Plate reconstructions and tomography reveal a fossil lower mantle slab below the Tasman Sea

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ABSTRACT

The Southwest Pacific region is tectonically complex and is home to numerous fossil and active subduction zones. At the Earth's surface, there remains a geological controversy regarding the polarity and continuity of fossil subduction zones in New Zealand and New Caledonia, the origin of obducted ophiolites, the presence of high-pressure metamorphism, the occurrence of widespread Cenozoic magmatism, and the potential disappearance of one or more ocean basins. This controversy can be solved by looking at the lower mantle rather than at the Earth's surface. New P-wave and S-wave mantle tomography models from the Southwest Pacific are presented, which identify a previously unrecognized lower-mantle high-velocity anomaly that cannot be linked to Pacific subduction. The anomaly is located below the Tasman Sea at ~1100 km depth, strikes NW–SE and is ~2200 km by 600–900 km in lateral extent. By combining relative and absolute plate motions it is demonstrated that when the geological structures at the surface are reinterpreted as a single northeast-dipping 2500-km middle Cenozoic subduction zone (the so-called New Caledonia subduction zone) the lower mantle anomaly can be accounted for, as it is found at the predicted location and depth. Discovery of the lower mantle slab anomaly thereby solves a long-standing geological controversy in the New Zealand–New Caledonia region. Finally, reconstructions and analytical calculations predict a lower mantle slab sinking velocity of ~1.5 cm/yr and a lower mantle viscosity of ~10²² Pa·s.

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1. Introduction

The geodynamic processes responsible for widespread Tertiary magmatism, crustal deformation, high-pressure metamorphism, and emplacement of two massive ophiolite sheets (surface area of ~3.5×10⁴ km² and ≥3×10⁴ km²) in Northland (New Zealand) and New Caledonia (Fig. 1) are controversial. The geology of the two regions was interpreted as resulting from two separate subduction zones, a northeast-dipping subduction zone in New Caledonia (Regnier, 1988; Aitchison et al., 1995; Cluzel et al., 2001; Spandler et al., 2005; Baldwin et al., 2007) and a southwest-dipping subduction zone (Malpas et al., 1992; Herzer, 1995; Hayward et al., 2001; Whattam et al., 2004; Mortimer et al., 2007) or northwest-dipping subduction zone (Brothers, 1984; Kamp, 1984) in Northland. However, such local models involving opposite-dipping slabs cannot be integrated into a unifying tectonic model for the entire Southwest Pacific (Schellart et al., 2006; Schellart, 2007). In New Caledonia, evidence for northeast-directed subduction is strong, including an Eocene volcanic arc

(Loyalty arc) (Aitchison et al., 1995; Crawford et al., 2003), subduction-induced high-pressure metamorphism (Cluzel et al., 2001; Spandler et al., 2005), fore-arc boninites in the ophiolite (Crawford et al., 2003), a north-dipping low velocity zone below New Caledonia and a gravity-low along the west coast of New Caledonia (Regnier, 1988).

In Northland, southwest or northwest-directed subduction models fail to meet several geological, geophysical and spatio-temporal constraints (Schellart, 2007). For example, geological and geophysical investigations do not show any evidence of a fossil trench or accretionary wedge to the northeast of Northland. For further details and descriptions of additional constraints please see Schellart (2007). It is found that Early Miocene volcanism in the Northland region is better explained with a northeast-dipping slab. In between New Caledonia and Northland, the Three Kings Ridge has been explained by west or east-dipping subduction, but recently recovered fore-arc boninites (Mortimer et al., 2007) and high-pressure metamorphic rocks (Meffre et al., 2006) from the western slope of the ridge support the latter interpretation.

A new hypothesis proposes the former existence of a ~83–55 Ma backarc basin (South Loyalty Basin) to the east of the Norfolk–New Caledonia Ridge and to the north of the New Zealand region, which subducted along a single northeast-dipping subduction zone (Crawford et al., 2003; Schellart et al., 2006; Schellart, 2007), the so-

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Fig. 1. Tectonic map of the Southwest Pacific region (modified from Schellart, 2007). Note the location of the hypothesized d'Entrecasteaux–New Caledonia–Three Kings–Northland (New Caledonia in short) fossil subduction zone indicated by the thick grey line with the white teeth. 1—normal fault; 2—strike-slip fault; 3—subduction zone (teeth on upper plate side); 4—spreading ridge and transform faults; 5—land; 6—Eocene–Early Miocene magmatism; 7—Obducted ophiolites; 8—continental/arc crust; 9—oceanic plateau; 10—ocean/deep sea basin floor. d'ER—d'Entrecasteaux Ridge; NHT—New Hebrides Trench.

