et al.*, 2005).

The 600 km-long and NW-SE trending Fairway Ridge, culminating in its northernmost part at the Lansdowne bank, thins and deepens southward. The origin of the ridge is still controversial. Previously interpreted as a ridge of oceanic nature (Ravenne *et al.*, 1977; Mignot, 1984) and as an oceanic piece of the New Caledonia basin crust overthrust along the Lord How Rise (Lafoy *et al.*, 1994; van de Beuque, 1999; Auzende *et al.*, 2000), it is now considered as thinned continental crust (Vially *et al.* 2003; Lafoy *et al*, 2005).

The New Caledonia Basin

The New Caledonia Basin extends from west of Northern New Zealand to west of New Caledonia, parallel to the Lord Howe Rise. The deepest (3600-3700 m) northern part (north of 22°30'S) strikes NW-SE while the central part (3000 m deep) strikes NNW-SSE. Origin of the basin is controversial from oceanic type (Shor *et al.*, 1971; Dubois *et al.*, 1974; Weissel et Hayes, 1977; Willcox *et al.*, 1980; Kroenke, 1984; Mignot, 1984 ; Sutherland, 1999 ; Auzende *et al.*, 2000) to thinned continental type (Etheridge *et al.*, 1989 ; Uruski et Wood, 1991 ; Sdrolias *et al.*, 2003 ; Vially *et al.*, 2003 ; Lafoy *et al.*, 2005). The horst and graben structure of the crust with westward tilted blocks and a sedimentary section up to 8 km thick suggests a thinned continental crust for the northern NW-SE segment. On this segment the crust dips to the east, toward the western margin of the New Caledonia ridge. At the base of this margin, buried deformation features interpreted as the result of compression have been recognized (Rigolot and Pelletier, 1988). In contrast to the northern segment, the central segment of the basin, with magnetic lineations and an axial ridge buried by a 4 km-thick sedimentary sequence, is interpreted as a segment floored with oceanic crust that formed during the Paleocene (possibly from 62 to 56 Ma) after Late Cretaceous–Earliest Paleocene stretching (Lafoy *et al.*, 2005).

The New Caledonia Ridge

The New Caledonia Ridge is the NW-SE trending northern segment of the 70-100 km-wide Norfolk Ridge which extends over 1500 km from the d'Entrecasteaux Reef to the northern tip of New Zealand. As for the Lord How Rise, the Norfolk ridge is interpreted to be a continental ribbon detached from Gondwana. Geology of the Grande Terre of New Caledonia (Lillie and Brothers, 1970; Paris, 1981; Picard, 1999) has great similarities with that of New Zealand. Main geologic features of the island have been tentatively extended along the submerged northward (Collot *et al.*, 1988) and southward (Rigolot, 1988) segments of the ridge.

The island is composed of a series of various terranes assembled during two tectonic events : a Late Jurassic to Early Cretaceous tectonic collage (Paris, 1981 ; Meffre 1995 ; Aitchison *et al.* 1998) and a Late Eocene subduction/collision resulting in the emplacement at the Latest Eocene (38-34 Ma) of a large ophiolitic nappe (Avias 1967; Paris, 1981 ; Collot *et al.*, 1987 ; Aitchison *et al.*, 1995 ; Cluzel *et al.*, 1994, 2001).

The pre-Cretaceous terranes, mainly located in the central chain, are unconformably overlain by Upper Cenomanian (Late Cretaceous) to Upper Eocene sediments, and include disrupted Late Carboniferous ophiolite (Meffre *et al.*, 1996 ; Aitchison *et al.*, 1998), mid-Triassic to late Jurassic volcano-sedimentary arc terrane (Meffre *et al.*, 1996), Mid Permien to Late Jurassic volcano-sedimentary arc terrane (Campbell *et al.*, 1985), post-Liassic unit composed of oceanic crust and volcano-sedimentary distal deposits (Cluzel, 1996) and affected by a Late Jurassic high pressure metamorphism (150 Ma : Blake *et al.*, 1977).

The post-Early Cretaceous terranes also involved in the Late Eocene major event include:

