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Geologic and kinematic constraints on Late Cretaceous to mid Eocene plate boundaries in the southwest Pacific

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A B S T R A C T

Starkly contrasting tectonic reconstructions have been proposed for the Late Cretaceous to mid Eocene (~85–45 Ma) evolution of the southwest Pacific, reflecting sparse and ambiguous data. Furthermore, uncertainty in the timing of and motion at plate boundaries in the region has led to controversy around how to implement a robust southwest Pacific plate circuit. It is agreed that the southwest Pacific comprised three spreading ridges during this time: in the Southeast Indian Ocean, Tasman Sea and Amundsen Sea. However, one and possibly two other plate boundaries also accommodated relative plate motions: in the West Antarctic Rift System (WARS) and between the Lord Howe Rise (LHR) and Pacific. Relevant geologic and kinematic data from the region are reviewed to better constrain its plate motion history during this period, and determine the time-dependent evolution of the southwest Pacific regional plate circuit. A model of (1) west-dipping subduction and basin opening to the east of the LHR from 85–55 Ma, and (2) initiation of northeast-dipping subduction and basin closure east of New Caledonia at ~55 Ma is supported. West-dipping subduction and basin opening were not driven by convergence, as has previously been proposed. Our plate circuit analysis suggests that between at least 74 Ma and subduction initiation at ~55 Ma there was little net relative motion between the Pacific plate and LHR, <20 km of convergence with a component of strike-slip motion. Subduction must therefore have been primarily driven by the negative buoyancy of the slab, or perhaps forced trench retreat due to orogenic collapse. We propose that at least two plate boundaries separated the Pacific plate from the LHR during this time, however, as there was little to no motion between these plates then a plate circuit which treats the Pacific plate and LHR as a single plate ("Australian" circuit) will produce similar kinematic results to a circuit which leaves their relative motion unconstrained and treats them as separate plates ("Antarctic" circuit). Prior to 74 Ma the reliability of magnetic anomalies from southwest Pacific spreading systems is questionable and it is difficult to properly test alternative plate circuits. After 55 Ma we advocate using an Antarctic plate circuit as the Australian plate circuit models that were tested predict significant net compression in the WARS, for which evidence is absent. Our preferred model makes testable predictions, such as burial of an arc beneath the Tonga and Vitiaz ridges, and Late Cretaceous to Eocene slabs in the mantle beneath the southwest Pacific, both of which can be investigated by future work. These predictions are particularly important for testing the earlier 85–55 Ma phase of the model, which is largely underpinned by ages and interpretations of South Loyalty Basin crust obducted onto New Caledonia, rather than an extinct arc or arc-related rocks.

Contents

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1. Introduction

The southwest Pacific (Fig. 1) has had a complex tectonic history since final Gondwanaland dispersal began, dominated by multiple episodes of marginal and back-arc basin formation, subduction and trench rollback (e.g. Crawford et al., 2003; Sdrolias et al., 2003; Schellart et al., 2006; Whattam et al., 2008; Cluzel et al., 2012a,b). Understanding the history of this complicated plate boundary activity in the southwest Pacific is of regional and global significance, including for understanding basin subsidence and hydrocarbon formation along the Lord Howe Rise (LHR) and around New Zealand, understanding the mechanisms accommodating Tasman Sea spreading (e.g. Schellart et al., 2006), and modeling mantle plumes within global plate kinematic models. However, less than 10% of the continental crust now in the South Pacific that rifted away from Australia and Antarctica is presently sub-aerially exposed and readily accessible for field exploration (Mortimer, 2008). Sub-aerial exposures are largely complicated by Cenozoic crustal thickening events in New Zealand and New Caledonia. Furthermore, data coverage in several offshore locations (e.g. Coral Sea) is sparse, and partial or complete basin subduction has destroyed large swaths of ocean crust. The magnetic isochrons created at the Tasman, Amundsen and Bellinghausen ridge systems provide the earliest robust constraints on the position of the Pacific plate relative to the rest of the global plate network — prior to this time the Pacific was completely surrounded by subduction zones (Seton et al., 2012). However, while spreading in the Tasman, Amundsen and Bellingshausen seas is reasonably well-constrained, alternative models exist for the early spreading history between Australia and Antarctica (Royer and Rollet, 1997; Tikku and Cande, 1999, 2000; Whittaker et al., 2007, 2013). Together, these data gaps have made it difficult to build well-constrained regional plate reconstruction models, and there remain many unresolved and hotly debated problems relating to southwest Pacific evolution. Questions remain concerning the timing, location and polarity of different subduction episodes, the driving mechanism for obduction events in New Caledonia and New Zealand, the mechanism for Tasman Sea opening and widespread rifting in Zealandia (Fig. 1), seafloor ages and the orientation of spreading in various basins (e.g. d'Entrecasteaux Basin), and the origin of several submerged tectonic features (e.g. South Rennell Trough). Conversely, the existence of Tonga–Kermadec subduction (Bloomer et al., 1995) and plate boundary activity within New Zealand (Sutherland, 1995) from at least 45 Ma is well established. Many of the unresolved issues related to southwest Pacific evolution are related to the poorly understood timing and type of plate boundary activity to the east of the LHR during the Late Cretaceous to mid Eocene. This problem will, therefore, be the focus of our study.