called New Caledonia subduction zone (Fig. 1). The terminal stage of subduction is hypothesized to be responsible for widespread magmatism and southwest-directed ophiolite obduction in New Caledonia and Northland (Schellart, 2007). Calculations from relative plate motions imply that subduction along the New Caledonia subduction zone started at 50 and 40 Ma (± 5 Ma) in the north and south, respectively (Schellart, 2007). Subduction amounted to a maximum of 1100, 600 and 500 km in the New Caledonia, Three Kings and Northland regions, respectively, as implied by the original geometry of the South Loyalty Basin (Schellart et al., 2006). Schellart (2007) proposed that the New Caledonia slab detached in the final stage of subduction and that the timing of detachment could be constrained by the timing of obduction and timing of magmatism. Assuming that obduction is largely controlled by slab pull forces, and assuming that certain magmatism in Northland, New Caledonia and the Norfolk basin is detachment-induced, then the timing of detachment has to be after the timing of obduction and before the timing of detachment-induced magmatism. The time of detachment was simply assumed to be right in between these two geological events. As such, the time of detachment is estimated at 30, 27 and 24 Ma (± 3 Ma) in southern New Caledonia, the Norfolk basin and northwest Northland, respectively (Schellart, 2007). The timing suggests that slab detachment started in the north and migrated southward along a sub-horizontal tear, comparable to detachment models proposed for the Mediterranean and Tethyan subduction zones (Wortel and Spakman, 2000; Hafkenscheid et al., 2006).

Here we test the new hypothesis by coupling the regional reconstructions to two lower mantle hotspot reference frames. This allows us to make predictions regarding the current geographical location and depth of the fossil slab in the mantle. Analytical calculations provide an additional estimate for the current depth of the slab anomaly. The reconstructions predict that subduction of the South Loyalty Basin lithosphere lasted from ~50 Ma until ~25 Ma, that the slab during this subduction period was 20 to 60 Myr old and thereby ~50–87 km thick (Schellart, 2007). As discussed in Cloos (1993), oceanic lithosphere as young as 10 Ma is already neutrally buoyant with respect to the underlying asthenosphere, so 20–60 Myr old oceanic lithosphere should readily subduct, more so if the crust becomes progressively eclogitized, and detach at the final stage of subduction, and should most likely be visible as a high-velocity anomaly in mantle tomography models. Results from an S-wave tomography model and a new P-wave tomography model will be presented, indeed showing a high-velocity anomaly at the predicted location and depth to represent the fossil slab that subducted along the New Caledonia subduction zone.

2. Predictions from reconstructions

Reconstructions of subduction zone migration can predict the location of fossil slabs when combined with the timing of slab detachment, predicted slab-sinking patterns and sinking velocities. Fig. 2 shows the migration of the New Caledonia fossil subduction

zone from 48 Ma to Present from combining subduction zone migration relative to Australia (Schellart et al., 2006; Schellart, 2007) and motion of the Australian plate with respect to two hotspot reference frames (Gordon and Jurdy, 1986; Müller et al., 1993). The reconstructions thereby constrain subduction zone migration relative to a lower mantle reference frame. Fig. 2 also shows the timing of termination of subduction zone activity and slab detachment as deduced from Schellart (2007). Prior to slab detachment, west-directed trench retreat along the New Caledonia subduction zone resulted primarily from slab rollback, while the northward motion of the trench resulted from the northward motion of the Australian plate to which the South Loyalty slab was still attached. After slab detachment, which occurred at ~30 Ma in the north and ~24 Ma in the south, the fossil trench was basically a passive feature that moved ~northward along with the ~northward motion of the Australian plate.