- an unmetamorphosed Upper Cretaceous to Upper Eocene sedimentary pile well exposed along the western side of the Grande Terre. This pile is composed of two sequences separated by an unconformity (Paris, 1981; Cluzel *et al.*, 2001). The lower sequence includes an Upper Cretaceous fining-upward clastic series of conglomerates, sandstones, coaly siltstones and volcanic rocks, overlain by Paleocene to Middle Eocene pelagic limestones and cherts, indicative of a deepening of the deposit environment. The upper sequence is a Upper Eocene (Upper Bartonian to Priabonian) flysch formation (Nouméa-Bourail and Népoui flyschs), deposited after a short period of deformation and erosion and showing a coarsening upward sequence with (a) basal neritic limestones, (b) a «lower flysch» member of marls and calcareous sandy marls, (c) an «upper flysch» member composed of fine-grained calcareous turbidites interbedded with mafic breccias, and (d) a "wildflysch" member that incorporates blocks and olistoliths of siliceous shales, limestones, basalts and flysch, topped by an olistostrome recording the Late Eocene tectonic paroxism.
- a mafic unit of oceanic basalts interbedded with argillite and cherts (named basalt nappe or Poya unit) of Late Cretaceous (Campanian) to Late Paleocene-Earliest Eocene age (85-55 Ma), with back-arc or fore-arc affinities (Routhier, 1953; Espirat, 1963; Eissen *et al.*, 1998; Cluzel *et al.*, 1997, 2001). This unit, severely sheared and folded, is mainly exposed along the northern half part of the west coast but also outcrops along the east coast; it always underlies the ultramafic nappe and tectonically overlies the Upper Eocene sedimentary rocks.
- mafic high pressure-low temperature metamorphic units (Pouebo and Diahot units) located in the northeastern part of the Grande Terre (Brothers, 1974; Paris, 1981; Yokoyama *et al.*, 1986; Maurizot *et al.*, 1989; Black *et al.*, 1993; Clarke *et al.*, 1997; Cluzel *et al.*, 1995; Baldwyn *et al*, 1999; Carson *et al.*, 1999; Rawling and Lister, 2002; Fitzherbert *et al.*, 2004, Spandler *et al.*, 2005). The units and especially the Pouebo unit are interpreted to be the equivalent of the basalt nappe metamorphosed under blueschist to eclogitic facies conditions, the latter (20 kbar, 650°C) indicating an underthrusting equivalent to a depth of 60-70 km. Radiometric dating of the metamorphism peak is 44 Ma (Spandler *et al.*, 2005) while those from the cooling ages range from 40 to 34 Ma (Baldwyn *et al.*, 1999), indicating a rapid unroofing and exhumation of the metamorphic units in the north and a synchronism with the final emplacement of the ophiolitic nappe (38-34 Ma) in the west and south of the island.
- the ophiolitic nappe mainly composed of peridotites and well exposed in the southern part of the island (3 km thick in the southern massif), Belep and Pines islands and as a series of klippes along the western northern half of the island (Avias, 1967; Guillon et Routhier 1971; Guillon, 1975; Prinzoffer *et al.*, 1980). If the age of its emplacement is Late Eocene (Paris *et al.*, 1979), the age of these mantellic rocks is not accurate and considered to be Late Cretaceous or older from radiometric datings of associated mafic and felsic dikes yielded Late Cretaceous (100-80 Ma) and Eocene (42-52 Ma) ages (Paris, 1981; Prinzhoffer, 1981). The peridotites are mainly harburgites and are the mother rocks for the nickel of New Caledonia.

The Oligocene is characterised by a lack of marine sediments, and post-obduction granodiorite intrusions (St Louis and Koum) radiometrically dated from 32 to 24 Ma (Guillon, 1975) and interpreted as the result of short-lived convergence episode along the western margin of the New Caledonia ridge (Cluzel *et al.*, 2005). The peridotites have been extensively weathered under aerial conditions since Oligocene, leading to the development of thick Ni-rich lateritic mantles (Trescases, 1973, 1975; Latham, 1986) and relict lateritic landsurfaces (Chevillotte *et al.*, 2006). Brittle extensional deformation plays an important role in the post-obduction morphotectonic evolution of the island (Leguere, 1976; Lagabrielle *et al*, 2005; Chardon and Chevillotte, 2006). This extension which may initiated in the Oligocene is expressed in the Neogene by the disruption of land surface formed during Oligocene planation and in the outcrop at Nepoui of Lower-Middle Miocene fluvial conglomerates with shallow water marine limestones (Coudray, 1976). Neogene ridge-normal then ridge-parallel to oblique extensional tectonics are also responsible for the shape and subsidence of the New Caledonia ridge margins (Daniel *et al.*, 1976; Dugas and Debenay, 1978; Bitoun and Récy, 1982; Rigolot, 1989; Chardon and Chevillotte, 2005; Flamand 2006). The Grande Terre displays one of the largest barrier-reefs in the world, isolating a locally-wide lagoon. The barrier reef settled in the Early Pleistocene (Coudray, 1976). Vertical motions from the 125 Ka reef indicate different tectonic blocks and a general slow subsidence (0.03 to 0.16 mm/year) of the coast except in the southeastern part of the Grande Terre and Pines island where coasts are uplifted (Launay and Récy, 1972; Launay, 1985; Cabioch, 1988, Cabioch *et al.*, 1996) and where quaternary faults have been observed (Lafoy *et al.*, 2000; Lagabrielle *et al.*, 2005 ; Flamand, 2006) and seismicity occurs (Régnier *et al.*, 1999; Pillet and Pelletier, 2004).

The South Loyalty Basin

Parallel to the Norfolk/New Caledonia Ridge, the South Loyalty Basin (called also the West Loyalty Basin) is a 1300 km-long, narrow (45-65 km wide), and northward deepening (from 2000 to 3800 m) basin with oceanic crust dipping northwestward and filled with thick (up to 8 km) sediments which are considered to be mainly post Eocene in age (Bitoun and Recy, 1982 ; Pontoise *et al.*, 1982 ; Collot, *et al.*, 1987). Geophysical data suggests that the oceanic basement is the continuity of the ophiolitic nappe of the Grande Terre. The age of the crust is unknown but considered as pre Late Cretaceous (Collot *et al.*, 1987) or Late Cretaceous to Paleocene (Cluzel *et al.*, 2001).

The Loyalty Ridge

The loyalty ridge is a narrow ridge parallel to the South Loyalty Basin and Norfolk Ridge and more or less continuous from the Cook Fracture zone in the South to the d'Entrecasteaux zone in the north. It is composed of a series of seamounts and guyots and supports the Loyalty islands.