A robust reconstruction model for the southwest Pacific has wider implications for global geodynamic studies. Plates in the Pacific realm

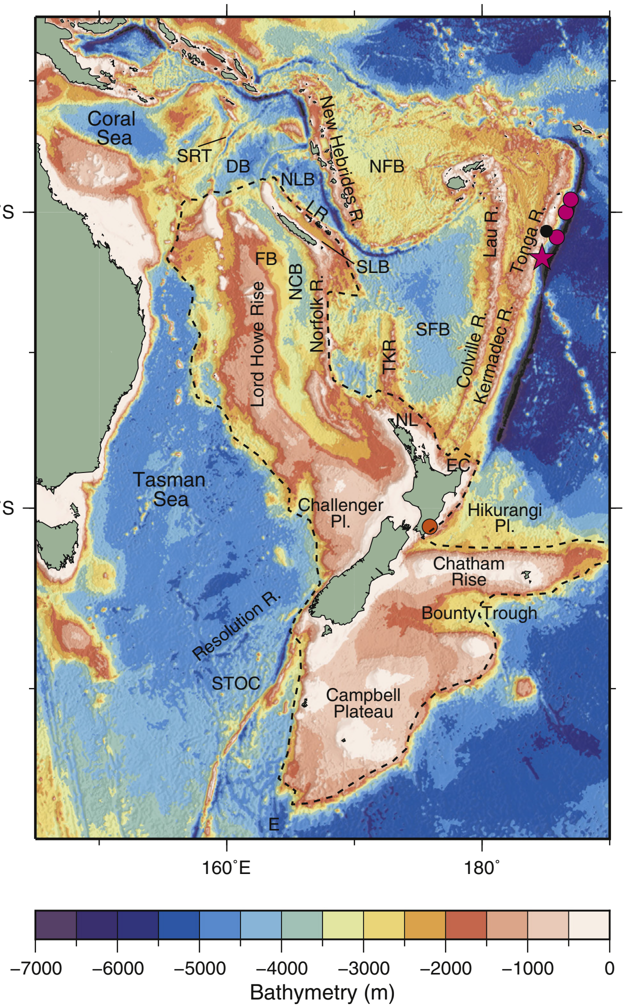

Fig. 1. Southwest Pacific bathymetry (ETOPO1, Amante and Eakins, 2009) showing major tectonic and elements of the region. Areas above present-day sealevel are shaded green and Zealandia is outlined with a black dashed line (Mortimer, 2008). DB, d'Entrecasteaux Basin; E, Emerald Basin; EC, East Cape; FB, Fairway Basin; LR, Loyalty Ridge; NCB, New Caledonia Basin; NFB, North Fiji Basin; NL, Northland; NLB, North Loyalty Basin; SFB, South Fiji Basin; SLB, South Loyalty Basin; SRT, South Rennell Trough; STOC, Southeast Tasman oceanic crust; TKR, Three Kings Ridge. Black filled circle shows the location of the Tongan island of 'Eua. Pink filled circles show the locations of dredge sites of Meffre et al. (2012) referred to in this paper. Pink star shows the location of ODP site 841B. Orange filled circle shows the location of the Kopi Boninite of Moore (1980). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

can only be linked with the motion of plates in the Indo-Atlantic realm via a southwest Pacific spreading corridor. The ability to link the Pacific plate to a relative plate motion circuit connected to the Indo-Atlantic realm enables tighter constraints to be placed on its relative and absolute motion history, and that of other plates in the Pacific ocean basin, such as the Kula and Farallon plates. This in turn affects computation of subduction budgets around circum-Pacific active margins. Predictions of the angle of convergence and amount of slab material subducted in the western, northern and eastern Pacific are influenced by how the Pacific plate is reconstructed (Sutherland, 2008). Furthermore, determining if and how Pacific hotspots have moved relative to Indo-Atlantic hot spots is a long-standing problem in plate tectonics and is of paramount importance for constructing absolute plate motion models (e.g. Tarduno and Gee, 1995). Global models of absolute plate motions based on the time-progressive volcanism along hot spot trails (Steinberger et al., 2004; Cuffaro and Doglioni, 2007; Doubrovine et al., 2012) rely on accurately linking observations between the Indo-Atlantic and Pacific realms. Steinberger et al. (2004) proposed that the ~50 Ma bend in the Hawaiian–Emperor seamount chain is best predicted using a plate model assuming no plate boundary between the Pacific and the LHR before chron 20 (43 Ma - timescale of Gee and Kent, 2007 is adopted). However, the assumptions within this plate-circuit are controversial (Schellart et al., 2006; Tarduno et al., 2009). A better understanding of the history of west versus east-dipping subduction in the southwest Pacific and how these intervals are related to episodes of opening versus closure of back-arc basins may contribute to our understanding of the effects of geographic polarity on subduction kinematics (Doglioni et al., 2007).

