Late Cretaceous to present-day opening of the southwest Pacific constrained by numerical models and seismic tomography

Kara J. Matthews[1], Maria Seton, Nicolas Flament, R. Dietmar Müller

Keywords: back-arc basin, eastern Gondwanaland, Lord Howe Rise, seismic tomography, South Loyalty Basin, southwest Pacific, subduction, Tonga-Kermadec Trench

Abstract

The southwest Pacific is a frontier region for petroleum exploration. A complex series of subduction and back-arc basin forming episodes characterises the late Cretaceous to present-day evolution of the region. Controversial aspects of the regional tectonic history include the presence or lack of subduction between 83 and 43 Ma, the polarity of subduction, the timing of back-arc basin formation, and whether or not Pacific plate motion can be tied to the motion of Australia via spreading in the Tasman Sea during the late Cretaceous-early Cenozoic. A combination of tectonic and geodynamic models has previously been used to propose that there was no subduction to the east of Australia between 83 and 43 Ma, with the Lord Howe Rise being part of the Pacific plate during this time period, contrary to alternative plate models that include a plate boundary to the east of the Lord Howe Rise. Determining which plate circuit to use for Pacific motion is critical for producing regional reconstructions for the southwest Pacific, and addressing specific problems on the chronology of tectonic and basin-forming events. To help resolve these long-standing disputes we test a recently published plate reconstruction in global mantle flow models with imposed plate motions. We use the 3D spherical mantle-convection code *CitcomS* coupled to the plate reconstruction software *GPlates*, with plate motions since 200 Ma and evolving plate boundaries imposed. We use seismic mantle tomography models to test the forward-modelled subduction history in the region. The reconstruction that we test incorporates east-dipping subduction from 85-45 Ma along the western margin of the Loyalty-Three Kings Ridge to close the South Loyalty Basin. Following collision of the Loyalty Ridge with New Caledonia, west-dipping Tonga-Kermadec subduction initiates along the eastern margin of the Loyalty Ridge and opens the North Loyalty, South Fiji, Norfolk and Lau basins. Contemporaneous with west-dipping Tonga-Kermadec subduction, there is short-lived east-dipping subduction between 36-18 Ma along the western margin of the Loyalty-Three Kings ridge. We find that subduction to the east of Australia during the period 85-45 Ma is necessary to account for the distribution of lower mantle slab material that is imaged by seismic tomography beneath New Zealand. Pacific plate motion therefore cannot be directly tied to Australia via spreading in the Tasman Sea as a convergent margin must have existed to the east of the Lord Howe Rise at this time. We suggest that adopting a plate circuit that ties Pacific plate motion directly to the Lord Howe Rise should be avoided for this period. An unexpected result is that the regional lower mantle structure provides strong evidence for a long-lived intra-oceanic subduction zone, located to the northeast of Australia at about 10-25°S and 170°E-170°W, that was active during at least the late Cretaceous. We propose that this subduction zone was located outboard of the plate boundary that separated the palaeo-Pacific ocean from the Tethys, and we speculate that arc remnants may be preserved in southeast Asia.

Introduction

Between Australia and New Zealand lies an important deepwater frontier region for hydrocarbon exploration (e.g. Hashimoto et al., 2010; Uruski, 2010). Basins along the Lord Howe Rise and surrounding New Zealand comprise thick sedimentary successions, in excess of several kilometres at some major depocentres (e.g. Hashimoto et al., 2010; Uruski, 2008). Prospectivity studies have shown petroleum potential in a number of basins such as the Fairway (Exon et al., 2007; Auzende et al., 2000), deep water Taranaki (Uruski, 2008), and Capel and Faust basins (Hashimoto et al., 2010; Rollet et al., 2012). However, well data and seismic lines are particularly sparse along the Lord Howe Rise, which makes robust plate kinematic models essential for interpreting and assessing hydrocarbon targets in a regional tectonic context.

The southwest Pacific (Fig. 1) has a complex tectonic history dominated by episodic back-arc basin formation (e.g. Whattam et al., 2008; Schellart et al., 2006; Crawford et al., 2003; Sdrolias et al., 2003), which may have started as early as the early Cretaceous behind the west-dipping eastern Gondwanaland subduction zone (Seton et al., 2012), prior to fragmentation of eastern Gondwanaland and break-up of Australia, New Zealand and Antarctica (Fig. 2). Several plate kinematic models have been presented for the late Cretaceous to present-day evolution of the region (e.g. Whattam et al., 2008; Schellart et al., 2006; Crawford et al., 2003; Sdrolias et al., 2003), while other studies have focused on constraining the evolution of specific basins, or the plate kinematics of the region during specific time-frames (e.g. Bache et al., 2012; Booden et al., 2011; Sutherland et al., 2010; DiCaprio et al., 2009).

Data coverage in parts of the southwest Pacific is sparse (Schellart et al., 2006) and several back-arc basins have been entirely or partially subducted, making it difficult to decipher the tectonic evolution of the region. Furthermore, it is difficult to date seafloor that formed during the Cretaceous Normal Superchron period, 120.4-83.5 Ma (timescale of Gradstein et al., 1994), as there were no reversals of Earth's magnetic field during this interval. These issues have contributed to several controversies concerning the tectonic evolution of the southwest Pacific. For instance, there is a long-standing disagreement over the nature of the southwest Panthalassa/proto-Pacific margin from the

1 EarthByte Group, School of Geosciences
The University of Sydney, NSW 2006, Australia
Email: kara.matthews@sydney.edu.au,
Phone: +61 2 9351 8093
Fax: +61 2 9351 2442

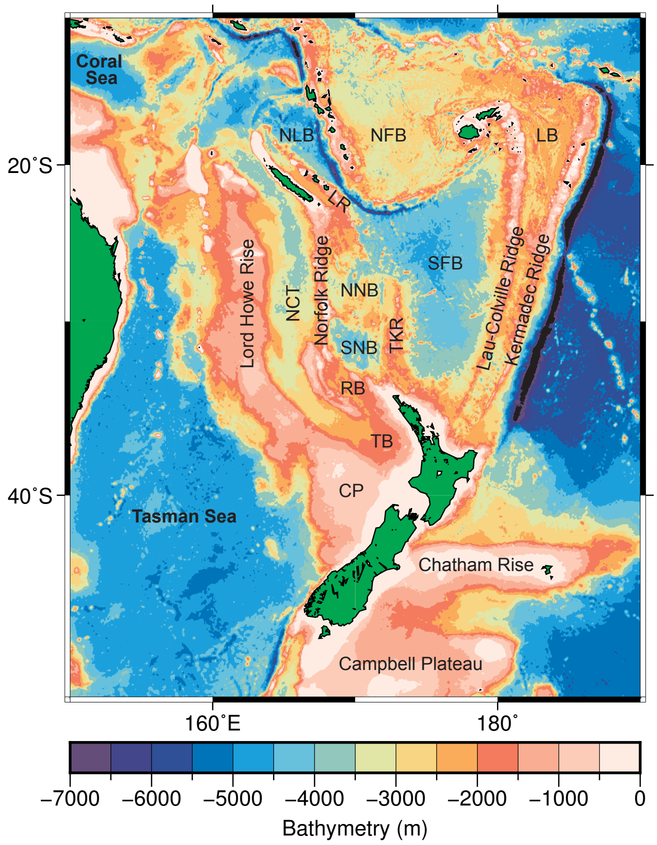


Figure 1. Bathymetry (ETOPOv2 - National Geophysical Data Center, 2006) for the southwest Pacific showing the major tectonic elements of the region. Areas above sea level are shaded green. CP, Challenger Plateau; LB, Lau Basin; LR, Loyalty Ridge; NFB, North Fiji Basin; NLB, North Loyalty Basin; NNB, North Norfolk Basin; RB, Reinga Basin; SFB, South Fiji Basin; SNB, South Norfolk Basin; TB, Taranaki Basin; TKR, Three Kings Ridge.

end of westward subduction beneath eastern Gondwanaland at 105-100 Ma until 45 Ma (Whattam et al., 2008). Seton et al. (2012) recently produced a global plate kinematic model for the past 200 million years. In the southwest Pacific their model incorporates eastward dipping subduction to close the South Loyalty Basin from 85-45 Ma after Sdrolias et al. (2003) and Sdrolias et al. (2004). Conversely, the model of Schellart et al. (2006) retains westward dipping subduction from 90 Ma until present-day. Crawford et al. (2003) and Whattam et al. (2008) incorporate westward dipping subduction until about 55 Ma, when northeast to eastward dipping subduction nucleates at the site of an extinct back-arc basin spreading centre. Steinberger et al. (2004) propose that from 83-43 Ma there was no subduction east of Australia, rather the Lord Howe Rise was attached to the Pacific plate and the Tasman Sea spreading ridge separated the Pacific Plate from the Australian plate.

In this study we test the robustness of the southwest Pacific reconstruction of Seton et al. (2012), which is embedded in a global plate kinematic model, here referred to as the EarthByte Kinematic Model 2012 (EKM12). EKM12 includes open-access time-dependent plate boundary geometries and a global rotation hierarchy. Additionally this model is the first global plate reconstruction model that incorporates the findings of Taylor (2006) that the Ontong-Java, Manihiki and Hikurangi plateaus erupted as a single large igneous province (LIP) at ~120 Ma that was subsequently rifted apart by seafloor spreading. In order to test model EKM12, we use global mantle flow models with imposed plate motions and comparisons with P- and S-wave seismic tomography. Our comparisons will provide an alternative constraint on how well plate motions are constrained and identify the current limitations of the plate model that can be addressed in future iterations. Furthermore, as model EKM12 incorporates subduction east of Australia during the period 85-45 Ma, we test Steinberger et al.'s (2004) proposal that subduction was absent at this time (83-43 Ma) and Pacific plate motion can therefore be directly tied to Australian plate motion through the Lord Howe Rise via opening of the Tasman Sea. The choice of which plate circuit to use for plate reconstructions during this timeframe has strong implications for Pacific plate motion (Steinberger et al., 2004), and the tectonic history of the southwest Pacific. The plate circuit of Steinberger et al. (2004) implies the absence of back-arc basin formation or closure during the period 83-43 Ma as both processes necessitate active subduction. Therefore, the model of Steinberger et al. (2004) requires significant modification of existing models for the evolution of the southwest Pacific which all incorporate subduction during this timeframe (Whattam et al., 2008; Schellart et al., 2006; Crawford et al., 2003; Sdrolias et al., 2003). In order to understand the geological evolution of the southwest Pacific, including the timing and driving mechanisms of back-arc basin opening and closing, it is essential to know whether subduction occurred during the late Cretaceous-early Cenozoic.