The geology of the ridge is poorly known and its origin and nature are unknown. Parallelism with other ridges bordering the Australian margin suggests an old and continental origin (Monzier, 1993). However, taking into account its possible link with the d'Entrecasteaux zone (an Eocene subduction zone) and for convenience in the understanding the geology of New Caledonia, the Loyalty Ridge is considered as an Eocene island arc in most of the reconstructions, (Maillet *et al.*, 1983; Kroenke, 1984; Eissen *et al.*, 1998 ; Cluzel *et al.*, 1994, 2001 ; Crawford *et al.*, 2003 ; Sdrolias *et al.*, 2003 ; Schellart *et al.*, 2006). Middle to Upper Oligocene non orogenic volcanism has been also proposed for the origin of the ridge (Rigolot, 1989; Monzier, 1993). Only few volcanic rocks have been recovered in two areas. Upper Miocene (9-11 Ma) alkalic basalts (Baubron *et al.*, 1976) outcrop on Mare island. Submersible dives off Mare along the eastern flank of the ridge (Monzier *et al.*, 1989) recovered volcanic breccias, Middle Oligocene (32 Ma) alkaline rhyolites, Middle Oligocene tuffaceous sandstones, Middle Upper Oligocene chalks, Upper Oligocene (27 Ma) alkalic basalts, Lower Miocene (20 Ma) back-arc basalts, and algae and reefal limestones with reworked Eocene-Oligocene and Mio-Pliocene fauna (Monzier, 1993). The ridge appears to be composed, at least partly, of non orogenic alkaline volcanics of Middle to Late Oligocene and Late Miocene age, the youngest being likely a part of a N-S trending hot spot track (Rigolot *et al.*, 1988). None of the recovered rocks argues for a volcanic arc origin, although this hypothesis, possible and attractive, is widely accepted in the literature.

The Loyalty islands are mainly composed of Late Miocene to Pleistocene uplifted reef formations covering the basement (Chevalier, 1968 ; Marshall et Launay, 1978 ; Bourrouilh, 1996; Carrière, 1987; Guyomard *et al.*, 1996). The varying altitude of the islands shows the bulge of the Australia plate in front of its subduction eastward beneath the Vanuatu arc (Dubois *et al.*, 1974, 1977, 1988). The islands diachronously emerged during the Pleistocene (possibly in the Latest Pliocene for Mare) and are still emerging and uplifting (as exemplified by Ouvea) except the ones which have passed the top of the bulge and are thus subsiding.

The North Loyalty Basin

The North Loyalty Basin (named also East Loyalty Basin) is a deep (3000 to 5000 m) basin floored with oceanic crust dated of pre Middle Eocene age in its northernmost part (Andrews *et al.*, 1975). Bounded northward by the d'Entrecasteaux zone, it is the remaining part of a larger basin which disappears eastward in the active Vanuatu subduction zone. Initially interpreted to have formed by spreading in Early Eocene and regarded as the old part of the South Fiji Basin (Lapouille, 1982; Weissel *et al.*, 1982), it is now considered to have formed in Late Eocene (44 to 35 Ma: Sdrolias *et al.*, 2003) as a back arc basin of the Loyalty arc (Maillet *et al.*, 1983; Cluzel *et al.*, 2001; Schellart *et al.*, 2006).

The Vanuatu Trench and Ridge, the North Fiji Basin and the Vitiaz Trench Lineament

The Vanuatu Ridge is a 1500 km-long active volcanic arc related to the subduction of the Australia plate since the Late Miocene (12-10 Ma). Oldest known arc volcanic rocks from the Vanuatu ridge are however Early Miocene in age (Mitchell and warden, 1971 ; Carney and MacFarlane, 1982) and related to the fossil west-dipping Vitiaz subduction zone along which Pacific plate subducted, Late Miocene subduction reversal and initiation of east-dipping Vanuatu subduction being due to collision of the Ontong Java Plateau and the Melanesian border plateau (Packham, 1973; Kroenke, 1984; Brocher, 1985; Pelletier and Auzende, 1996). The Vanuatu arc rotated clockwise leading to the formation of the active north Fiji back arc complex basin (Chase *et al.*, 1971 ; Falvey, 1975 ; Auzende *et al.*, 1988, 1995).

The Vanuatu Trench is a segment of the present-day Australia-Pacific converging plate boundary along which Australia plate dips eastward and is consumed. Relative motion of convergence at trench is ENE-WSW and rate of motion varies along the trench and is about 12 cm/year at the latitude of the Loyalty islands (Dubois *et al*, 1977; Louat et Pelletier 1989 ; Pelletier *et al*, 1998 ; Calmant *et al*, 1995, 2003). Near 22°S, the Loyalty Ridge enters the trench and subducts/collides with the Vanuatu arc since 300 ka (Monzier *et al.*, 1989) , reducing the convergence motion south of the impact point and forming sinistral E-W strike-slip motion across the arc and isolating a micro plate (Louat and Pelletier 1989 ; Calmant *et al.*, 2003) on which Mattew and Hunter islands are active volcanoes (Maillet *et al.*, 1986). This incipient collision is also supposed to have tectonic effects on the Loyalty Ridge (Lafoy *et al.*, 1996).

Tectonic evolution of the New Caledonia domain

Numerous Cretaceous to Cenozoic reconstructions of the Southwest Pacific and New Caledonia region have been proposed in the past years (Kroenke 1984 ; Yan and Kroenke, 1993; Veevers, 2000 ; Muller *et al.*, 2000 ; Cluzel *et al.*, 2001 ; Sutherland *et al.*, 2001 ; Hall, 2002 ; Crawford *et al.*, 2003 ; Sdrolias *et al.*, 2003 ; Schellart *et al.*, 2006). Tectonic evolution concerning the New Caledonia domain can be divided into 5 stages (Fig. 3, plate 2/2).

The pre-Late Cretaceous period

The Paleozoic to Early Cretaceous period is thought to be marked by a subduction zone along the eastern margin of Gondwana. It ended with a Late Jurassic to Early Cretaceous tectonic orogeny (correlated with the Rangitata orogeny in New Zealand) that resulted in a collage of different units. The various Late Carboniferous to Late Jurassic units found in New Caledonia assembled at that time to form the old core of New Caledonia (Paris, 1981 ; Meffre, 1995 ; Cluzel *et al.*, 2001).