To better constrain the broad scale tectonic evolution of the southwest Pacific during the Late Cretaceous to mid Eocene period, and so provide a better understanding of how the Pacific plate should be reconstructed with respect to a relative plate motion chain, we review geologic and kinematic data from the southwest and southern Pacific. Our paper is divided into two parts. In Part A we review geologic data from the southwest Pacific and West Antarctic Rift System to better constrain the timing, types and locations of plate boundaries that were active during the Late Cretaceous to mid Eocene. In Part B we investigate the kinematic consequences of adopting alternative Pacific plate motion circuits for motion in the West Antarctic Rift System and between the LHR and Pacific, where relative motion histories are poorly constrained. This process allows us to identify which plate circuits produce realistic or unrealistic amounts of motion and implied deformation, based on a comparison with geologic observations and documented histories of relative motion and quiescence.

1.1. Tectonic setting and previous reconstruction models

It is well established that subduction along the eastern margin of Gondwanaland was long-lived. Subduction related magmatism preserved in the New England Fold Belt (Leitch, 1975; McPhie, 1987) is evidence that a convergent margin paralleled Eastern Gondwanaland from at least the Carboniferous. Arc related products are also preserved in the Téremba Terrane in New Caledonia (Paris, 1981; Campbell, 1984; Adams et al., 2009) and the Murihiku Terrane and Median Batholith in New Zealand (Ballance and Campbell, 1993; Mortimer et al., 1999; Roser et al., 2002). Subduction along Eastern Gondwanaland waned in the mid Cretaceous at about 105–100 Ma (Veevers, 1984; Bradshaw, 1989; Laird and Bradshaw, 2004), coinciding with a major plate boundary reorganization event (Veevers, 2000; Matthews et al., 2012). Convergence gave way to strike-slip motion along the margin (Veevers, 1984; Sutherland and Hollis, 2001; Siddoway, 2008), and ultimately widespread continental extension and fragmentation (Gaina et al., 1998; Sutherland, 1999; Rey and Müller, 2010), represented by a regional-scale unconformity ('Eastern Gondwana Composite Surface') in submerged portions of Zealandia (Bache et al., 2014), and a major near continent-wide angular unconformity in New Zealand (Laird and Bradshaw, 2004). Observations from Marie Byrd Land and southernmost New Zealand (Kula et al., 2007; Siddoway, 2008; McFadden et al., 2010; Saito et al., 2013) indicate widespread continental deformation, with a transition from wrench to transtensional regimes before intitation of seafloor spreading within the Tasman and Amundsen seas at around chron 34y time (83 Ma), although Vry et al. (2004) suggested that subduction may have continued along the New Zealand portion of the margin until ~85 Ma. Evidence for subduction continuing after 105–100 Ma (Worthington et al., 2006) comes from the South Island from calc-alkaline activity at ~89 Ma in the Canterbury region (Smith and Cole, 1997) and high-grade metamorphism in the Alpine Schist at ~86 Ma (Vry et al., 2004). Elsewhere around the circum-Pacific subduction continued beneath eastern Asia, North and South America, and the Antarctic Peninsula.

Another well accepted aspect of southwest Pacific evolution is that from at least 45 Ma to the present one, or multiple plate boundaries have separated the Pacific plate from the LHR. Tonga–Kermadec subduction was active from at least 45 Ma, as supported by dated arc tholeiites from the Tonga forearc and Tongan island of 'Eua (e.g.