Fig. 3 provides a three-dimensional perspective of the progressive subduction, sinking and detachment of the South Loyalty slab from 48 Ma to the Present. Subduction starts in the north (New Caledonia region, Fig. 3a) and progressively grows southward (Fig. 3b). At the final stage of subduction in the New Caledonia region a sub-horizontal slab tear develops (Fig. 3c) that propagates southward (Fig. 3d), resulting in detachment and sinking of the South Loyalty slab into the mantle. When slab detachment is complete, the entire South Loyalty Basin slab sinks vertically into the mantle (Fig. 3e–f). The diagrams further show development of a sub-vertical slab tear at the northernmost part of the subduction zone (below the d'Entrecasteaux zone) that allows for efficient slab-rollback-induced toroidal upper mantle return flow from behind the slab towards the mantle wedge side (e.g. Schellart et al., 2007; Schellart, 2008), thereby facilitating rapid slab rollback at the northernmost part of the subduction zone.

Geodynamic models of progressive subduction can provide constraints on the sinking patterns of slabs in the mantle. Such sinking

patterns are important to be able to determine the location of fossil slabs in the mantle. One can obtain an estimate of the sinking angle of upper mantle slabs from quantifying the degree of subduction partitioning, i.e. how much of the total subduction velocity (v_s) is due to the trench migration (v_T , trench retreat is positive) and how much is due to trenchward subducting plate motion (v_{sp}). Subduction partitioning can be expressed as: v_T/v_s . Laboratory models (e.g. Schellart, 2005, 2008) show that for $v_T/v_s > 0.5$ the slab sinks backwards (oceanward), for $0 \leq v_T/v_s < 0.5$ the slab sinks subvertical, and for $v_T/v_s < 0$ the slab sinks forward (towards the overriding plate). Fig. 2 indicates that for the New Caledonia slab, some 50–75% of the total subduction rate is due to trench retreat ($v_T/v_s = 0.50–0.75$). 3D laboratory models of progressive subduction with a free trench show that when $v_T/v_s = 0.50–0.75$, then slabs sink vertically (90°) to backward at angles of 70° in the upper mantle (Schellart, 2005, 2008; see for example Fig. 4b, c in Schellart, 2005). One can thus expect that the South Loyalty slab would have sunk vertically downward to backward (west to south-westward) at $\sim 70^\circ$ during its descent in the upper mantle. Furthermore, lower-mantle slabs sink sub-vertically (Pysklywec et al., 2003; Schellart et al., 2007).

The NW–SE extent of the fossil slab is constrained by the width (trench-parallel extent) of the subduction zone (max. 2500 km), while its NE–SW extent is constrained by the original size of the South Loyalty Basin (average of 750 km) (Schellart et al., 2006). These geometrical constraints, in combination with the trench migration patterns (Fig. 2) and inferred sinking kinematics for the slab as deduced from geodynamic modelling indicate that the fossil South Loyalty slab is laterally confined by the green dashed line in Fig. 2.

The reconstructions further indicate that the subduction velocity is on average ~6 cm/yr (500–1100 km of subduction in 10–16 Myr), implying an upper mantle sinking velocity of similar magnitude (Schellart, 2005). Sinking velocities are thought to reduce by a factor of four upon entering the lower mantle (Lithgow-Bertelloni and Richards,

Fig. 2. Migration of the New Caledonia subduction zone from 48 Ma to 0 Ma (Present) in two reference frames by combining subduction zone migration relative to Australia (Schellart et al., 2006), timing of slab detachment (Schellart, 2007) and motion of the Australian plate in (a) a global hotspot reference frame (Gordon and Jurdy, 1986) and (b) an Indo-Atlantic hotspot reference frame (Müller et al., 1993). Different trench positions are found in the two reference frames, which are small (0–100 km) from 25 to 0 Ma but amount up to 370 km at 48 Ma. Location of the subduction zone for distinct times is indicated by the blue, purple, red, orange, yellow and brown lines. Numbers indicate time in Ma. Teeth are located on the upper plate side and indicate active subduction. Dotted lines indicate detached slab while arrows indicate direction of lateral slab tear propagation. Dashed green line indicates predicted lateral extent of the detached South Loyalty slab.