The Early Late Cretaceous (120-100 Ma) to Earliest Eocene (55-50 Ma) period: marginal rifting and spreading along the east Gondwana margin

This extensional tectonics period is marked by dislocation of the east Gondwana margin by Late Cretaceous rifting and subsequent Latest Cretaceous to Latest Paleocene/Earliest Eocene spreading. This results in a series of at least two thinned fragments of continental crust (the Lord Howe Rise, the Norfolk Ridge and possibly the Loyalty Ridge ? and/or a ridge further to the east ?) and at least three main basins floored with oceanic crust (the Tasman Sea Basin (100-85 Ma stretching, 85/80-52 Ma spreading), the New Caledonia Basin (95-62 ? Ma stretching, 62-56 Ma ? spreading) and the South Loyalty Basin (100/80 ? or 85/80 -55 Ma ? spreading). Ages of the crust of the two latter basins still derive from interpretations.

The major problem in the reconstruction for this period is the nature of the plate boundary east of the area. Reconstructions have suggested this boundary is an east dipping subduction zone, a west-dipping subduction zone, a strike-slip boundary, or no boundary at all. The recent reconstructions (Cluzel *et al.*, 2001 ; Crawford *et al.*, 2003; Shellart *et al.*, 2006) propose a continous west-dipping subduction of the Pacific plate that rolled back eastward to accomodate the basins opening. However, this attractive model does not fit well with the « classic » formation of successive back-arc basins in which the closest basin to the trench is the youngest, because the basins are thought to be more or less similar in age (except if an older age -Early Cretaceous- for the South Loyalty Basin is chosen). Also, the associated volcanic arc accompanying this long term subduction is still largely undocumented, even if one may consider that it disappeared by erosion or it is hidden beneath younger volcanic arcs or sediments.

The Early Eocene (55-50 Ma) to Latest Eocene/Early Oligocene period: lithospheric shortening and the New Caledonian orogen

This period is marked by convergence inside the previously dismembered east Gondwana margin, by the partial closure of North Loyalty Basin and finally by the emplacement of the ophiolitic nappe in New Caledonia. Recent reconstructions (Cluzel *et al.*, 2001 ; Schellart *et al.*, 2006) propose a relatively long-lived intra-oceanic east-dipping subduction zone inside the South Loyalty Basin during Early to Middle Eocene. Subduction was locked in the Late Eocene (38-34 Ma) by the underthrusting of the Norfolk ridge, resulting in the thrusting over the New Caledonia block of mafic and ultramafic units with coeval (60-70 km) exhumation of metamorphic rocks by buoyancy-driven uplift. In the final stage, the convergence motion jumped westward along the west margin of the New Caledonia ridge, as suggested by compressive features at the toe of the western margin, the abandoned slab beneath southern New Caledonia (Régnier, 1988) and the post-orogenic Lower Oligocene intrusives. Compression also affected the Fairway Ridge and Lord How Rise.

In the above-mentioned model, a large portion of the oceanic lithosphere of the wide South Loyalty Basin would have been absorbed in subduction, and the subduction would be responsible for the development of the Loyalty Ridge arc and the North Loyalty back-arc basin. However, the fore arc of this subduction zone does not resemble classic fore arc domain, and volcanic rocks with arc affinity are still unknown in the Loyalty ridge, and the direction of spreading in the North Loyalty Basin, deduced from E-W trending magnetic anomaly lineations, is parallel -instead of normal- to the arc and the subduction zone. An alternative model proposed here is to consider a narrow original North Loyalty basin from which a small panel of lithosphere forming the foot of the New Caledonia Ridge's margin has been obducted, with no requirement of Eocene subduction to create the Loyalty Ridge.

The Oligocene to Late Miocene period : planation, extensional tectonics and subsidence, volcanism

This poorly-documented post orogenic period is marked by (low to moderate ?) uplift, erosion and planation of the New Caledonia ridge. Isostatic uplift is likely accompanied by ridge-normal extensional tectonics and subsidence of the margins, and could be due to the buoyancy force of the orogenic root thickened during collision, to loading by the overthrusted oceanic lithosphere and to erosion. Ni-rich lateritic mantles likely mainly developed during this period of extensive alteration of the peridotites.

This period is also marked by alkaline volcanism related to several hot spots that affect the Australia plate, as exemplified by the Tasmantid chain in the Tasman Sea Basin, the Lord Howe volcanic chain and the volcanic edifices along the Norfolk Ridge/Loyalty Ridge.

The Late Miocene (10 Ma) to Present period : extension, initiation of subduction, bulge-related deformations and arc-ridge subduction/collision
The Latest Miocene is marked, to the east of the New Caledonia domain, by the initiation of the east-dipping Vanuatu subduction zone. This zone of convergence, behind which the active North Fiji Basin opened and along which part of the North Loyalty Basin (and other basins and features ?) was consumed, appears to play a significant role in the Late Neogene evolution of New Caledonia and Loyalty Ridges, as shown by (1) Late Neogene ridge-parallel to oblique extensional tectonics that affected the ridges, (2) Quaternary faults, bulge-related Quaternary vertical motions of the Loyalty islands and southern part of the New Caledonia Ridge, (3) Loyalty Ridge-Vanuatu arc interaction-related deformation and (4) shallow seismicity of relatively low intensity in the Grande Terre (intense seismicity, however, occurs in the easternmost part of the Loyalty Ridge close to the active Vanuatu plate boundary).