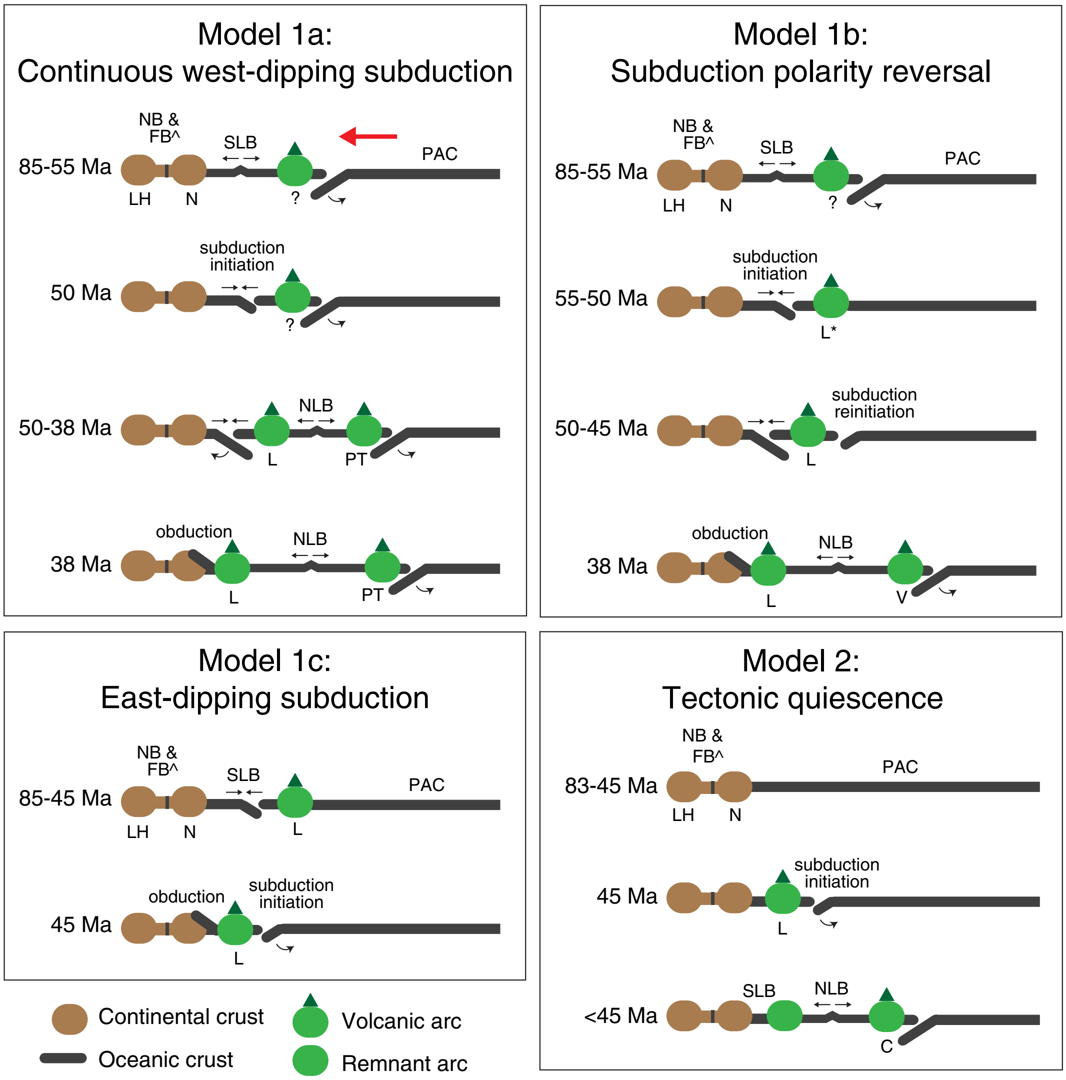

Fig. 2. Alternative tectonic models for the southwest Pacific since the mid Cretaceous, based on previously published reconstructions (see main text for a detailed explanation of each model). Model 1a is based on Schellart et al. (2006) and involves continuous west-dipping subduction to the east of the LHR due to convergence between the LHR and Pacific plate. The main features of Model 1a include: opening of the South Loyalty Basin (SLB) due to west-dipping subduction and associated slab roll-back and back-arc spreading behind an unnamed arc, and initiation of east-dipping subduction in the SLB at 50 Ma to consume the basin. In this model east and west-dipping subduction occur contemporaneously. Model 1b is based on Whattam et al. (2008) and Crawford et al. (2003) and involves continuous subduction to the east of the LHR, including a subduction polarity reversal. This model is similar to Model 1a, however west-dipping subduction is not continuous throughout the period from 85–45 Ma, and relative motion between the Pacific and LHR is not specified. Model 1c is based on Sdrolias et al. (2003) and Sdrolias et al. (2004) and involves east-dipping subduction during the entire 85–45 Ma period to consume the South Loyalty Basin that, in this particular model, had opened earlier during the period from 140–120 Ma. Model 2 involves tectonic quiescence in the southwest Pacific until ~45 Ma and is based on Steinberger et al. (2004), Mortimer et al. (2007), Sutherland et al. (2010) and Doubrovine et al. (2012). In Model 2 there is no plate boundary between the Pacific plate (PAC) and LHR (LH) until ~45 Ma when west-dipping Tonga–Kermadec subduction initiates. ^Opening of the Fairway and New Caledonia basins, the latter of which is partially floored by oceanic crust, occurs prior to 85 Ma according to Collot et al. (2009), however in the model of Schellart et al. (2006) is opens from 62–56 Ma. *In the model of Whattam et al. (2008) the Loyalty Arc develops as a new intra-oceanic arc due to subduction within the SLB and is separate to the Late Cretaceous unnamed arc further east (denoted by "?") that was associated with west-dipping subduction, whereas in the model of Crawford et al. (2003) the Loyalty Arc establishes on top of a pre-existing arc (as is shown in this figure). C, Colville Arc; L, Loyalty Arc; N, Norfolk Ridge; NLB, North Loyalty Basin; PT, proto-Tonga–Kermadec Arc; V, Vitiaz Arc.