Southwest Pacific plate model

The model EKM12 includes a revived model for the evolution of the south proto-Pacific (Fig. 2). The South Loyalty Basin opens to the east of Australia as a back-arc basin from 140-120 Ma, due to eastward rollback of the long-lived west-dipping eastern Gondwanaland subduction zone. Spreading cessation at 120 Ma coincides with the eruption of the Ontong-Java-Manihiki-Hikurangi LIP and a major ridge reorganisation in the south proto-Pacific (Taylor, 2006). The eruption of the LIP leads to the initiation of a complex ridge system which rifts the plateaus apart and fragments the Phoenix plate. Fast orthogonal convergence continues until 100 Ma, when slow and very oblique convergence initiates. This coincides with a major global plate reorganisation event (Matthews et al., 2012; Veevers, 2000) that may have been triggered by events along this margin (Matthews et al., 2012). Several events occur at ~85 Ma in model EKM12. The Hikurangi plateau docks with the Chatham Rise (86 Ma) (Worthington et al., 2006; Billen and Stock, 2000), Tasman Sea spreading initiates (83.5 Ma) (Gaina et al. (1998) and east-dipping subduction initiates along the Loyalty-Three Kings Ridge (85 Ma) (Sdrolias et al., 2004). The plate model incorporates Coral Sea spreading from 61-52 Ma following Gaina et al. (1999).

For the late Cretaceous and Cenozoic model EKM12 incorporates the models of Sdrolias et al. (2003), Sdrolias et al. (2004) and DiCaprio et al. (2009). The east-dipping Loyalty-Three Kings Ridge subduction zone consumes the South Loyalty Basin until its termination at ~45 Ma when the arc collides with New Caledonia. At this time west-dipping proto-Tonga-Kermadec subduction initiates, continuing to present-day (Sdrolias et al., 2003). Clockwise eastward rollback of the west-dipping Tonga-Kermadec subduction zone is associated with opening of the North Loyalty (43-33 Ma), South Fiji (36-26 Ma), Norfolk (25-18 Ma) and Lau (12-0) basins. Contemporaneous with west-dipping Tonga-Kermadec subduction, an eastward-dipping subduction zone initiates at ~36 Ma along the western margin of the Loyalty-Three Kings ridge and is active until 18 Ma. This dual subduction

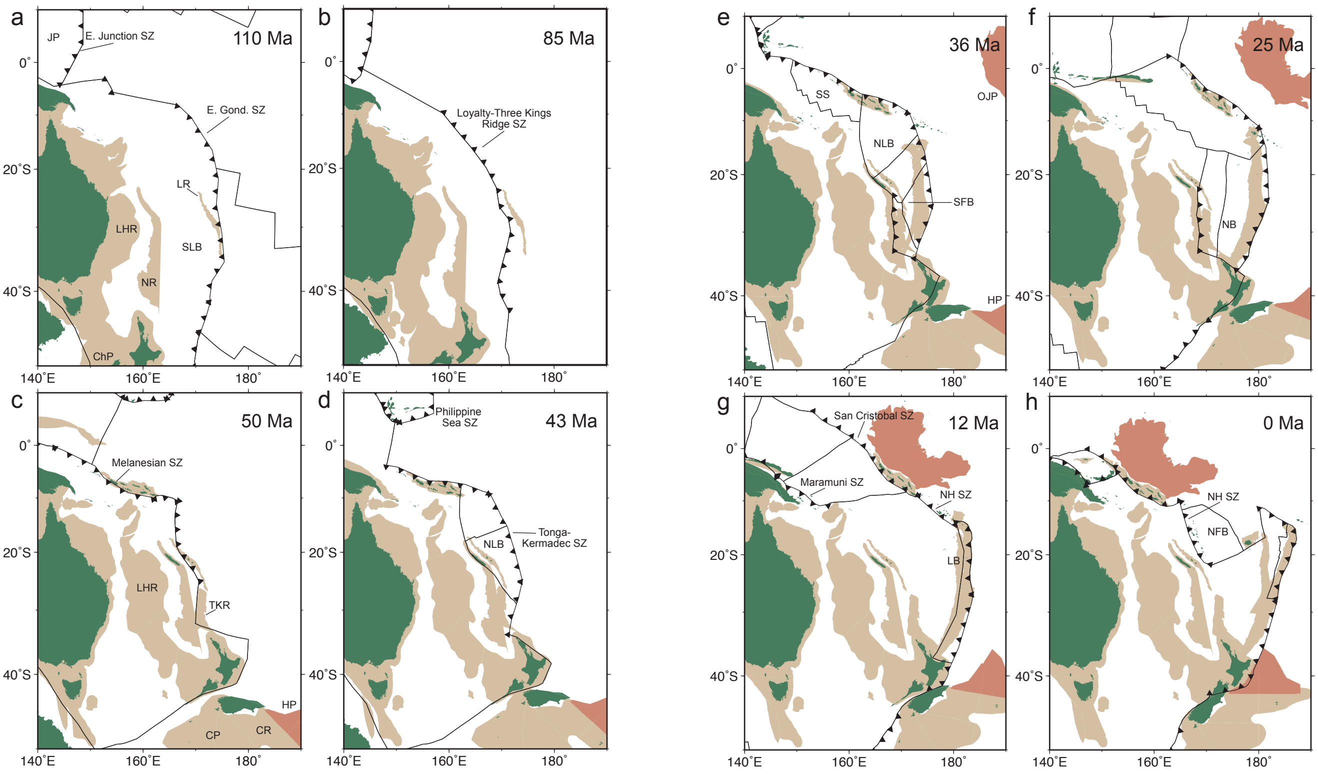


Figure 2. Plate reconstruction model of Seton et al. (2012) in a fixed Australia reference frame. Present-day continental landmasses are green, submerged continental crust and arcs are brown, and the Ontong Java plateau (OJP) and Hikurangi plateau (HP) are dark brown. SZ denotes a subduction zone. CP, Campbell Plateau; ChP, Challenger Plateau; CR, Chatham Rise; JP, Junction plate; LB, Lau Basin; LHR, Lord Howe Rise; LR, Loyalty Ridge; NB, Norfolk Basin; NFB, North Fiji Basin; NH SZ, New Hebrides subduction zone; NLB, North Loyalty Basin; NR, Norfolk Ridge; SFB, South Fiji Basin; SLB, South Loyalty Basin; SS, Solomon Sea; TKR, Three Kings Ridge.

system of opposing polarities influenced evolution of the southern South Fiji Basin and Norfolk Basin (DiCaprio et al., 2009).

During the final stages of eastward-dipping Loyalty-Three Kings ridge subduction, south-dipping Melanesian subduction initiates at 50 Ma resulting in subduction of the Pacific plate until 16 Ma. To the southwest of the Melanesian subduction zone Solomon Sea spreading takes place from 40 Ma until 16 Ma, when there is a change in the subduction configuration. At 15 Ma northeast-dipping subduction at the San Cristobal and New Hebrides trenches initiates. From 12 Ma southwest-directed rollback of the New Hebrides subduction zone consumes most of the North Loyalty Basin, resulting in formation of the North Fiji Basin which is still active at present-day (Fig. 2).

Scenario with no subduction east of Australia between 83-43 Ma

Steinberger et al. (2004) proposed a plate circuit for the period 83-43 Ma that ties the Pacific plate to Africa through Australia at the Lord Howe Rise (Africa-East Antarctica-Australia-Lord Howe Rise-Pacific). They showed that this plate circuit correctly predicts the Hawaiian-Emperor chain back to 65 Ma, reproducing the bend in the chain that occurs at ~50 Ma. According to Steinberger et al. (2004) this plate circuit may be a better alternative than determining Africa-Pacific relative motion via a plate circuit through west Antarctica (Africa-East Antarctica-Marie Byrd Land-Campbell Plateau-Pacific), since this latter circuit does not produce a bend in the Hawaii-Emperor chain, and predicts a large amount of deformation between east and west Antarctica prior to ~43 Ma. Steinberger et al.'s (2004) plate circuit implies that the Lord Howe Rise was attached to the Pacific plate during the period 83-43 Ma, and therefore that no subduction east of the Lord Howe Rise was occurring during this time interval.

Methodology

Mantle convection models driven by imposed plate motions, and fully geodynamic models driven by mantle density heterogeneities, are two methods of predicting where slabs are positioned in the mantle. These inferred slab locations can be compared with seismic tomography models of mantle structure (Ricard et al., 1993) in order to test the robustness of the plate reconstruction and evaluate alternative subduction zone histories. This methodology has been used in regional and global studies of subduction histories (e.g. Zahirovic et al, 2012; Matthews et al., 2011; Steinberger, 2000), and will be used here to test the robustness of model EKM12 and determine whether or not subduction likely occurred during the period 83-43 Ma to the east of Australia.

CitcomS model setup

We use the spherical finite element code *CitcomS* (Tan et al., 2006; Zhong et al., 2000) to compute forward numerical models of mantle convection for the past 200 Myr. Our methodology follows that used in Zahirovic et al. (2012). Plate velocities derived from model EKM12 are applied as kinematic surface boundary conditions of the model. Plate velocities are exported using the plate reconstruction software *GPlates* (Gurnis et al., 2012; Boyden et al., 2011) at 1 Myr intervals, with linear interpolation for intervening times.