Some geological aspects for the understanding of the New Caledonia biodiversity and endemism

The problem of lands for refuge of the Gondwanian flora and fauna
The geology of the Grande Terre of New Caledonia indicates that after the Cretaceous the Norfolk Ridge/New Caledonia Ridge was below sea level up to the Late Eocene (or possibly up to the Middle Eocene), thus for a period of about 20 Ma (from 65 to 45 Ma). Indeed, the Paleocene-Lower Eocene pelagic limestones and cherts following the Upper Cretaceous fining-upward clastic sequence indicate relatively deep water deposits. Presence of an island on the New Caledonia Ridge is established since the Late Eocene, the size of which was larger than today since the d'Entrecasteaux Reef area, the wide northern lagoon, the wide southern lagoon, as well as parts of the upper slope of the present-day margins were probably above sea level. Other islands likely existed in the Late Eocene-Early Oligocene, on the Lord Howe Rise, Fairway Ridge and Loyalty Ridge. Islands also existed in Late Oligocene on the Chesterfield/Bellona zone, and in Late Miocene on the Loyalty Ridge. Thus, one may infer that the Gondwanian fauna and flora were introduced in New Caledonia during or after the Middle to Late Eocene.
If New Caledonia was not a refuge for the Gondwanian biodiversity, did other land -except Australia - exist in the region before the Middle-Late Eocene ? This in turn poses the major question concerning the nature of the plate boundary east of Australia before 45-50 ma. If we consider (as proposed by Veevers *et al.*, 2000; Cluzel *et al.*, 2001; Crawford *et al.*, 2003; Schellart *et al.*, 2006) that a west dipping subduction continuously proceeded since that time during rifting and subsequent basin spreading, therefore a volcanic arc must have been always active. This provides the possibility for the building of an alignment of islands likely restricted in size east of the Norfolk-New Caledonia Ridge, such as the present-day islands of the active Tonga or Vanuatu arcs.

The influence of the New Caledonian ophiolite
It is well known that the nature of the ultra basic ophiolitic nappe of New Caledonia plays an important role in the floral endemism of the island. The peculiar speciation and diversification are driven by the unusual type of ultra basic rock-derived soils rich in metallic elements and depleted in mineral salts. Most of the flora living on lateritic soils is endemic. Today ultrabasic rocks and associated soils only cover about one third of the island. The ophiolitic nappe constituting the top of the tectonic pile of the Late Eocene orogen, it is likely that when it emerged the Grande Terre island was fully covered by the ophiolite, leading to species selection. However, the present-day ophiolitic nappe

lacks a complete crustal sequence (sheeted dike complex and mafic rocks) and we do not know if the mafic crustal sequence existed on the top of the ultramafic rocks or if it has been removed by erosion.

An other point about the possible role of the ophiolitic nappe for the biodiversity is alkaline hydrothermalism resulting from reactions between water and upper mantle rocks (serpentinization). Recently (Pelletier *et al.*, 2006) an extensive active alkaline hydrothermal field has been revealed by 2004 and 2005 swath mapping and scuba diving in the floor of Prony Bay which developed in the large peridotitic massif at the southern tip of the Grande Terre, the "Prony aiguille" chimney was previously known (Launay and Fontes, 1985). Uncommon characteristics of this alkaline hydrothermal system are similar (same high pH, same deposits of CaCo3 and Mg hydroxyde and same peridotitic substratum) to those of the Lost City site recently discovered on the Mid-Atlantic Ridge (Kelley *et al.*, 2001) and which produces methane- and hydrogen-rich fluids serving as energy sources for archaeal and eubacterial communities (Kelley *et al.*, 2005). Such alkaline systems which derive from hydration of the outcropping ultramafic rocks is interesting because it may share several characteristics with hydrothermal environments at the beginning of life on Earth. The Prony Bay site is accessible by scuba diving and thus may be of wide interest. It is possible that such other active or recent hydrothermal sites exist in the present-day submerged portions of the remaining ophiolitic nappe (parts of southeastern and eastern lagoons of the Garnde Terre). Probably, such hydrothermal fields already existed in the past since the Late Eocene emplacement of the ophiolitic nappe, and may have played a first-order role in the evolution of the New Caledonian biodiversity.

Pleistocene sea level variation

The sea level has 100 Ka cyclic high and low stands since the last 0.9 Ma (Shackelton, 1987, 2000). Low stands were 120-130 m below present-day sea level and high stands were sometimes 5 to 10 m above present-day sea level. Such high amplitude and rapid variations may have significant effects on the distribution of the New Caledonia marine biodiversity. Taking into account the slow subsidence of the margins, the wide lagoon around the Grande Terre with depths much lower than 120 m has been largely emerged and submerged several times during the last million years. Detailed bathymetry of the lagoon however suggests that lakes with brackish water existed during sea falls in the deepest parts of the lagoon, the depths of the passes being generally shallower than that the lagoon itself (Dugas *et al.*, 1980; Chevillotte *et al.*, 2005). Lakes were likely in the relatively deep (60 m) and closed northern lagoon (Collot *et al.*, 1988), as well as in the southern lagoon as suggested by 25-35 m deep closed basins in front of the Pirogue river and east of the Mato Pass. The production of maps at different times during the last cycles, especially during the last glaciation/deglaciation period (last 125 ka), taking into account subsidence, sedimentation and detailed bathymetry of the lagoon, barrier reef and passes may be useful to locate possible refuges through Late Pleistocene for the species of the lagoon.