Duncan et al., 1985; Bloomer et al., 1995), and basin subsidence events (e.g. Sutherland et al., 2010; Bache et al., 2012; Hackney et al., 2012) of a similar age. At ~45 Ma it has also been proposed that rifting initiated between the Challenger and Campbell plateaus and a plate boundary propagated through New Zealand which has been active ever since (Sutherland, 1995).

Starkly contrasting kinematic interpretations have been proposed for the tectonic evolution of the southwest Pacific in the intervening period (~85–45 Ma). Schellart et al. (2006) proposed that from 82–45 Ma there was 1500 km of subduction of the Pacific plate, while Steinberger et al. (2004) proposed that the LHR was part of the Pacific plate, with no intervening plate boundary. Amongst published tectonic reconstructions for the Late Cretaceous to Eocene evolution of the southwest Pacific, these two scenarios can be considered as end-member cases (Mortimer et al., 2007) (Fig. 2). We are going to consider four alternative reconstruction scenarios for the evolution of the southwest Pacific which explore different plate boundary configurations to the east of the LHR (Fig. 2). The first end-member, Model 1, involves continuous subduction in the southwest Pacific to the north of New Zealand, with one or several subduction polarity reversals (Crawford et al., 2003; Sdrolias et al., 2003, 2004; Schellart et al., 2006; Whattam et al., 2008; Cluzel et al., 2012a,b; Meffre et al., 2012). In striking contrast is end-member scenario Model 2, of tectonic quiescence east of the LHR until 45 Ma (Steinberger et al., 2004; Mortimer et al., 2007; Sutherland et al., 2010; Doubrovine et al., 2012). Alternative reconstruction models that incorporate subduction to the east of the LHR have been presented for this timeframe that vary in subduction polarity and timing of subduction, these are presented as models 1a–c.

1.1.1. Model 1a: Continuous west-dipping subduction

In the southwest Pacific plate reconstruction model of Schellart et al. (2006) west-dipping subduction occurs to the east of New Caledonia driven by convergence between the Pacifc and LHR. Eastward slab roll-back drives opening of the South Loyalty Basin to the east of New Caledonia from ~85–55 Ma. Schellart et al. (2006) further suggested that back-arc basin extension associated with the west-dipping subduction zone accommodated opening of the New Caledonia Basin to the west of New Caledonia from 62–56 Ma (after Lafoy et al., 2005), and partially accommodated opening of the Tasman Sea. In this model west-dipping subduction is continuous in the southwest Pacific to present-day, and occurs contemporaneous with east-dipping subduction that initiates at ~50 Ma in the South Loyalty Basin. This east-dipping subduction zone had consumed the South Loyalty Basin and subduction ended diachronously with New Caledonia obduction at ~38 Ma in the north and Northland oduction at ~25 Ma in the south.

1.1.2. Model 1b: Subduction polarity reversal

Model 1b is primarily based on the reconstructions of Crawford et al. (2003) and Whattam et al. (2008). In this model west-dipping subduction to the east of an unnamed continental ribbon, and eastward slab roll-back accommodate opening of the South Loyalty Basin from 85–55 Ma. A subduction polarity reversal occurs at 55 Ma, with the initiation of northeast-dipping subduction in the South Loyalty Basin at the site of the recently extinct spreading ridge. Reinitiation of west-dipping subduction occurs at ~50 (Whattam et al., 2008) or 45 Ma (Crawford et al., 2003; Meffre et al., 2012). In contrast to the model of Schellart et al. (2006) west-dipping subduction ceases during the Cretaceous, and east and west-dipping subduction is only contemporaneous for a short interval of time towards the end of the period during which the South Loyalty Basin is closing. In the model of Crawford et al. (2003) the subduction polarity flip at 55 Ma occurs across the unnamed continental ribbon to the east of New Caledonia. In the model of Whattam et al. (2008) an intra-oceanic arc is established by northeast-dipping subduction within the South Loyalty Basin that is separate from this continental ribbon further west. Relative motion between the Pacific and LHR is not specified in these models.