The thickness of the oceanic lithosphere is assimilated in the model from reconstructed seafloor ages (Seton et al., 2012) using a half-space cooling formulation (e.g. Davis and Lister, 1974). The shallow thermal structure of the slabs is computed from the age of the subducting oceanic lithosphere using an error function temperature profile that is symmetric about the centre of the slab. The slabs are assimilated into the mantle to 350 km depth. Subduction zones that appear during the model run are progressively assimilated assuming a sinking rate of 3 cm/yr.

We have run two *CitcomS* models, referred to as SWP1 and SWP2. SWP1 uses a temperature-dependent viscosity law in which the reference viscosity of the upper mantle is 10^{21} Pa s, the transition zone (410-670 km depth) viscosity is 5×10^{21} Pa s, and the lower mantle (>670 km) viscosity is 5×10^{22} Pa s (Fig. 3). In SWP2, the upper mantle (<670 km) is assigned a temperature-independent viscosity of 10^{21} Pa s, and the lower mantle (>670 km) a temperature-independent viscosity of 5×10^{22} Pa s (Fig. 3).

The average vertical resolution is 26 km for the upper mantle, and 56 km for the lower mantle. Model parameters are listed in Table 1.

Constant variables	Values
Reference density	4000 kg/m^3
Reference viscosity	$1\text{x}10^{21}$ Pa s
Thermal diffusivity	$1\text{x}10^{-6}$ m^2/s
Coefficient of thermal expansion	$3\text{x}10^{-5}$ K^{-1}
Activation energy (upper mantle)	100 kJ/mol
Activation energy (lower mantle)	33 kJ/mol
Activation temperature	225 K
Earth radius	6371 km
Gravitational acceleration	9.81 m/s
Temperature contrast	2800 K

Table 1. Numerical modelling parameters held constant between models SWP1 and SWP2.

Seismic tomography

Mantle seismic tomography models contain information relative to the thermal structure of the mantle, and to its chemical composition. Seismic waves travel faster through colder and denser structures compared to warmer regions of the mantle, and therefore fast seismic velocities are typically associated with subducted slabs. As seismic tomography enables the identification of dense slabs in the mantle, it can be used to ground truth the final state of our forward model and validate or invalidate the proposed subduction history. We use the P-wave model MIT-P08 (Li et al., 2008) and the S-wave model GyPSum-S (Simmons et al., 2010) as images of the mantle beneath the southwest Pacific. P-wave models give high-resolution imaging of subduction zones due to the high concentration of receiver stations in proximity to seismic wave sources (earthquakes), whereas S-wave models provide better coverage of large wavelength features and ocean basins due to the sampling of broadband data (Romanowicz, 2003). S-waves

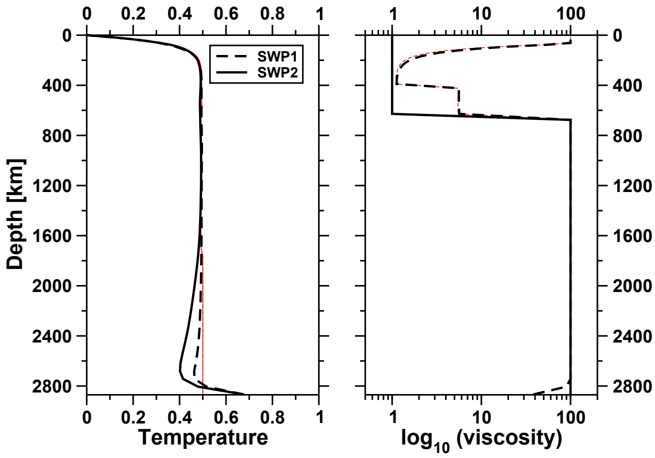


Figure 3. Radial mantle temperature and viscosity profiles for SWP1 (dashed lines) and SWP2 (solid lines). Red lines show initial profiles and black lines show present-day profiles. The *CitcomS* temperature field is non-dimensionalised with a value of 1 being equivalent to 2800 K.

are better for imaging the lower half of the mantle, for instance due to having higher amplitudes compared to P-waves (Ritsema and Allen, 2003).

We have constructed east-west trending cross-sections through the tomography models, overall perpendicular to the north-south trending subduction zones in the southwest Pacific since the late Cretaceous. The early development of the southwest Pacific back-arc basin domain occurred at higher latitudes compared to its present-day position, and therefore cross-sections have been produced at 10°S, 20°S, 30°S and 38°S in order to capture the entire region of mantle where associated slab material is expected. At 40°S the north-northeast to south-southwest trending Tonga-Kermadec subduction zone curves round to a near east-west orientation, and therefore constructing a cross-section at this latitude would image slab material from both subduction zone segments. We have chosen 38°S to facilitate more straightforward analysis of the seismic tomography and comparison with *CitcomS* temperature output. The decrease in resolution of the seismic tomography models further south, resulting from a decrease in the number of receiver stations, restricts us from presenting a robust analysis at these latitudes.

Based on the locations of model EKM12 palaeo-plate boundaries (Fig. 4), predictions can be made about where seismically fast material, associated with slabs, should be positioned in the mantle. This requires assumptions to be made about slab sinking rates and the dynamics of slab sinking. A factor of ~4 decrease in sinking rates is expected to occur as slabs penetrate the higher viscosity lower mantle (Ricard, 1993; Lithgow-Bertelloni and Richards, 1998), and slabs may not immediately penetrate the lower mantle, becoming transiently stalled at the 670 km upper-lower mantle transition zone (Fukao et al., 2001). Furthermore, the assumption that slabs sink vertically (e.g. van der Meer et al., 2010) is helpful for first-order analyses, yet it has been shown to be an oversimplification (e.g. Zahirovic et al., 2012). Slabs may be swept laterally due to the influence of mantle flow, and slab material in the mantle may also appear offset from where it was subducted at the surface due to lithospheric net rotation. Lower mantle sinking rates are on the order of 1-2 cm/yr (van der Meer et al., 2010; Schellart et al., 2009; Hafkensheid et al., 2006; van der Voo et al., 1999), and upper mantle sinking rates are larger and more variable depending on convergence rates at subduction zones.