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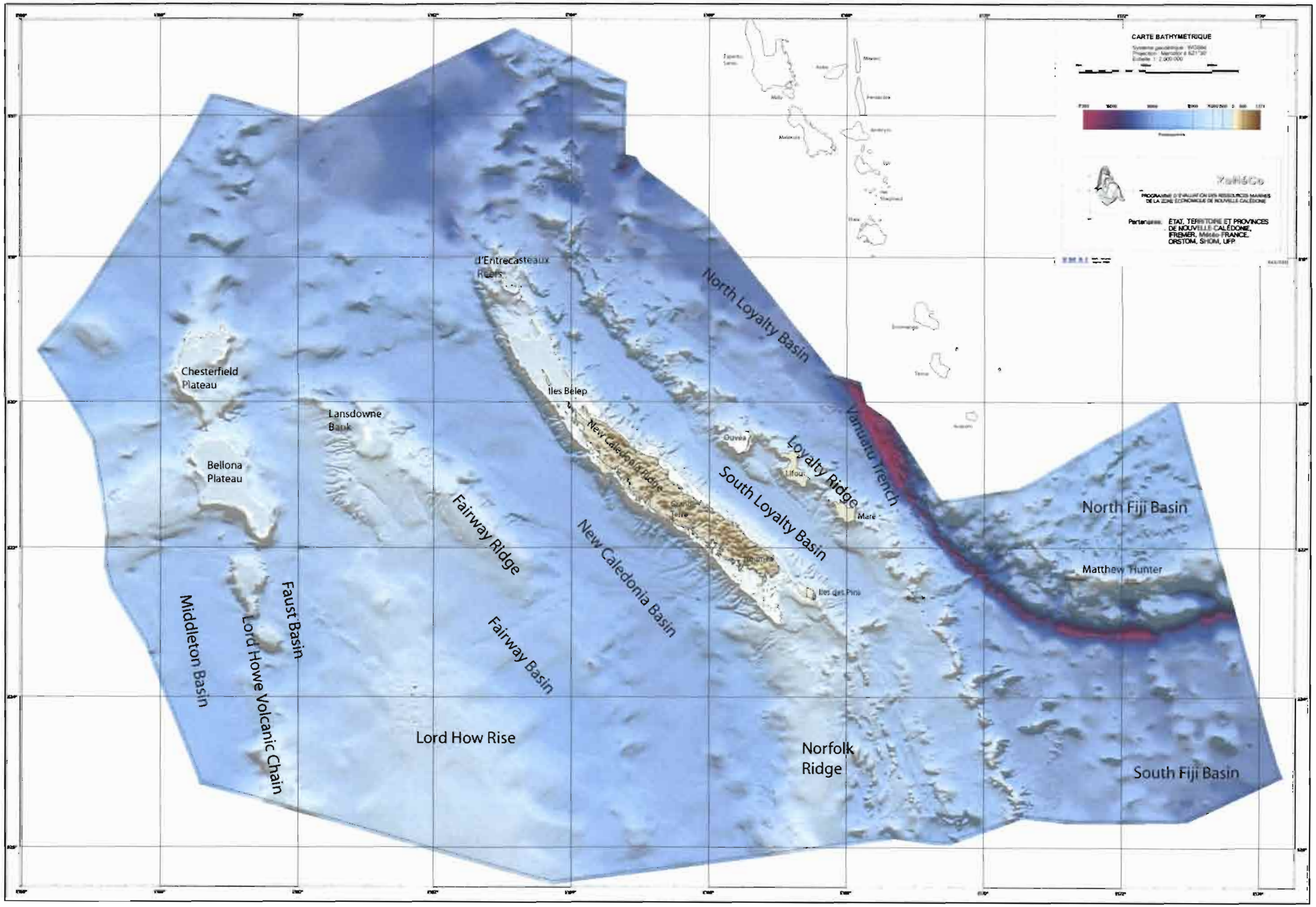

Figure 1: Bathymetry of the New Caldonia Exclusive Economic zone (From ZoNéCo Programm, 1998)