Eissen et al. (1998) also support a model of opening of the South Loyalty Basin from at least 85–55 Ma, including subduction initiation at ~50 Ma within the basin. However due to a lack of evidence for a Late Cretaceous arc to the east of New Caledonia their model does not incorporate subduction prior to ~50 Ma, rather the South Loyalty Basin opens as a marginal basin.

1.1.3. Model 1c: East-dipping subduction

In contrast to the southwest Pacific reconstructions described above, the model of Sdrolias et al. (2004) and Sdrolias et al. (2003) incorporates long-lived east-dipping subduction rather than west-dipping subduction. In this model east-dipping subduction beneath the Loyalty arc from ~85–45 Ma closes a Cretaceous aged basin that existed to the east of New Caledonia. Obduction onto New Caledonia begins at ~45 Ma. According to a later iteration of this model (Seton et al., 2012), this east-dipping subduction zone specifically closes an Early Cretaceous aged back-arc basin that opened as a result of eastward roll-back of the west-dipping Eastern Gondwanaland subduction zone from 140–120 Ma.

1.1.4. Model 2: Tectonic Quiescence

An objection to Models 1a–c is that relative motion between the LHR and the Pacific implies a plate boundary running through New Zealand throughout the Late Cretaceous and early Paleocene, whereas most observations suggest this was a period of tectonic quiescence in New Zealand (Sutherland, 2008). Further weighing against Model 1 is the lack of evidence for arc magmatism expected from subduction beneath the LHR or a continental fragment rifted from Eastern Gondwanaland between ~100 and 55 Ma (Eissen et al., 1998). Instead of subduction and backarc basin formation, this scenario favours widespread extension in the southwest Pacific during the Late Cretaceous to mid Eocene that was unrelated to subduction zone processes, followed by Tonga–Kermadec subduction initiation at ~45 Ma coincident with plate boundary activity in New Zealand (Steinberger et al., 2004; Mortimer et al., 2007; Sutherland et al., 2010; Doubrovine et al., 2012). In this model the ~84–55 Ma Tasman Sea mid-ocean ridge formed the plate boundary between the Pacific and Australian plates and elsewhere there was no plate boundary activity, rather the LHR was part of the Pacific plate (Mortimer et al., 2007).

2. PART A: Geologic constraints on plate boundary activity

In the following sections we review geologic and geophysical data from the southwest Pacific and West Antarctic Rift System that provide information about the nature and timing of plate boundary activity in the region during the Late Cretaceous to mid Eocene. These data are summarized in Fig. 3 and Table 1. Where possible we focus on primary geologic observations rather than tectonic reconstruction models. We also outline the constraints on relative motion between East and West Antarctica, the other region in the wider southwest Pacific where the timing and amount of deformation is poorly constrained.

2.1. Observations from New Caledonia and northern New Zealand

The geology of New Caledonia and the northern North Island of New Zealand preserves vital clues for deciphering southwest Pacific plate boundary evolution, particularly as slices of ocean floor have been obducted onto these continental regions thereby providing information about paleo-ocean basin development to the east of the LHR. If subduction to the east of the LHR occurred continuously from the mid or Late Cretaceous until present-day, as has been proposed in several plate tectonic reconstructions for the region (Crawford et al., 2003; Sdrolias et al., 2003; Schellart et al., 2006; Whattam et al., 2008; Ulrich et al., 2010; Cluzel et al., 2012a,b), then there should be evidence for this ongoing subduction north of New Zealand. Alternatively, if the Pacific Plate and LHR formed a single plate during this time (e.g. Steinberger et al., 2004; Doubrovine et al., 2012) then there should be no evidence for subduction nor for any other plate boundary activity, such as the opening of marginal basins, until the onset of Tonga–Kermadec subduction at ~45 Ma (e.g. Gurnis et al., 2004). Establishing the presence or absence of subduction zone indicators in the geology of New Caledonia and northern North Island is an important first step for distinguishing between southwest Pacific plate reconstruction models (Fig. 2).

The most direct evidence for subduction comes from remnants of volcanic arcs. However, rather than the absence of substantial volcanic arc material in the southwest Pacific during this period indicating a lack of subduction, it may instead reflect poor data coverage, with few wells reaching old strata, as well as the scarce volcanism that can be associated with some subduction scenarios (Crawford et al., 2003; Schellart et al., 2006; Leng and Gurnis, 2012). Back-arc basins also provide evidence for subduction, as their driving mechanism is controlled by subduction and slab roll-back. We have consequently reviewed the information about basins that opened and closed during the Late Cretaceous to mid Eocene to understand their origin as back-arc or marginal basins to complement information about arc activity.