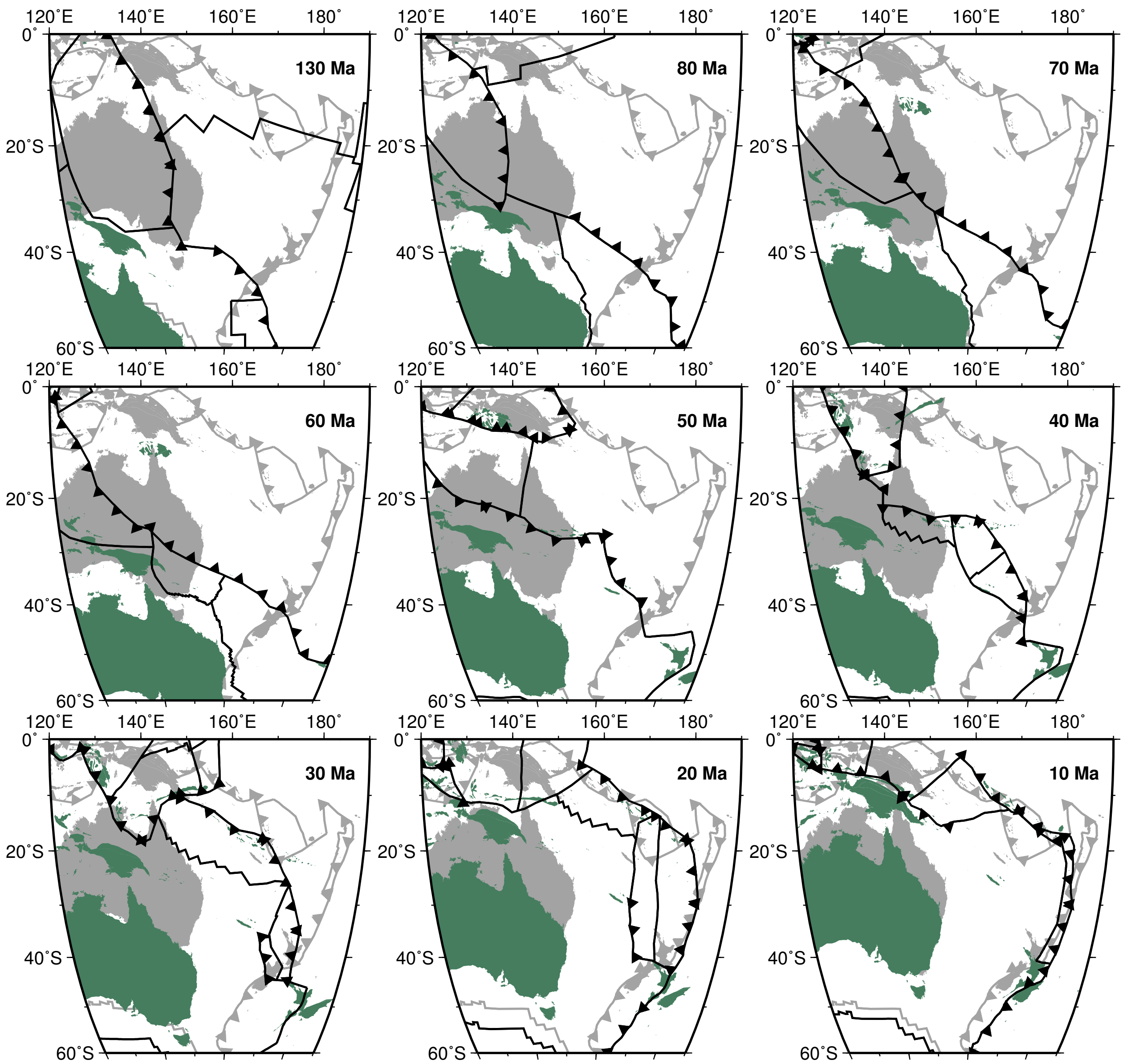


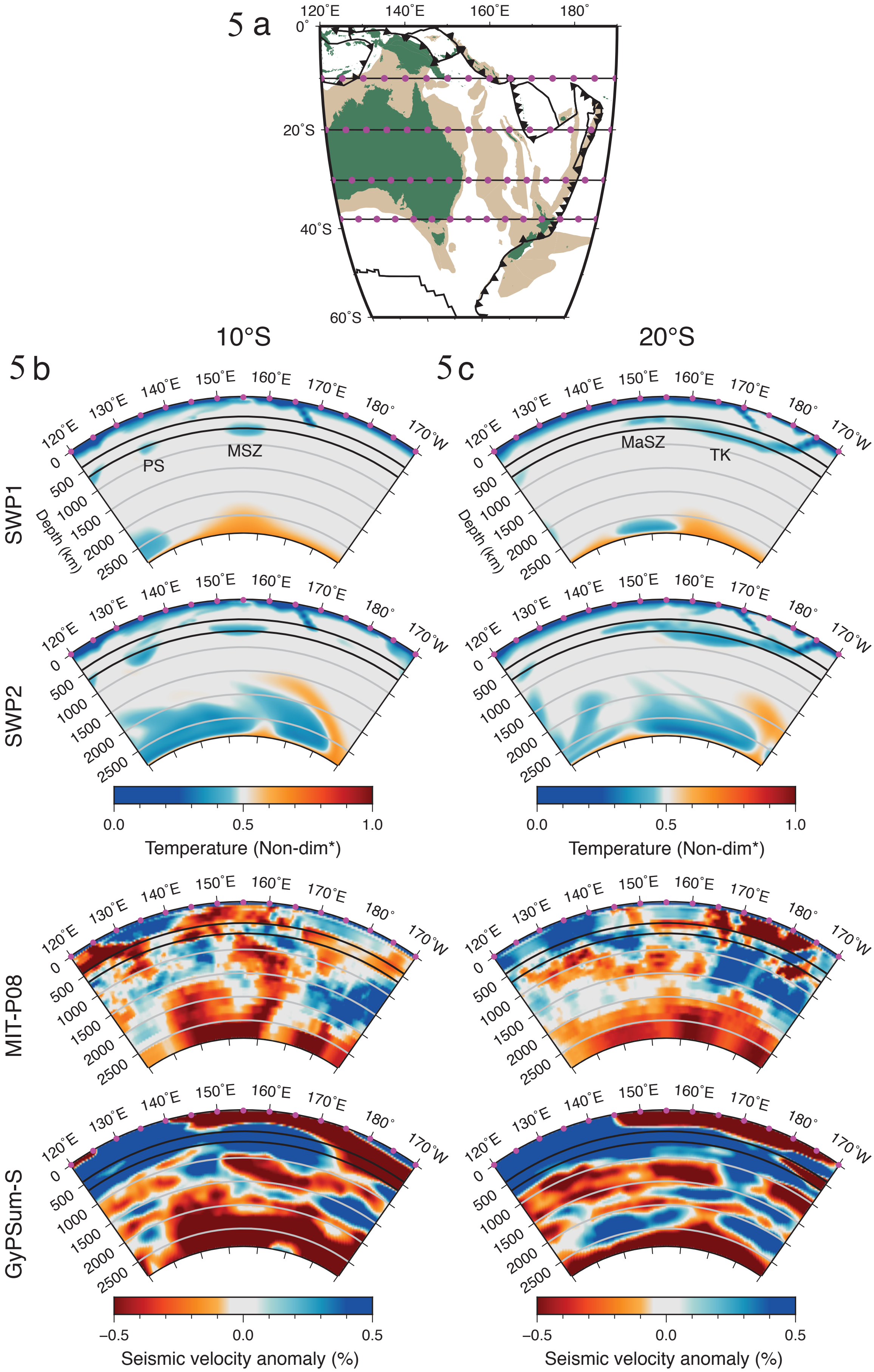
Figure 4. Plate reconstructions at selected times since the Cretaceous, in an absolute reference frame (Seton et al., 2012). Present-day plate boundaries and continents are grey, and reconstructed data are black and green. Interpreting the distribution of slab material in the mantle beneath the southwest Pacific, imaged by seismic tomography, requires knowledge of how plate boundaries have evolved in the region.

For a first-order analysis of the seismic tomography models, to determine the age of subduction of inferred slab material in the mantle, we assume an upper mantle sinking rate of 4 cm/yr and a mid-range lower mantle sinking rate of 1.5 cm/yr, and we further assume vertical sinking and immediate lower mantle penetration. A sinking rate of 4 cm/yr in the upper mantle is a conservative estimate considering that proto-Pacific plate motions were high (Müller et al., 2008), and subduction hinge rollback which is a primary mechanism driving back-arc basin formation can be very high (e.g. 8 cm/yr for the Izu-Bonin-Marianas trench since 20 Ma, and 13 cm/yr for the northern part of the Tonga-Kermadec trench since 30 Ma; Sdrolias and Müller, 2006). A mid-range lower mantle sinking rate of 1.5 cm/yr was calculated by Schellart et al. (2009) for southwest Pacific slabs, and is comparable to the global average sinking rate determined by van der Meer et al. (2010) of 1.2 cm/yr.

Results

Mantle structure inferred from seismic tomography

The seismic tomography models image a large volume of fast seismic velocity material in the mantle beneath the southwest Pacific and Australia, which may be associated with subducted oceanic slabs from palaeo-subduction zones (Figs 4,5). As a result of back-arc extension and seafloor spreading, and associated passive upwelling of warm material, slow seismic velocities dominate the upper mantle to the east of the Australian continent. Seismically fast slab material extends to the surface where subduction is presently active, but this is not resolved in each tomography model. For instance, while slabs associated with the active Tonga-Kermadec subduction zone are clearly resolved connecting to the surface (~180±10°) in each tomography model, slab material associated with present-day New Hebrides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120°E
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60°S
10°S
20°S
5 b
5 c
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SWP2
MIT-P08
GyPSum-S
Depth (km)
Temperature (Non-dim*)
Seismic velocity anomaly (%)

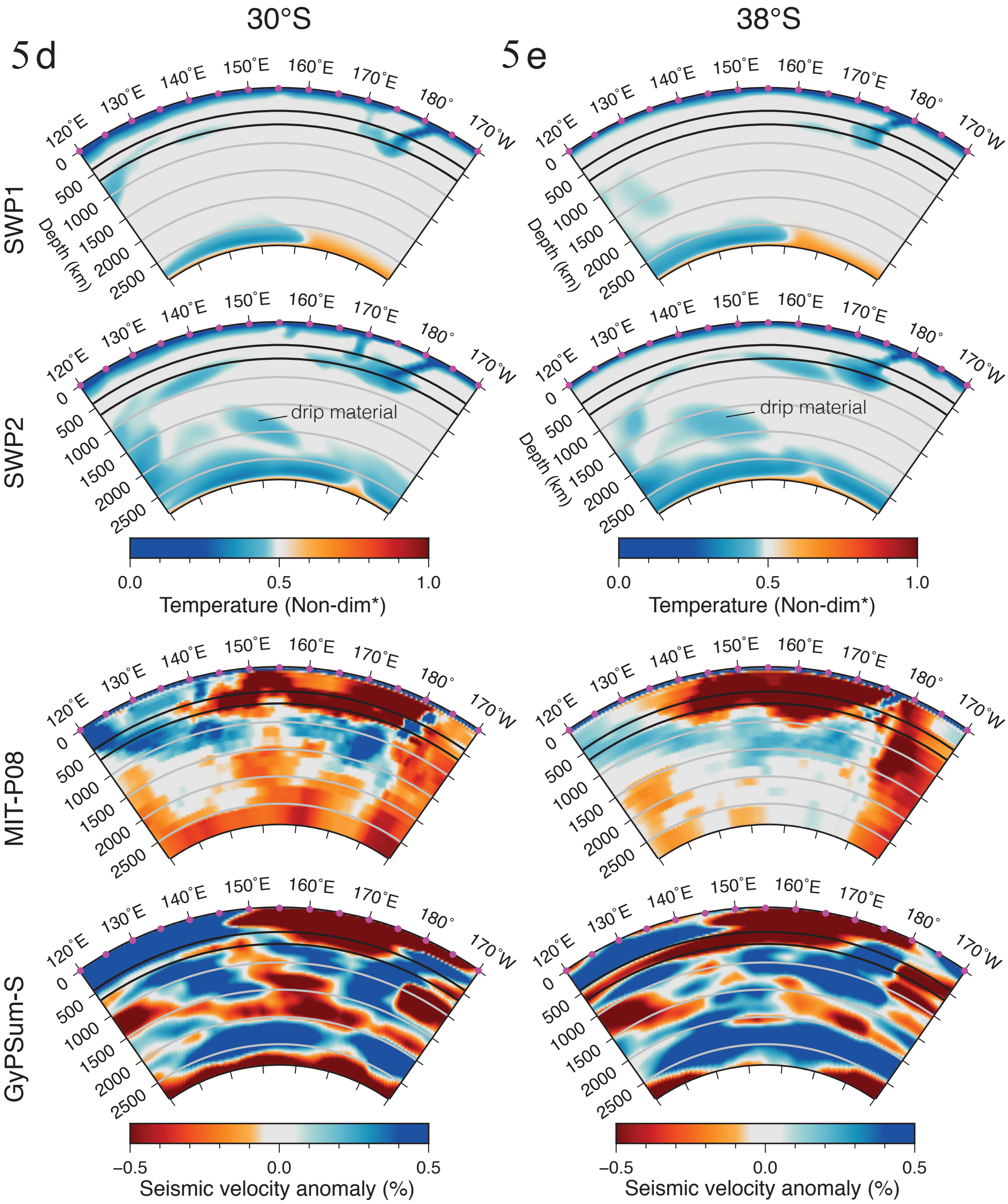

Figure 5. (a) Locations of cross-sections through the mantle temperature field of *CitcomS* models SWP1 and SWP2, and seismic tomography models MIT-P08 and GyPSum-S. (b-e) Cross-sections at 10°S, 20°S, 30°S and 38°S through the predicted present-day mantle temperature field of *CitcomS* models SWP1 and SWP2 (top), and seismic tomography (bottom). In tomography images, anomalously fast seismic velocities are inferred to represent slab material. MaSZ, slabs from the Maramuni subduction zone; MSZ, slabs from the Melanesian subduction zone; PS, slabs from Philippine Sea subduction; TK, slabs from Tonga-Kermadec subduction. *The *CitcomS* temperature field is non-dimensionalised with a value of 1 being equivalent to 2800 K.

subduction is only clearly resolved connecting to the surface in the MIT-P08 model (~170°W) and this is likely because P-wave models are better at resolving subduction zones compared to S-wave models.