1998), suggesting a lower mantle sinking velocity of ~ 1.5 cm/yr. The lower mantle sinking velocity can also be estimated by calculating the sinking velocity of an oblate ellipsoid (V_E) parallel to its shortest axis (Kerr and Lister, 1991), with $V_E = Sg(\rho_S - \rho_M)D^2/18\mu_{LM}$. In the special case where the ellipsoid is a sphere, then $S=1$ and the equation is identical to the Stokes equation. Here, S is the shape factor (Kerr and Lister, 1991) with $S \approx 0.37$ for the South Loyalty slab, g is the gravitational acceleration (9.8 m/s^2), $(\rho_S - \rho_M)$ is the density contrast between slab and mantle (80 kg/m^3), $D = (abc)^{1/3}$, μ_{LM} is the lower mantle viscosity ($\sim 10^{22} \text{ Pa}\cdot\text{s}$), and a , b and c are the shortest ($\sim 70 \text{ km}$), intermediate ($\sim 750 \text{ km}$) and longest axis ($\sim 2500 \text{ km}$) of the ellipsoid (detached South Loyalty slab), respectively. The calculations give $V_E \approx 1.3 \text{ cm/yr}$, which is comparable to the lower mantle sinking velocity cited above.

One assumption that is incorporated into this calculation regards the density contrast between slab and mantle. This was previously investigated by Cloos (1993), who did calculations for oceanic lithosphere with a 7 km thick oceanic crust, of which the crust was eclogitized during subduction. Following these calculations, the density contrast between the South Loyalty slab with eclogitized crust (which was $20\text{--}60 \text{ Myr}$ old at the time of subduction and $\sim 50\text{--}87 \text{ km}$ thick) and the ambient upper mantle is then $\sim 78\text{--}84 \text{ kg/m}^3$. Considering that the upper mantle sinking velocity was relatively high ($\sim 6 \text{ cm/yr}$) and the time that the fossil slab has spent in the lower mantle (decreasing from an average of 30 Myr in the north to an average of 20 Myr in the south), it is expected that the slab has retained a large part of its thermal structure. So as a first-order approximation, a slab-mantle density contrast of 80 kg/m^3 for both the upper mantle and lower mantle appears reasonable.

With subduction starting at $\sim 50 \text{ Ma}$ in the north, slab detachment in New Caledonia at $\sim 30 \text{ Ma}$, average upper and lower mantle sinking velocities of 6.0 cm/yr and 1.5 cm/yr , and a transition zone at 670 km , the slab tip and trailing edge of the slab would presently be at 1250 and 950 km depth, respectively. Along the Three Kings and Northland sections, subduction started at ~ 45 and $\sim 40 \text{ Ma}$ and detachment occurred at ~ 27 and $\sim 24 \text{ Ma}$, respectively. Thus, the slab tip and trailing edge would now have reached a depth of 1180 and 910 km (Three Kings), and 1100 and 860 km (Northland). Thus, we predict that the fossil South Loyalty slab is confined by the green dashed line in Fig. 2 at $\sim 950\text{--}1250 \text{ km}$ depth, $\sim 910\text{--}1180 \text{ km}$ depth and $\sim 860\text{--}1100 \text{ km}$ depth in the north, middle and south, respectively. To test this prediction we use mantle tomography to identify structures in the mantle beneath the Southwest Pacific that could represent the fossil slab.

3. Southwest Pacific tomography models

Tomographic imaging of the Earth's mantle has allowed detailed visualization of actively subducting slabs and detached slabs in the deep mantle. Figs. 4 and 5 illustrate the Southwest Pacific mantle

Fig. 3. Block diagrams illustrating the progressive migration of the New Caledonia subduction zone (in the Indo-Atlantic hotspot reference frame from Müller et al., 1993) and the progressive subduction, sinking and detachment of the South Loyalty slab. (a) Incipient subduction in the North. (b) Progressive subduction by slab rollback and southward growth of the subduction zone. At the northernmost part of the subduction zone (below the d'Entrecasteaux zone) a sub-vertical slab tear develops that allows for efficient slab-rollback-induced toroidal upper mantle return flow from behind the slab towards the mantle wedge, thereby facilitating rapid slab rollback at the northernmost part of the subduction zone. (c) Final stage of subduction in the north (New Caledonia region), with inception of slab detachment due to formation of a sub-horizontal slab tear that propagates southward. Subduction continues in the Norfolk–Northland region. (d) Continued southward propagation of the sub-horizontal slab tear. Active subduction is present only in the southernmost part (Northland region). (e) Slab detachment is now complete and the entire South Loyalty Basin slab is now located in the lowermost upper mantle and uppermost lower mantle. (f) Continued sinking of the South Loyalty Basin slab in the lower mantle.