Plate 1 / 1

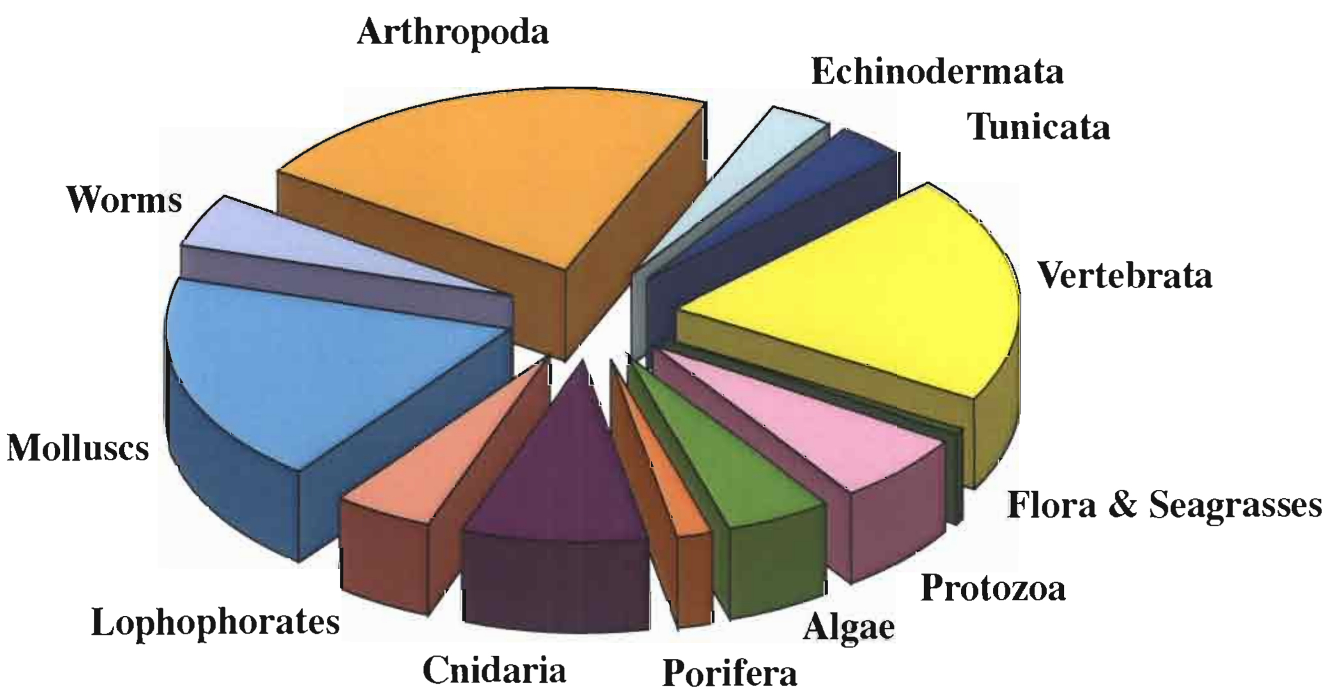

Figure 2. Divisions of the marine biodiversity of New Caledonia between the different groups as established from the species lits in this volume (total of 8783 species).