4.1.1 10°S

At 10°S we expect to see slab material associated with post-Mesozoic subduction restricted to the west of 180°, where the tip of the present-day Tonga-Kermadec subduction zone is located (Fig. 4). Model EKM12 incorporates west-dipping subduction at ~140°E to the north of Australia during the Jurassic, Cretaceous and earliest Cenozoic. At this time proto-Pacific seafloor is subducting beneath eastern Gondwanaland and the Junction region that separated the Tethys and Panthalassa oceans (Seton and Müller, 2008). We expect to see slab material associated with this subduction zone extending from the core-mantle boundary to ~1300 km depth. During the Cenozoic a more complex system of

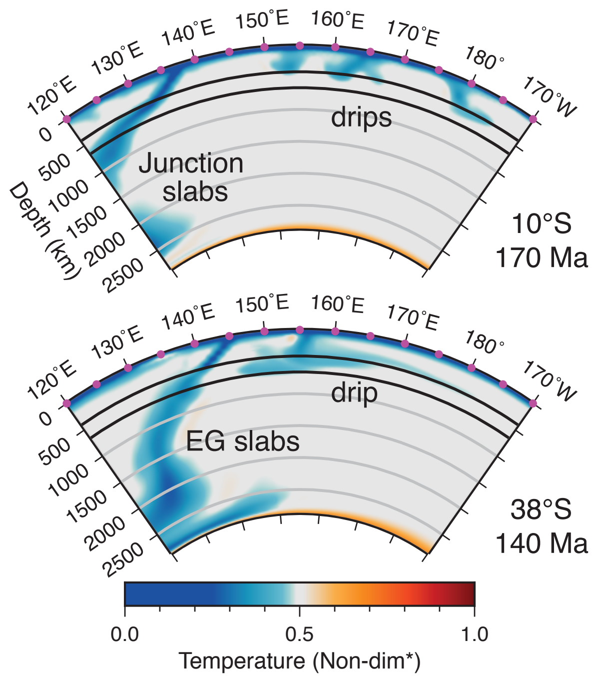

Figure 6. Cross-sections through the mantle temperature field of SWP2 at 10°S at 170 Ma (top), and 38°S at 140 Ma (bottom). Delaminating cold lithosphere can be seen in the upper mantle east of active subduction beneath the Junction region north of Australia (top) and eastern Gondwanaland (bottom).

plate boundaries develop in this region, particularly after 50 Ma, coinciding with a global plate reorganisation event that was associated with a reconfiguration of plate boundaries in the western Pacific (e.g. Whittaker et al., 2007; Sharp and Clague, 2006). Subduction at the margins of the Philippine Sea plate, and later the Caroline Sea plate, occurs between 130-150°E during the period 50-30 Ma, and therefore we expect to see slab material between ~1200-850 km depth. Subduction near 135°W continues to present day beneath Papua New Guinea. Southwest-dipping subduction at the Melanesian subduction zone in the region ~155-165°E begins around 27 Ma and continues to 16 Ma. At 15 Ma there is a polarity reversal and northeast-dipping San Cristobal subduction initiates, continuing to present-day. We predict that slab material associated with Melanesian subduction should be seen at 820-650 km depth, and slab material associated with San Cristobal subduction should extend from 650 km depth to the surface.

MIT-P08 and GyPSum-S image fast seismic velocity material west of 170°E, to the north and northeast of Australia, consistent with the expected location of slab material based on the Cenozoic evolution of the plate boundaries described above (Fig. 5). We focus on model MIT-P08 for the mid-upper mantle in this region, as at 10°S there are lots of subduction zones and P-wave models resolve subduction better than S-wave models. Between 135-150°E the fast seismic velocity material penetrates to 1500 km depth, slightly deeper than where we expect material associated with Philippine Sea subduction. Further east between 155-170°E fast seismic velocity material extends between 1000-400 km, and connects to the surface between ~160-165°E. The position of this fast seismic velocity material matches where model EKM12 predicts slab material associated with Melanesian, San Cristobal and New Hebrides subduction to be. The trend of deeper material in the west compared to the east is to be expected since subduction at the Melanesian and San Cristobal trenches initiated more recently than subduction in the Philippine and Caroline seas (Figs 2,4).

Fast seismic velocity above the core-mantle boundary west of 130°E is possibly associated with subduction beneath the Junction region north of Australia (Fig. 5). Interestingly, MIT-P08 and GyPSum-S clearly resolve a large volume of fast seismic velocity material at ~900-2000 km depth between ~170°E and 170°W, implying the presence of a long-lived subduction zone much further east of where Mesozoic-Cenozoic subduction occurs in EKM12. This material does not extend into the upper mantle. This observation implies that subduction may have been active in this area from about 100 Ma until as recently as about 30 Ma.

4.1.2 20°S

From at least 250 Ma (the oldest reconstruction time in EKM12) until 35 Ma subduction is restricted to the region west of 160°E in EKM12 (Fig. 4). West-dipping subduction beneath the Junction region to the north of Australia occurred at about 140°E until 60 Ma and therefore we expect slab material to be position in the lower mantle beneath ~1300 km depth. Since 35 Ma there has been extensive northeastward directed rollback of the Tonga-Kermadec trench from ~160°E-175°W. Assuming that the slabs did not become stalled in the transition zone we expect slab material to extend to 950 km depth. Between ~145-155°E from 15-6 Ma, southwest-dipping subduction of the Solomon Sea plate beneath Australia at the Maramuni subduction zone occurred to the north, however we may still expect to see slab material at this latitude in the transition zone. The New Hebrides subduction zone has only been active at this latitude since ~8 Ma, at ~170°E, following its southwest-directed rollback. We therefore predict slab material from the New Hebrides subduction zone to be in the upper mantle and transition zone between 540-0 km depth. Overall, based on the evolution of plate boundaries in model EKM12 we expect fast seismic velocity material to be located deeper in the mantle in the west of the study region compared to the east.

Between 120-165°E tomography model MIT-P08 shows a clear break in fast velocity material at about 1300 km depth, and GyPSum-S at 1500 km depth. This is consistent with the northward and eastward migration of subduction zones away from this region during the Cenozoic. Between 140-150°E fast seismic velocity material in the transition zone to 1000 km depth, imaged by MIT-P08 may be associated with Maramuni subduction to the north. Inferred slab material is in the expected longitudinal location, however deeper than we predict based on our assumed slab sinking rates. Beneath Australia we see fast seismic velocity material extending from 2500-1500 km in tomography model GyPSum-S, and from 2000-1300 km depth in tomography model MIT-P08. At depths greater than ~1300 km is roughly where we predict slab material associated with west-dipping Jurassic-late Cretaceous subduction of the proto-Pacific seafloor beneath the Junction plate region should be located.

There is a large amount of fast seismic velocity material in the lower mantle to the east of 165°E, more then we expect from the Tonga-Kermadec subduction zone that was only active at this latitude since about 35 Ma. As observed at 10°S, fast seismic velocity material is clearly seen in the lower mantle between 170°E-170°W penetrating to 2000 km depth according to MIT-P08, or 2500 km according to GyPSum-S. We cannot link slab material at these depths, in this location, to any subduction zones incorporated in the kinematic model tested here. This suggests that an additional subduction zone may have been located here from at least 100 Ma until about 30 Ma.