Fig. 4. S-wave (a–d) and P-wave (e–h) tomographic images from the Southwest Pacific region at depths of (a) 910 km, (b) 1100 km, (c) 1299 km, (d) 1500 km, (e) 900 km, (f) 1100 km, (g) 1300 km and (h) 1500 km. Note the strong NW–SE striking fast anomaly below the Tasman Sea between New Zealand and Australia (SLBS), which is most pronounced at 1100 km depth. The anomaly is interpreted as the fossil slab of the subducted South Loyalty Basin. The high velocity anomaly below the North and South Fiji basins represents the subducted Tonga–Kermadec–Vitiaz slab (TKVS). Velocity perturbation varies between -1.5% and $+1.5\%$ for a–d, -0.9% and $+0.9\%$ for e, -0.6% and $+0.6\%$ for f, and -0.5% and $+0.5\%$ for g–h.

structure with a global S-wave model and P-wave model. Details on the S-wave model can be found elsewhere (Kennett and Gorbato, 2004). The new global P-wave model P06 (Amaru, 2007) is based on a travel time data set that has approximately tripled in size compared to the data set used for an older model (BS2000, Bijwaard and Spakman, 2000). In particular, core phases have been incorporated to better constrain the lower mantle structure and the data set contains considerably more regional travel times.

The S-wave and P-wave models show high velocity anomalies at 900 km and 1100 km depth below the North and South Fiji basins, which represent the deep continuation of the Tonga–Kermadec–Vitiaz slab (van der Hilst, 1995; Hall and Spakman, 2002; Pysklywec et al., 2003). Indeed, the tomography models, in particular the P-wave model, show a north–south striking high-velocity anomaly. This is in agreement with Cenozoic reconstruction models for the central and southern part of the Tonga–Kermadec–Vitiaz subduction zone (e.g. Hall and Spakman, 2002; Schellart et al., 2006; Sdrolas and Müller, 2006).

Of particular interest for this study is a pronounced fast NW–SE striking anomaly in the lower mantle below the Tasman Sea. The fast anomaly is best resolved at 1100 km depth (Fig. 4b,f) and has lateral dimensions of ~ 2200 km by ~ 600 – 900 km. The high-velocity anomaly is less resolved at 1300 km depth, and disappears at 900 and 1500 km depth. Cross-sections shows that the anomaly is limited to a depth of ~ 950 – 1350 km and is most pronounced at a depth of ~ 1000 – 1200 km (Fig. 5). The S-wave anomaly has a higher amplitude than the P-wave anomaly, as generally observed for lower mantle slabs (Kennett and Gorbato, 2004). The NW–SE striking high-velocity anomaly is situated in the predicted location to represent the fossil South Loyalty slab with the predicted NW–SE orientation and predicted lateral dimensions (Fig. 6).

The results from the S-wave tomography model and P-wave tomography model show a good agreement with respect to the geometry of velocity anomalies. For example, the cross-sections in Fig. 5 compare very well, both showing the flat-lying South Loyalty Basin slab at 1000–1200 km, the upper mantle Kermadec slab, an apparent hole in the Kermadec slab, the potential lower mantle continuation of the Kermadec slab, and two pronounced low velocity zones, one above the South Loyalty Basin slab anomaly and one to the southeast of the Kermadec subduction zone. The fact that the two different models show such good agreement makes us confident that the South Loyalty Basin slab anomaly is well resolved.

Several other high-velocity anomalies are observed in Fig. 4. In particular, a high-velocity anomaly at ~ 900 depth below the north-eastern Australia–Coral Sea–New Guinea region (Fig. 4a and e) could represent fossil slab remnants related to Eocene–Oligocene subduction along the Pocklington subduction zone (e.g. Hall and Spakman, 2002; Schellart et al., 2006). The S-wave model and P-wave model also show a high-velocity anomaly below southeast Australia at ~ 900 km depth. This anomaly remains visible in the P-wave model at 1100 km and 1300 km depth, with a WNW–ESE strike and extending towards the South Island of New Zealand. The amplitude of the anomaly, however, is rather small. For example, at 1300 km depth, it is only about $+0.20\%$ with a range of $+0.06$ to $+0.31\%$ in the P-wave model, while the S-wave model does not show a WNW–ESE striking feature. Thus, the anomaly might be insignificant.