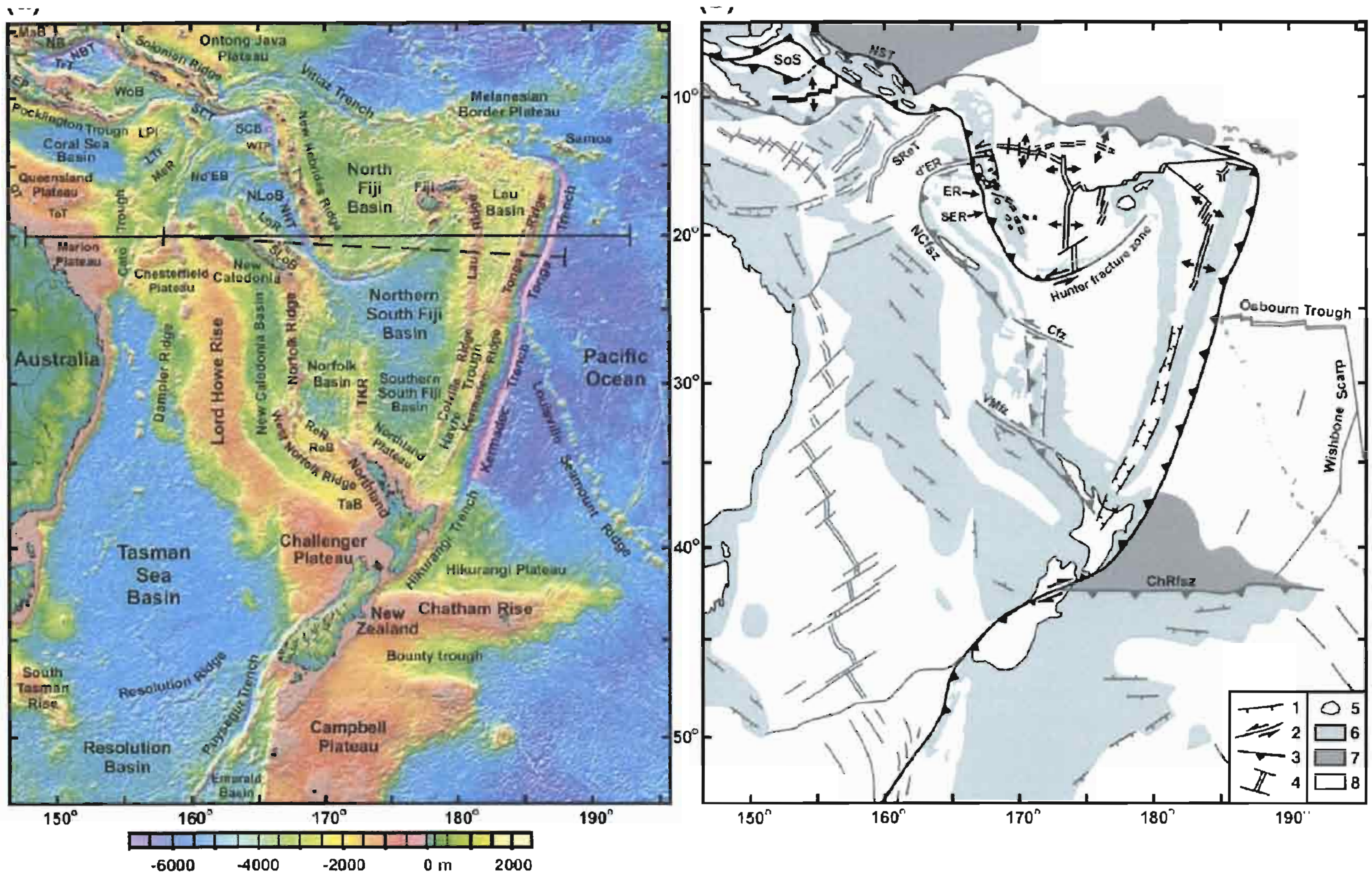

Fig. 2 . (a) Topography and bathymetry of the Southwest Pacific region (fromSmith and Sandwell (1997)) and (b) regional tectonic setting of (a). Cfz, Cook fracture zone; ChRfsz, Chatham Rise fossil subduction zone; d'ER, d'Entrecasteaux Ridge; EP, East Papua; ER, Efate Re-entrant; LPl, Louisiade Plateau; LoR, Loyalty Ridge; LTr, Louisiade Trough; MaB, Manus Basin; MeR, Melish Rise; NB, New Britain; NBT, New Britain Trench; NCfsz, New Caledonia fossil subduction zone; Nd'EB, North d'Entrecasteaux Basin; NHT, New Hebrides Trench; NLoB, North Loyalty Basin; NST, North Solomon Trough; QT, Queensland Trough; ReB, Reinga Basin, ReR, Reinga Ridge; SCB, Santa Cruz Basin; SCT, San Cristobal Trench; SER, South Efate Re-entrant; SLoB, South Loyalty Basin; SoS, Solomon Sea; SReT, South Rennell Trough; TaB, Taranaki Basin; TKR, Three Kings Ridge; ToT, Townsville Trough; TrT, Trobriand Trough; VMfz, Vening Meinesz fracture zone; WoB, Woodlark Basin; WTP, West Torres Plateau. 1, normal fault; 2, strike-slip fault; 3, subduction zone; 4, spreading ridge (double line) and transform faults (single lines); 5, land; 6-8, sea, with 6, continental or arc crust; 7, oceanic plateau; and 8, basin/ocean floor. Structures in light grey indicate that they are inactive. Thick continuous east-west line at latitude 20° S in panel (a) shows location of cross-section plotted in Fig. 3 h.

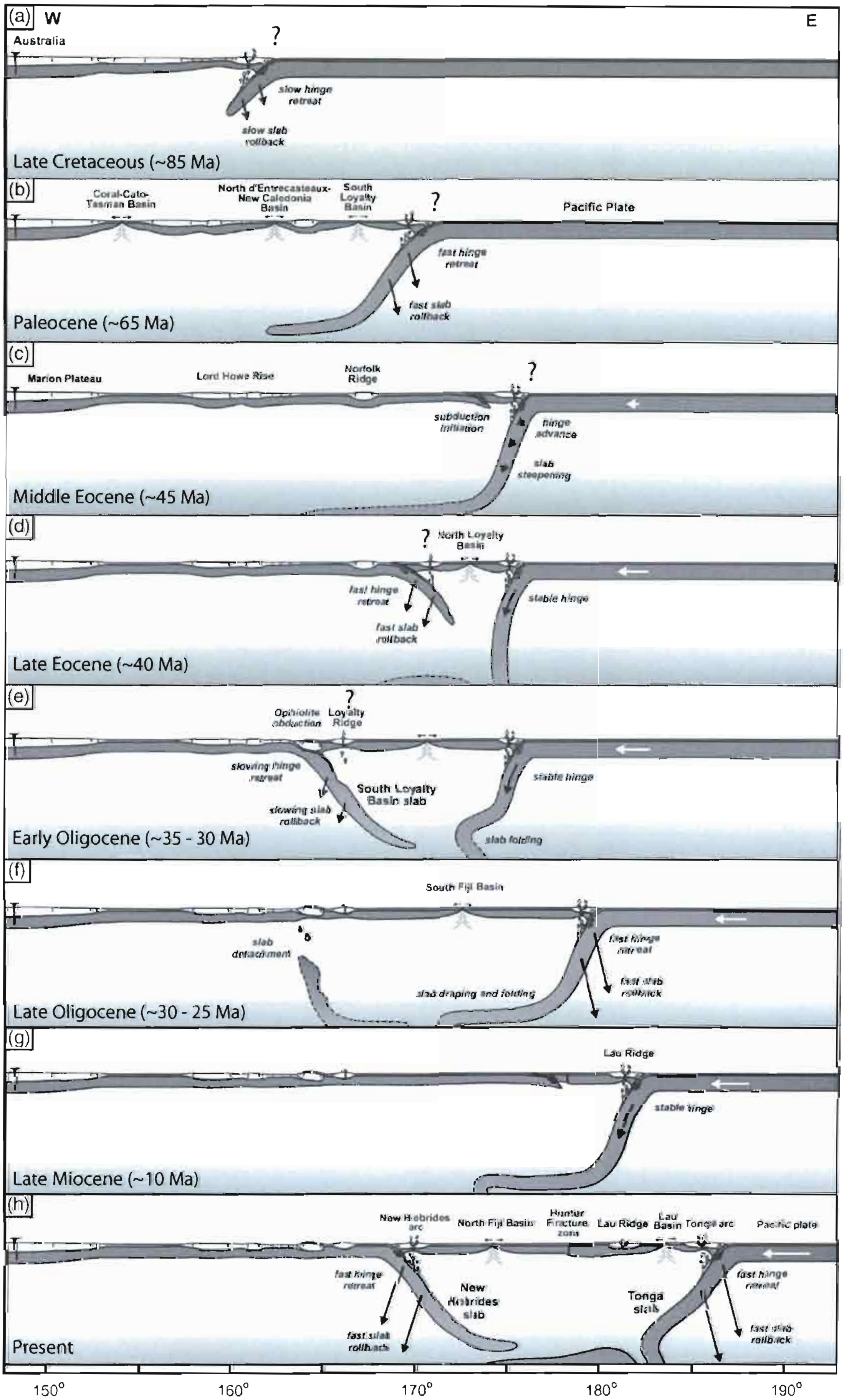

Figure 3 : East-west cross-sections illustrating the evolution of the Southwest Pacific region since the Late Cretaceous. Line of section is shown on Fig 2a. (modified from Schellart et al, 2006)

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Volume spécial

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