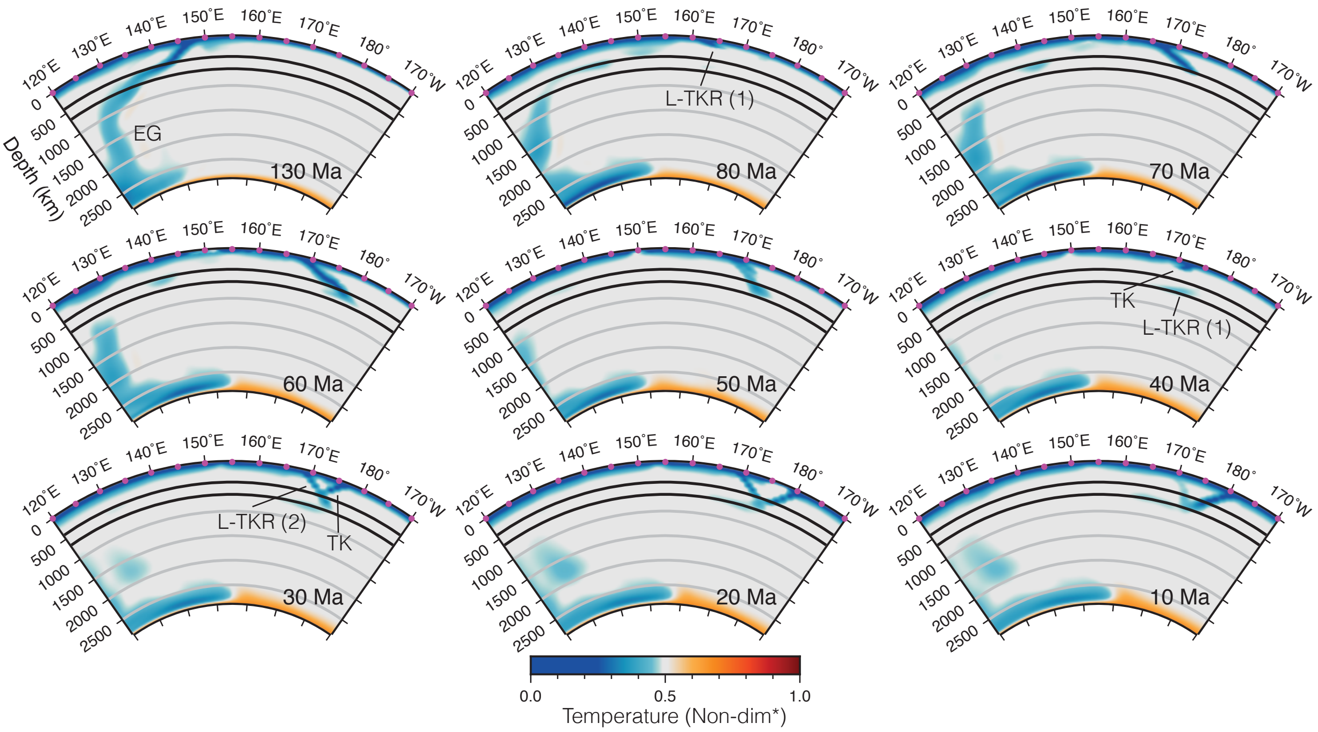

Figure 7. Time-dependent cross-sections through SWP1 mantle temperature field at 38°S. Reconstruction times corresponding to Fig. 4. It can be seen that while slab material associated with long-lived west-dipping eastern Gondwanaland subduction (EG) penetrates to the core-mantle boundary, slabs associated with shorted lived subduction zones from the late Cretaceous to present-day do not have a long residence time in the mantle resulting in a lack of slab material below ~900 km depth at present-day. L-TKR (1), slabs from Loyalty-Three Kings Ride subduction from 85-45 Ma; L-TKR (2), slabs from Loyalty-Three Kings Ridge subduction from 36-18 Ma; TK, slabs from Tonga-Kermadec subduction. *The *CitcomS* temperature field is non-dimensionalised with a value of 1 being equivalent to 2800 K.

4.1.3 30°S

Slab material associated with Jurassic to late Cretaceous aged west-dipping subduction to the north of Australia is expected, at present-day, to be positioned beneath Australia below a depth of ~1500 km as subduction occurred here continuously until 70 Ma. We also expect at least 150 km of slab material associated with south-dipping Melanesian subduction to be positioned beneath Australia at 1200 km depth. Between 160-180°E we should see evidence in the mantle north of New Zealand, above ~1700 km depth, of slab material originating from east-dipping subduction of the South Loyalty Basin from 85-45 Ma, the west-dipping Tonga-Kermadec subduction zone that initiated at 46 Ma, and the east-dipping subduction zone that facilitated opening of the Norfolk Basin from 36-18 Ma.

Both tomography models show fast seismic velocity material in the lower mantle beneath central and eastern Australia in agreement with predictions from the plate model (Fig. 5). Fast seismic velocity material is clearly resolved between ~130-160°E below the transition zone. This material may be attributed to south-dipping subduction at the Melanesian subduction zone from ~50-40 Ma, during the period when it was positioned immediately to the north at ~25-28°N. In agreement with the plate model fast seismic velocity material is also observed at 1600 km depth between 160-180°E, extending to the surface between 175-180°E where the active Tonga-Kermadec trench is located. Tomography model GyPSum-S better resolves the lower half of the mantle and comprises a continuous band of fast seismic velocity material below 1500 km in the west and 2000 km in the east. We associate this material with Jurassic to late Cretaceous aged west-dipping subduction of the proto-Pacific.

4.1.4 38°S

Based on EKM12 we expect to see fast seismic velocity material in the mantle beneath Australia and the Tasman Sea, as the eastern Gondwanaland subduction zone migrated east from ~130°E to ~160°E during the Jurassic to mid-Cretaceous. Subduction has occurred continuously at 170-180°E since 110 Ma. At 100 Ma model EKM12 incorporates a major change in convergence along eastern Gondwanaland from orthogonal to highly oblique, and it has also been suggested that subduction ended at this time (Laird and Bradshaw, 2004). If subduction ended at 100 Ma, or there was slab break-off due to a major change in convergence, then we may expect to see a break in fast seismic velocity material below slabs subducted at the Loyalty-Three Kings Ridge subduction zone after 85 Ma.

MIT-P08 has poor resolution at this latitude, apart from at the Tonga-Kermadec subduction zone, due to fewer seismic receiver stations this far south. Additionally, P-wave models produce high resolution models of subduction zones due to proximal seismic sources and receiver stations, however Australia is a tectonically stable continent and therefore much of the 38°S transect is positioned far from seismic sources. GyPSum-S resolves fast seismic velocity material along the core-mantle boundary, which we expect from eastern Gondwanaland subduction, including the east-directed rollback of the trench to accommodate opening of the South Loyalty Basin from 140-120 Ma. Between ~160-180°E there is a break in inferred slab material of about 500 km, from 2000-1500 km depth, which suggests that subduction ended at ~100 Ma (Laird and Bradshaw, 2004) or there was slab break-off due to the major change in convergence along eastern Gondwanaland from orthogonal to highly oblique at 100Ma (this is discussed further in

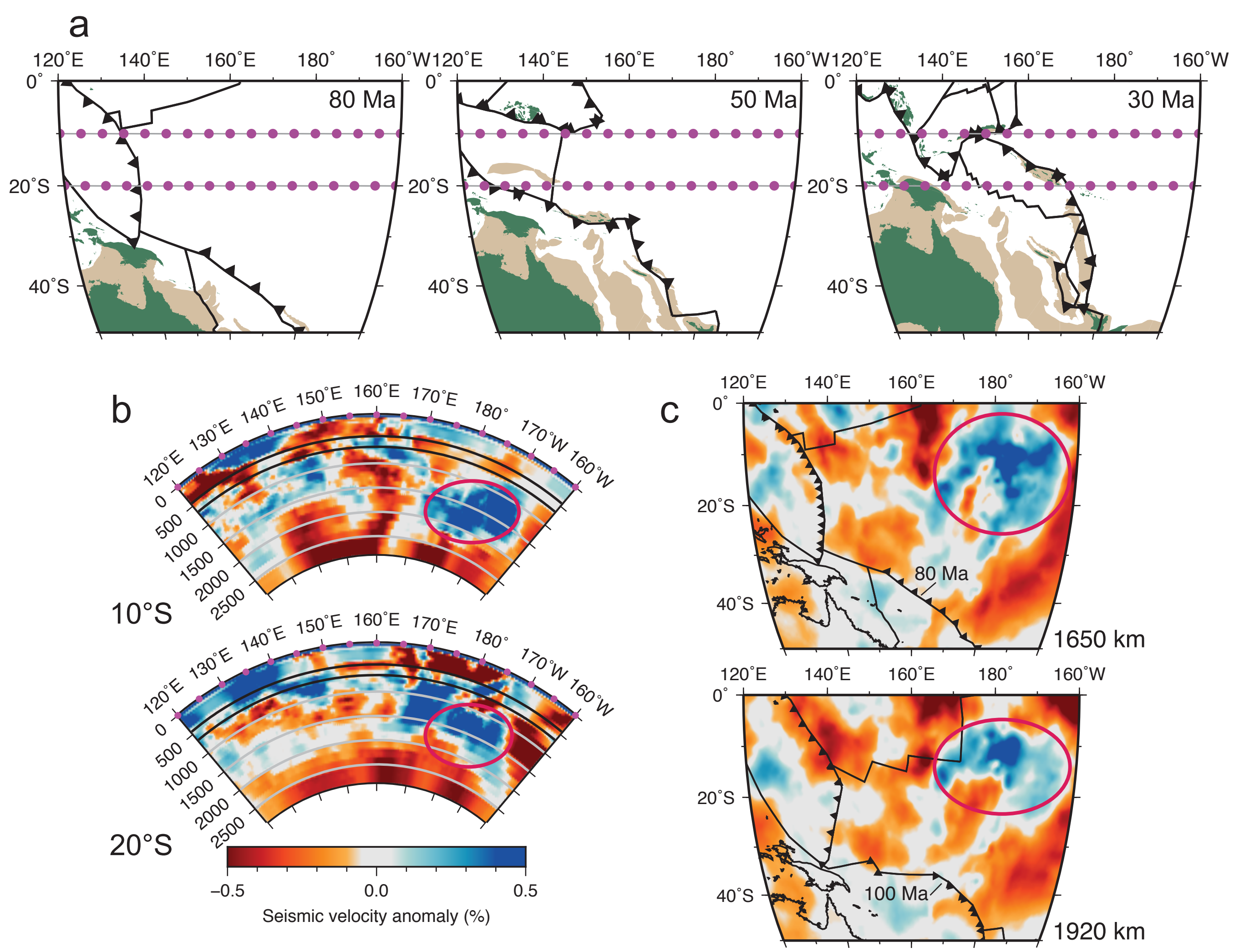

Figure 8. (a) Plate reconstructions at 80, 50 and 30 Ma (Seton et al., 2012) showing the locations of seismic tomography cross-sections shown in b-c. (b) Seismic tomography cross-sections through model MIT-P08 at 10°S and 20°S. A large volume of fast seismic velocity material in the lower mantle (denoted by pink circles) is positioned beneath a region of the southwest Pacific where subduction did not occur until about 30 Ma. (c) Horizontal seismic tomography slices (MIT-P08) at 1650 km depth (top) and 1920 km depth (bottom). Plate boundaries at 80 Ma (top) and 100 Ma, when subduction ended east of Gondwanaland (bottom), are plotted as these are the approximate ages of subduction of slab material at these depths assuming a mid-range lower mantle sinking of 1.5 cm/yr and upper mantle sinking rate of 4 cm/yr. Fast seismic velocity material aligns with the locations of subduction zones, yet a strong slab signature is also seen to the northeast ~5-25°S (pink circles).