2.1.1. New Caledonia

Autochthonous basement rocks of New Caledonia (Téremba, Boghen and Koh terranes) formed in the Carboniferous to Jurassic (e.g. Paris, 1981; Aitchison et al., 1998; Cluzel et al., 2012b), and are overlain by Late Cretaceous and younger sedimentary rocks, and obducted terranes (Aitchison et al., 1995; Maurizot and Vendé-Leclerc, 2009). Structurally overlying the basement terranes is the mafic allochthonous Poya Terrane that comprises slices of oceanic crust and abyssal sediments (Aitchison et al., 1995). In turn the Poya Terrane is structurally overlain by the New Caledonia Ophiolite, also referred to as the Peridotite Nappe (Paris, 1981) (Fig. 4), however their formational relationship with each other is a matter of ongoing debate (e.g. Lagabrielle et al., 2013). These units were obducted onto New Caledonia during the Eocene and hold important clues for understanding plate boundary activity during the Late Cretaceous to Eocene evolution of the southwest Pacific.

The Eocene obduction event is the most widely studied event in the geologic history of New Caledonia. The same event also strongly affected tectonics in the wider region, causing subsidence of the LHR and eastward tilting of the northern New Caledonia Basin, which in turn is linked with uplift of the Fairway Ridge (its western margin) and subsidence of the western margin of New Caledonia (Collot et al., 2008). Convergence leading to obduction began by ~50 Ma, as indicated by the deposition of flysch-type deposits in the Koumac–Gomen region of New Caledonia, including calci-turbidites (Maurizot, 2011). The Poya Terrane was thrust onto New Caledonia first (e.g. Cluzel et al., 2001), with peak high-pressure metamorphism occurring at ~44 Ma (Spandler et al., 2005), and obduction of the ophiolite occurred in the late Eocene by 37 Ma (Ghent et al., 1994; Fitzherbert et al., 2004). Alternatively, Lagabrielle et al. (2013) propose that the Poya Terrane originally formed the cover of the Peridotite Nappe, despite structurally underlying it at present. In their model, continuous uplift and exhumation drove detachment and gravity-driven sliding of slices of oceanic crust and mantle. The Poya Terrane first detached from the Peridotite Nappe and was emplaced on New Caledonia by gravity-driven sliding along a detachment surface. This was followed by overthrusting of the Peridotite Nappe that also experienced gravity driven sliding, this time along its serpentinite sole (Lagabrielle et al., 2013).

The Poya Terrane is a key element in deciphering the tectonic development of the southwest Pacific and represents seafloor spreading to the east of the LHR at some stage following the mid Cretaceous and prior to the Eocene. It is predominantly composed of enriched mid-ocean ridge basalts (>80%), yet also comprises back-arc basin basalt-like tholeiites (~5%) and oceanic island basalts (seamount volcanism) (Eissen et al., 1998; Cluzel et al., 2001). The back-arc basin basalt-like rocks are younger, and exhibit a Nb–Ta depletion and large-ion lithophile (LILE) enrichment with similarities to lavas erupted in well studied active back-arc basins. The Poya Terrane is related to the high pressure-low temperature Pouebo Terrane (Aitchison et al., 1995), largely exposed to the northeast of the island near the Pam Peninsula. Their similar geochemistry and isotope characteristics suggest they formed in the same basin or at least under the same conditions (Cluzel et al., 2001).

The age of formation of the Poya Terrane is uncertain. Dating has largely relied on the identification of microfossils in cherts that are interbedded in and overlay the basalts. This technique has yielded Late Cretaceous (Campanian) to latest Paleocene or earliest Eocene ages, indicating that the basin must have existed throughout these times (Cluzel et al., 2001). Due to the interbedded nature of the cherts the fossil data have been interpreted as evidence for opening of a marginal basin from the Campanian to latest Paleocene or earliest Eocene (Cluzel et al., 2001). Limited radiometric dating of Poya Terrane basalts has revealed only Cenozoic ages in the range ~37 to 65 Ma (Guillon and Gonord, 1972; Eissen et al., 1998). However, reliability of ages derived from the study of Guillon and Gonord (1972) has been questioned due to uncertainties over the amount of argon loss from the samples (Rodgers, 1975). The authors regarded their ages as discordant, and calculated an isochron age of 38.5 ± 1.5 Ma. Furthermore, Eissen et al. (1998) also noted that their K–Ar ages were likely affected by varying degrees of resetting due to metamorphism during ophiolite emplacement.