4. Resolution tests for tomography models

Resolution tests for the P-wave model P06 have been conducted and are shown in Figs. 7 and 8. Fig. 7 shows results from spike

Fig. 5. Northwest–southeast striking S-wave and P-wave tomographic cross-sections from the Southwest Pacific region. (a) Cross-section of the S-wave model. (b) Cross-section of the P-wave model. (c) Map showing the location of the cross-sections. Note the flat-lying high-velocity anomaly below the Tasman Sea (SLBS), which is most pronounced at 1000–1200 km depth. The anomaly is interpreted as the fossil slab of the subducted South Loyalty Basin. Both the S-wave model and P-wave model also show the northwest-dipping Kermadec slab in the upper part of the upper mantle, a hole in the lower part of the upper mantle and uppermost lower mantle, followed by a high-velocity anomaly that could be interpreted as the lower mantle continuation of the Kermadec slab. In the P-wave model, this lower mantle high velocity anomaly dips northwestward at ~70° and continues down to ~2400 km depth. The hole could be due to insufficient ray coverage in that area. Note that the white lines in (b) mark 1% intervals.

resolution tests. Spatially the spikes correspond well to their input locations and strong smearing effects along possible dominant ray directions do not occur. The spikes are recovered with reduced amplitude primarily as a result of the imposed smoothness constraints (damping) during tomographic inversion. We infer from this, and other sensitivity tests, that the imaged mantle structures shown in Fig. 4e–h are sufficiently resolved to identify the fossil South Loyalty slab.

Fig. 8 shows a synthetic slab test to investigate if the ray coverage in the region is sufficient to reproduce the South Loyalty slab at 1100 km depth and 1300 km depth. The test demonstrates that the outline of an anomaly of the assumed 3D shape of the South Loyalty slab can be imaged properly with only minor leakage to the surrounding mantle in both the horizontal direction (Fig. 8f and g) and vertical direction (Fig. 8e and h). This demonstrates the detectability of the bulk volume and its orientation. The overall amplitude recovery shows systematic underestimation of synthetic amplitudes, which relates to incomplete convergence of the LSQR algorithm from Paige and Saunders (1982) combined with applied damping. Particularly under New Zealand the test demonstrates lack of detectability of detailed structure within the synthetic anomaly. This results from the non-homogeneous sampling of this mantle

Fig. 6. Predicted location of the South Loyalty slab from the global (model in Fig. 2a) and Indo-Atlantic (model in Fig. 2b) hotspot reference frames, and the observed location of the S-wave anomaly and P-wave anomaly at 1100 km depth.

Fig. 7. Spike resolution tests for the P-wave model for the Southwest Pacific region at depths of 900, 1100, 1300 and 1500 km. The spike size is 3° (~ 250 km at these depths) and 120 km, 150 km or 200 km in thickness. The input synthetic spikes have amplitudes of +5% or –5% and their shapes are outlined by the thin solid lines. The non-square character of some of the input-spikes is due to the model parameterization with irregular cells. The mantle region between the spikes has a 0% anomaly value to permit detection of smearing effects between spikes in all directions. Synthetic data were generated from the global spike models by integration of the synthetic slowness along the same ray paths as used in the real data experiment. A normal distributed error with standard deviation of 0.5 s was added to the synthetic data prior to inversion. The colour contours depict the results of the inversion. Contour limits of +2.0% and –2.0% (40% of the input value) were chosen as a compromise between showing higher-amplitude spike response and lower-amplitude smearing effects. The amplitudes returned from the inversion have a range of –2.0% to 2.5%, –1.5% to 1.25%, –1.5% to 1.0% and –1.25% to 1.25% for 900 km, 1100 km, 1300 km and 1500 km depth, respectively.

volume, which locally affects convergence properties of the solution (Spakman and Nolet, 1988).

5. Discussion

5.1. The South Loyalty fossil slab anomaly

The continuity and sharp southwestern boundary of the SLBS anomaly as shown in Fig. 4b and f agree with the proposed model of a

continuous ~ 2500 km subduction zone, and negate earlier models involving two separate subduction zones for New Caledonia and Northland. Indeed, previous tectonic models involving a northeast-dipping subduction zone in New Caledonia and a west or southwest-dipping subduction zone in Northland would predict two small separate high-velocity anomalies rather than one continuous high-velocity anomaly as shown here. The sharp northwesternmost boundary of the anomaly, in particular for the P-wave model, implies

Fig. 8. Synthetic slab test for the P-wave model to investigate if the ray coverage in the region is sufficient to reproduce the South Loyalty slab at 1100 km depth and 1300 km depth. (a–d) Synthetic anomaly input of the assumed 3D shape of the South Loyalty slab at four depths. (e–h) Amplitude recovery at four depths.

that the slab was indeed of limited trench-parallel extent, bounded on its northern side by the d'Entrecasteaux ridge, a fossil transform plate boundary. The d'Entrecasteaux ridge basically marks the surface expression of the fossil sub-vertical slab tear that migrated westward (Fig. 3b–c), conform the reconstructions from Schellart et al. (2006) and Schellart (2007), similar in function as the Hunter Fracture zone bounding the North Fiji Basin (e.g. Schellart et al., 2002) and conform a STEP fault as described in Govers and Wortel (2005).