Section 5.1). Fast seismic velocity material extends to the surface from ~1500 km depth which we associate with Loyalty-Three Kings Ridge and Tonga-Kermadec subduction.

Comparison of CitcomS model results with mantle structure inferred from seismic tomography

At present-day models SWP1 and SWP2 predict a similar distribution of slabs in the mantle to 900 km depth, although SWP2 preserves a larger volume of slab material (Fig. 5). The models strongly differ below this depth. SWP2 predicts a larger volume of slab material in the lower mantle, compared to SWP1 (Fig. 5). SWP2 additionally shows plume generation at the core-mantle boundary. The viscosity of the upper mantle is much smaller in model SWP2 in which viscosity does not depend on temperature and there is no viscosity increase at 410 km but rather a single viscosity increase at 670 km depth (Fig. 3). As a result slabs sink faster through the upper mantle in SWP2 compared with SWP1, in which slab descent is slowed at 410 km and then again at 670 km. Due to faster upper mantle sinking rates in SWP2, slab material from the long-lived eastern Gondwanaland subduction zone descends faster and does not thermally diffuse away as quickly, which results in a larger amount of cold material predicted in the lower half of the mantle. Once the slab material reaches the core-mantle boundary it moves laterally, and at 10°S we can see that it has deflected ascending buoyant plume material (Fig. 5b). One consequence of temperature-independent viscosity in model SWP2 (Fig. 3) is the Rayleigh-Taylor instability of the cold and dense oceanic lithosphere that delaminates in the upper mantle (Fig. 6). This delamination is most important in the early stages of the model run when the oceanic lithosphere to the north and east of eastern Gondwanaland is oldest. This behaviour results in the formation of unrealistic cold "drips" (Fig. 6), and therefore we focus on the results of SWP1 (Fig. 7). Note that delaminated lithosphere is positioned in the lower mantle at present-day (Fig. 5d-e) and should be disregarded when comparing the model output with seismic tomography images.

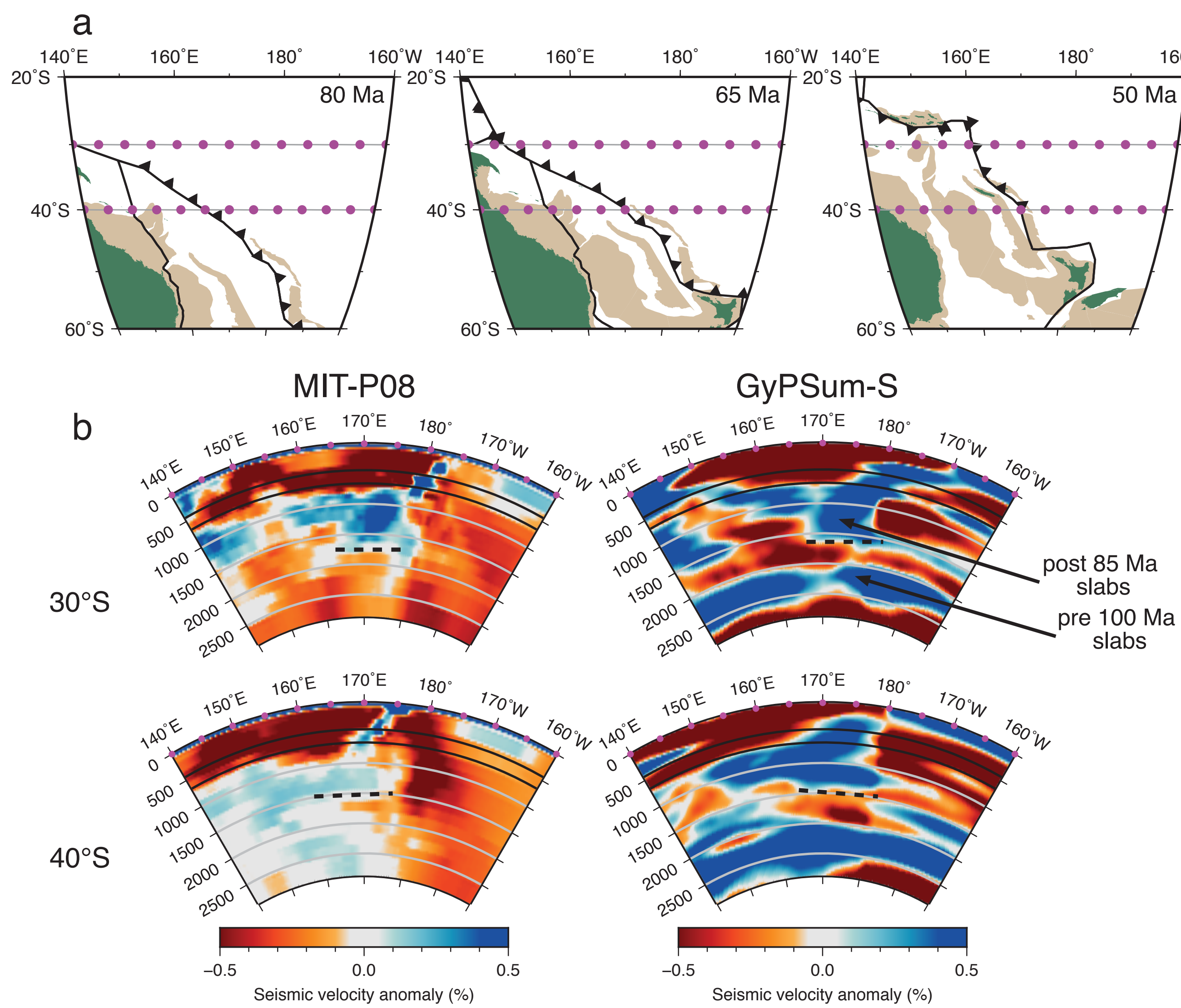

Figure 9. (a) Plate reconstructions at 80, 65 and 50 Ma (Seton et al., 2012) showing the locations of seismic tomography cross-sections shown in b. At these times there was spreading in the Tasman Sea, and Steinberger et al. (2004) propose that there was no subduction east of Australia. (b) Seismic tomography cross-sections at 30°S and 40°S. At these latitudes model EKM12 predicts that there should be slab material in the lower mantle above ~1450 km depth assuming an upper mantle sinking rate of 2 cm/yr and a lower mantle sinking rate of 1.5 cm/yr (see text for more details). Below this depth the plate model predicts that there should be a break in fast seismic velocity material to account for the preceding period where there was no subduction after eastern Gondwanaland subduction ceased at ~100 Ma. Dashed line indicates where we propose slab material associated with post-85 Ma subduction is located.

CitcomS models SWP1 and SWP2 produce a good match with seismic tomography in the upper mantle and transition zone. For instance at 10°S between 150-170°E (Fig. 5b), cold material appears to be penetrating the lower mantle between 500-700 km depth. This material was subducted at the southwest-dipping Melanesian subduction zone during its rollback. Fast seismic velocity material is observed in the tomography models in this same region, although focused between 155-170°E and penetrating deeper into the mantle over a wider area. Further to the west, at 130-135°E, slab material at 550-900 km depth was produced by Philippine Sea and Caroline Sea subduction. Tomography models also reveal fast seismic velocity material in this region, however to a larger longitudinal (135-150°E) and radial extent (1500-670 km). At 20°S slab material subducted at the Tonga-Kermadec trench during extensive eastward directed rollback initially became stalled in the transition zone between 155-180°E (Fig. 5c). As a consequence of rollback the cold material predicted by the models is flat lying and appears to have begun sinking into the lower mantle. The additional slab material in the transition zone between 145-155°E originates from southwestward subduction of the Soloman Sea plate beneath the Australian plate at the Maramuni trench from ~15-5 Ma. GyPSum-S images fast seismic velocity material in the transition zone and upper most lower mantle across most of the study area from 120-180°. This may be attributed, in part, to slab material from the Tonga-Kermadec and Maramuni trenches, however as the signal is constant for the entire region there is likely a resolution problem. MIT-P08 also shows fast seismic velocity material in the transition zone and lower mantle, yet it is less continuous than GyPSum-S. Fast seismic velocity material matches where SWP1 and SWP2 show Maramuni slabs, however MIT-P08 images a larger volume of inferred slab material extending to deeper depths. The volume of fast seismic velocity material imaged by MIT-P08 to the east of 160°E it very large suggesting that it originates from more than just Tonga-Kermadec subduction, as discussed above (Sections 4.1.1-4.1.2), which predicts slab material to 950 km. As SWP1 and SWP2 do not incorporate subduction at 20°S prior to 35 Ma, we do not expect slab material to be located beneath about 950 km depth.

Further south at 30°S and 38°S, SWP1 and SWP2 show slab material from the dual Tonga-Kermadec and Loyalty-Three Kings Ridge subduction zones is positioned in the transition zone and uppermost lower mantle. At 30°S where there has been comparatively more eastward rollback of the Tonga-Kermadec trench, compared to further south, and therefore modelled slab material covers a wider area. P-wave models produce high-resolution models of subduction zones (Romanowicz, 2003) and only MIT-P08 images fast seismic velocity material associated with present-day Tonga-Kermadec subduction. GyPSum-S does not resolve slabs connecting to the surface at these latitudes and therefore we focus on MIT-P08 results for our analysis of the upper mantle and transition zone. MIT-P08 upper mantle Tonga-Kermadec slabs match those in the SWP1 model, yet there is less fast seismic velocity material in the transition zone compared to SWP1 and SWP2 (Fig. 5d). Particularly at 30°S there is no fast seismic velocity material in the transition zone between 170-178°S as seen in the SWP1 and SWP2 results, rather it sits just below the 670 km transition zone boundary having clearly penetrated the lower mantle.

SWP1 and SWP2 produce a poor match with tomographic fast seismic velocity trends in the lower mantle. SWP1 does not preserve enough slab material in the lower half of the mantle, while SWP2 preserves too much slab material above the core-mantle boundary at 10 and 20°S. In some regions we do not expect lower mantle fast seismic velocity material due to the absence of subduction. In the lower mantle to the east of 170°E a large volume of fast seismic velocity material is resolved by the tomography models at 10-20°S. Model EKM12 does not incorporate subduction in this region during the Jurassic to mid-Cenozoic (Fig. 4) and therefore SWP1 and SWP2 do not predict slab material in this location. In other regions it appears that SWP1 and SWP2 fail to produce the amount of lower mantle slab material that is predicted by the seismic tomography models. This could be due to the viscosity structure used in the model. A smaller viscosity jump across the upper/lower mantle boundary would result in greater amounts of cold material preserved in the lower mantle. SWP1 slab material is also heating up to ambient mantle temperature too quickly, which reflects slow sinking rates.

Discussion

Based on a qualitative comparison between the plate boundary evolution and the seismic tomography alone, the position of fast seismic velocity material in the upper and lower mantle matches where we expect it to be located based on the subduction history of the kinematic model. The *CitcomS* derived temperature field for the upper mantle and upper-most lower mantle, to a depth of ~900 km, produces a close match with seismic tomography (Fig. 5). The latitudinal and longitudinal location of slab material subducted since about 50 Ma matches that of fast seismic velocity material imaged by tomography, particularly the MIT-P08 tomography model which better resolves subduction zones and the upper mantle compared to GyPSum-S. In addition, the modelled slab material at the core-mantle boundary also matches both seismic tomography models, particularly GyPSum-S which better resolves the lower mantle compared to MIT-P08, suggesting that the location of the long-lived west-dipping eastern Gondwanaland subduction zone in model EKM12 is well constrained. However, the volume of slab material throughout the lower mantle inferred from the tomography is larger than what we see in the SWP1 and SWP2 results, which is a function of the plate model and of our choice of modelling parameters. From a plate reconstruction perspective, the timing of initiation of the multiple subduction zones in the area and the age of the reconstructed ocean floor are ill constrained. Both parameters are critical, because ephemeral subduction zones of young ocean floor will not get advected to depth. Fig. 7 clearly shows that long-lived subduction zones result in slabs with longer residence time in the lower mantle than ephemeral subduction zones. From a modelling perspective, increasing the assumed sinking rate for the early stages of new subduction zones, reducing the viscosity jump across the upper/lower mantle boundary, and using a depth-dependent reference viscosity in the lower mantle (e.g. Gurnis et al., 2000) will increase the volume of material entering the mantle and increase its residence time. The overall good match between mantle tomography and EKM12 makes the southwest Pacific a key area to investigate the viscosity structure of the mantle.

A large volume of fast seismic velocity material in the mid-lower mantle at ~900-2000 km depth is located between ~5-25°S at the northern end of present-day Tonga-Kermadec subduction, and to the east of 170°E (Fig. 8). Both P- and S-wave tomography resolve this inferred slab material (Fig. 5b-c), strongly suggesting the presence of long-lived subduction in this region. In EKM12 this region does not have a history of subduction until about 30 Ma when east directed roll-back of the Tonga-Kermadec trench introduces subduction to this latitude (Figs 4,8). Therefore the tomographically imaged slab material in this region has important implications for Panthalassic plate margin evolution. In EKM12 a single subduction zone accommodates subduction of Panthalassa beneath eastern Gondwanaland during the Jurassic to mid-Cretaceous. Based on comparing mantle convection modelling results to seismic tomography we suggest that there may have been an intra-oceanic subduction zone active for several tens of millions of years, due to the radial extent of the fast seismic velocity material imaged by tomography. We suggest the subduction zone was located at ~180-170°W. This subduction zone would have been located ocean-ward of the Junction plate region and therefore material associated with a remnant volcanic arc would likely be positioned in southeast Asia in the Philippines or north of Australia. This finding has further implications for deciphering the history of accreted terranes in southeast Asia, and future work should involve locating potential arc remnants.

Van der Meer et al. (2010) suggested that a linear north-south trending band of fast seismic velocity material located at 1325-920 km depth, in the P-wave tomography model of Amaru (2007), may be associated with a north-south trending subduction zone at ~175°E between ~15-35°S. The tomography model they analysed did not resolve the large volume of slab material to the east of 180° that is resolved by MIT-P08 and GyPSum-S.

Was there subduction east of Australia during the late Cretaceous to early Cenozoic?

Steinberger et al. (2004) proposed that there was no subduction east of Australia during the period 83-43 Ma. As a result the Lord Howe Rise would be part of the Pacific plate with Pacific plate motion tied to Africa through Australia, via spreading in the Tasman Sea. This is contrary to many kinematic studies of the southwest Pacific (Whattam et al., 2008; Schellart et al., 2006; Crawford et al., 2003; Sdrolias et al., 2003). Model EKM12 incorporates subduction of the South Loyalty Basin at the east-dipping Loyalty-Three Kings Ridge subduction zone from 85-45 Ma (Fig. 9a).

During this period the back-arc basin was completely subducted. In their model the South Loyalty Basin reached a maximum width of approximately 800 km, which implies a convergence rate of around 2 cm/yr at the subduction zone. If we further assume (1) a lower-mantle sinking rate of 1.5 cm/yr, and (2) subduction has been continuous from 85 Ma until present day between 30-40°S, then we expect to see a continuous volume of slab material in the mantle above 1450 km depth. At 105-100 Ma it has been proposed that subduction of Panthalassa beneath eastern Gondwanaland ceased (Laird and Bradshaw, 2004), accompanied by a change to strike-slip motion at the margin (Veevers, 2000), and a slab breakoff event (e.g. Tappenden, 2003). EKM12 incorporates a major change in the nature of convergence along the eastern Gondwanaland margin from orthogonal to highly oblique at 100 Ma. This major change in relative motion between the proto-Pacific and eastern Gondwanaland may be represented by a break in slab material in the lower mantle, preceding the initiation east-dipping subduction of the South Loyalty Basin at 85 Ma.

At 30 and 40° S the GyPSum-S tomography model images a continuous volume of fast seismic velocity material extending from 1600 km and 1500 km depth to the upper mantle, respectively, below which there is a break in inferred slab material to 2000 km depth (Fig. 9b). Model MIT-P08 only images fast seismic velocity material extending from ~1750 km depth to the surface at 30°S, with no fast seismic velocity material at the core-mantle boundary. At 40° fast seismic velocity material extends from the surface to 1600 km depth, below which velocity anomalies are very weak, i.e. close to 0% (Fig. 9b). These observations are consistent with the subduction history inferred from model EKM12 and other studies of eastern Gondwanaland evolution during the mid-Cretaceous (Laird and Bradshaw, 2004; Tappenden, 2003; Veevers, 2000). Subduction occurred after 85 Ma and is in fact necessary to account for the large volume of fast seismic velocity material in the lower mantle. Schellart et al. (2006) showed that based on their reconstruction for the southwest Pacific there was a small amount of convergence in the New Zealand region during the period 83-43 Ma, which is supported by geological evidence. We support their conclusion that the Lord Howe Rise was not part of the Pacific plate during this timeframe.

Conclusions

The southwest Pacific is a frontier region for petroleum exploration and several studies indicate thick sedimentary sequences and petroleum potential in a number of basins. Sparse data coverage highlights the value of regional plate tectonic and geodynamic models based on constraints from surface geology and geophysics as well as mantle tomography. We have analysed P- and S-wave seismic tomography images, and incorporated the plate motions of Seton et al. (2012) into global mantle flow models (‘SWP1’ and ‘SWP2’) using the numerical code *CitcomS*. SWP1 produces a good match with seismic tomography from the surface to 900 km depth and at the core-mantle boundary in terms of the latitudinal and longitudinal location of slab material. However, the residence time of ephemeral subduction zones is too short in this particular model, which results in a poor match to mantle tomography. This implies that the timing of initiation of subduction zones and the age of subducting oceanic lithosphere may have been underestimated in model EKM12.

An unexpected result of our investigation is that we have found evidence for a long-lived intra-oceanic subduction zone to the northeast of Australia between about 10-25°S and 170°E-170°W. Well defined fast seismic velocity material in this region extends from ~900-2000 km. Assuming an upper mantle sinking rate of 4 cm/yr and a lower mantle sinking rate of 1.5 cm/yr the subduction zone was active from about 100-35 Ma, and was located outboard of the Junction plate that separated the Tethys and Panthalassa oceans (Seton and Müller, 2008). We suggest that arc material associated with this subduction zone is likely preserved in the southeast Asia region, due to the overall westward directed subduction of proto-Pacific seafloor during the Cretaceous and Cenozoic seen in model EKM12.

Steinberger et al. (2004) proposed that from 83-43 Ma the Lord Howe Rise may have been attached to the Pacific plate, and that Pacific plate motion can be tied to Africa via seafloor spreading between Australia and the Lord Howe Rise. Their plate circuit implies that there was no subduction to the east of Australia during this 83-43 Ma timeframe, contrary to most plate reconstruction models of the southwest Pacific (Whattam et al., 2008; Schellart et al., 2006; Crawford et al., 2003; Sdrolias et al., 2003). We find that the distribution of slab material in the mantle inferred from seismic tomography matches the subduction history of Seton et al.’s (2012) EKM12 plate model, which includes east-dipping subduction east of the Lord Howe Rise from 85 to 45 Ma. Mantle tomography images show that a continuous volume of slab material extends from the surface to ~1500 km depth which is the depth where we predict 85 Ma slabs to be located. Below this depth there is a break in inferred slab material which we associate with the major change in relative plate motion along eastern Gondwanaland at 100 Ma, which may have been associated with subduction termination (Laird and Bradshaw, 2004) and slab-breakoff (Tappenden, 2003). We therefore suggest that the Lord Howe Rise was not attached to the Pacific during this time period, and Pacific plate motion cannot be tied to Africa via the Lord Howe Rise and Australia, as was proposed by Steinberger et al. (2004).

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