The fast anomaly is best resolved at ~1100 km depth (Fig. 4b,f) and is situated in the predicted location to represent the fossil South Loyalty slab with the predicted NW–SE orientation and lateral dimensions (~2200 km by ~600–900 km) (Fig. 6). Cross-sections shows that the anomaly is limited to a depth of ~950–1350 km and is most pronounced at a depth of ~1000–1200 km (Fig. 5), as predicted by the reconstructions and the average upper and lower mantle sinking velocities of 6 cm/yr and 1.5 cm/yr, respectively. The P-wave model, and to a lesser extent the S-wave model, show that the southeastern extent of the South Loyalty anomaly abuts with the ~north–south striking Tonga–Kermadec–Vitiāz anomaly (Fig. 4f). This agrees with predictions from the regional reconstruction by Schellart et al. (2006), in which the opposite-dipping subduction zones were active contemporaneously, were widely separated in the north (~2000 km), but were very close (~100 km) at the southernmost extent of the New Caledonia subduction zone (southeastern Northland region).

5.2. Sinking velocities for lower mantle slabs

Fossil slabs have been recognized previously with tomographic imaging, including the Farallon slab underneath North America (Grand et al., 1997), the Mongol–Okhotsk slab underneath Siberia (Van der Voo et al., 1999), the Tethyan slabs underneath South Asia (Hafkenscheid et al., 2006), the fossil Gondwanaland margin slab underneath the Australian–Antarctic discordance (Gurnis and Müller, 2003) and the Vitiāz slab below the North Fiji Basin (Hall and Spakman, 2002; Pysklywec et al., 2003). The lower mantle free sinking velocity of 1.5 cm/yr for the South Loyalty slab is slower than that estimated for the Tethyan slabs (2 cm/yr; Hafkenscheid et al., 2006), but faster than the average mantle sinking velocity for the Mongol–Okhotsk slab (1 cm/yr; Van der Voo et al., 1999). The Tethyan slab volumes are larger than the South Loyalty slab, thereby enhancing faster sinking. The lowermost part of the Mongol–Okhotsk slab appears to be resting on the core–mantle boundary, thereby partly supporting the trailing slab and retarding sinking. The South Loyalty slab is sinking freely, and thus a lower mantle sinking velocity of 1.5 cm/yr appears realistic considering its rapid trench and subduction velocities (4–8 cm/yr).

6. Conclusions

Discovery of the fossil lower mantle slab solves a long-standing geological controversy related to the lateral continuity of the New Caledonia fossil subduction zone, polarity of subduction, source of middle Cenozoic magmatic rocks, and mode of ophiolite obduction. The fossil subduction zone ran from New Caledonia in the north to Northland in the south, stretching for some 2500 km. The d'Entrecasteaux ridge functioned as a transform plate boundary (i.e. a STEP fault conform Govers and Wortel, 2005), and thereby represents the northern limit of the fossil subduction zone. The slab dipped north-east to eastward, and was responsible for arc volcanism along the d'Entrecasteaux–Loyalty–Three Kings–Northland Plateau seamount chain. In the final stage of subduction, the slab detached, causing detachment-induced magmatism and volcanism in New Caledonia, the Norfolk Basin and Northland (Schellart, 2007).

The discovery of the lower mantle slab necessitates an obduction mode in Northland and New Caledonia involving a subduction zone dipping away from the continental margin (~northeastward), which conforms the original theoretical obduction model from Moores

(1970) in which fore-arc ophiolites are obducted on top of a passive margin. The reconstructions predict a lower mantle slab sinking velocity of ~1.5 cm/yr and a lower mantle viscosity of ~ 10^{22} Pa·s. The considerable size of the subduction zone and fossil slab requires new models of global mantle circulation to incorporate this structure in future calculations.

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