Prinzhofer (1981) obtained K–Ar dates for intrusive rocks from the Massif du Sud in eastern New Caledonia that range from ~100–80 Ma, and these can also be used to constrain the timing of formation of oceanic lithosphere adjacent to New Caledonia. He dated undeformed dolerite veins that cross-cut the Peridotite Nappe harzburgites in the Montagne des Sources region (100 ± 10 and 94 ± 9 Ma), gabbros from the Bogota Peninsula (100 ± 10 and 96 ± 10 Ma) and hornblendite and meladiorite samples from a pluton in Mouirange that also cuts through all formations of the Nappe (81 ± 8 and 80 ± 8 Ma) (Fig. 4). The Peridotite Nappe is also intruded by mafic and felsic dykes, including boninitic dykes, dated 53.1 ± 1.6 to 49.6 ± 2.8 Ma (Cluzel et al., 2006), and ~55.8 ± 1.7 Ma ($^{40}Ar/^{39}Ar$ age) amphibolite lenses are found beneath its serpentinite sole, and formed from Poya Terrane protolith according to Cluzel et al. (2012a).

2.1.2. New Zealand — North Island

In the Northland and East Cape regions of North Island, New Zealand (Fig. 1) there are exposures of oceanic crust that have been obducted onto basement rocks. Emplacement was from the northeast, and this is well-constrained based on detailed structural mapping and stratigraphic relationships (Ballance and Spörli, 1979; Spörli, 1982; Hanson, 1991; Rait et al., 1991). The Northland and East Coast (from East Cape) allochthons (Fig. 5) comprise a Cretaceous to Oligocene assemblage of thick clastic sedimentary sequences and volcanic rocks including massive and pillowed basalt, referred to as the Tangihua Volcanics (Northland Allochthon) and Matakaoa Volcanics (East Coast Allochthon). The allochthons are generally regarded as correlative although they have typically been studied separately, and the Northland Allochthon has received more attention in the literature.

Studies of the Northland Allochthon have revealed that tholeiitic basalts are dominant, although a suite of alkalic rocks are also apparent (Malpas et al., 1992). Initial geochemical analyses revealed the tholeiites were normal mid-ocean ridge basalts (Malpas et al., 1992), however later tests identified Nb, Zr and Ta depletions and LILE enrichment suggesting a subduction influence and possible formation in a back-arc basin (e.g. Thompson et al., 1997; Nicholson et al., 2000a; Whattam et al., 2004), as well as island arc tholeiites (Nicholson et al., 2000a). There is uncertainty over the petrogenic and age relationship between the tholeiitic and alkalic rocks (Thompson et al., 1997). The alkalic rocks may be related to seamounts (Malpas et al., 1992), a deep plume influence (Thompson et al., 1997), or flow of undepleted mantle from behind a retreating slab into the mantle wedge (Thompson et al., 1997). Geochemical data on the Matakaoa Volcanics were not available until the study of Cluzel et al. (2010b), and they concluded that both mafic melanges comprise mid-ocean ridge, back-arc basin and island arc tholeiite-like basalts, have similar rare earth element patterns, show depletion in Ta and Nb, and contain sediments of similar ages (Cluzel et al., 2010b).

The age and emplacement mechanism of the allochthons is a matter of ongoing debate (e.g. Whattam et al., 2006; Mortimer et al., 2007; Nicholson et al., 2007; Herzer et al., 2009). Sediments are of Cretaceous to Oligocene age, based on radiolarian and foraminifera microfossil dating (e.g. Katz, 1976; Strong, 1976, 1980; Larsen and Spörli, 1989; Hollis and Hanson, 1991; Spörli and Aita, 1994; Cluzel et al., 2010b). The age range of fossils found in the sedimentary sequences is in general agreement with the radiometric ages of lavas of 100–42 Ma reported by Brothers

Fig. 3. Time-space diagram summarizing the geologic and tectonic observations from the southwest Pacific (a–b), and wider Pacific (c) regions, including: (a) rifting (dotted lines) and seafloor spreading, (b) ages of geologic observations from New Caledonia and New Zealand, and (c) other Pacific events. The period from 83–45 Ma is highlighted in pink, and the timing of a subduction initiation event to the east of New Caledonia (55 Ma) is marked with a dashed line. Dated rock samples are filled circles with their error bars shown as gray lines. Microfossil ages are green. All other events are stars if denoting a single age, or black lines if denoting a period of time. Observations related to subduction are purple. EC, East Cape; HEB, Hawaiian–Emperor seamount chain bend; LVB, Louisville seamount chain bend; NI, North Island. *Samples from the Tonga forearc are included with the New Caledonia observations. ^The final two samples from Brothers and Delaloye (1982) are from the East Cape region (Fig. 1). **A wide spread of microfossil ages have been identified and may represent multiple volcanic events (see main text for details). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

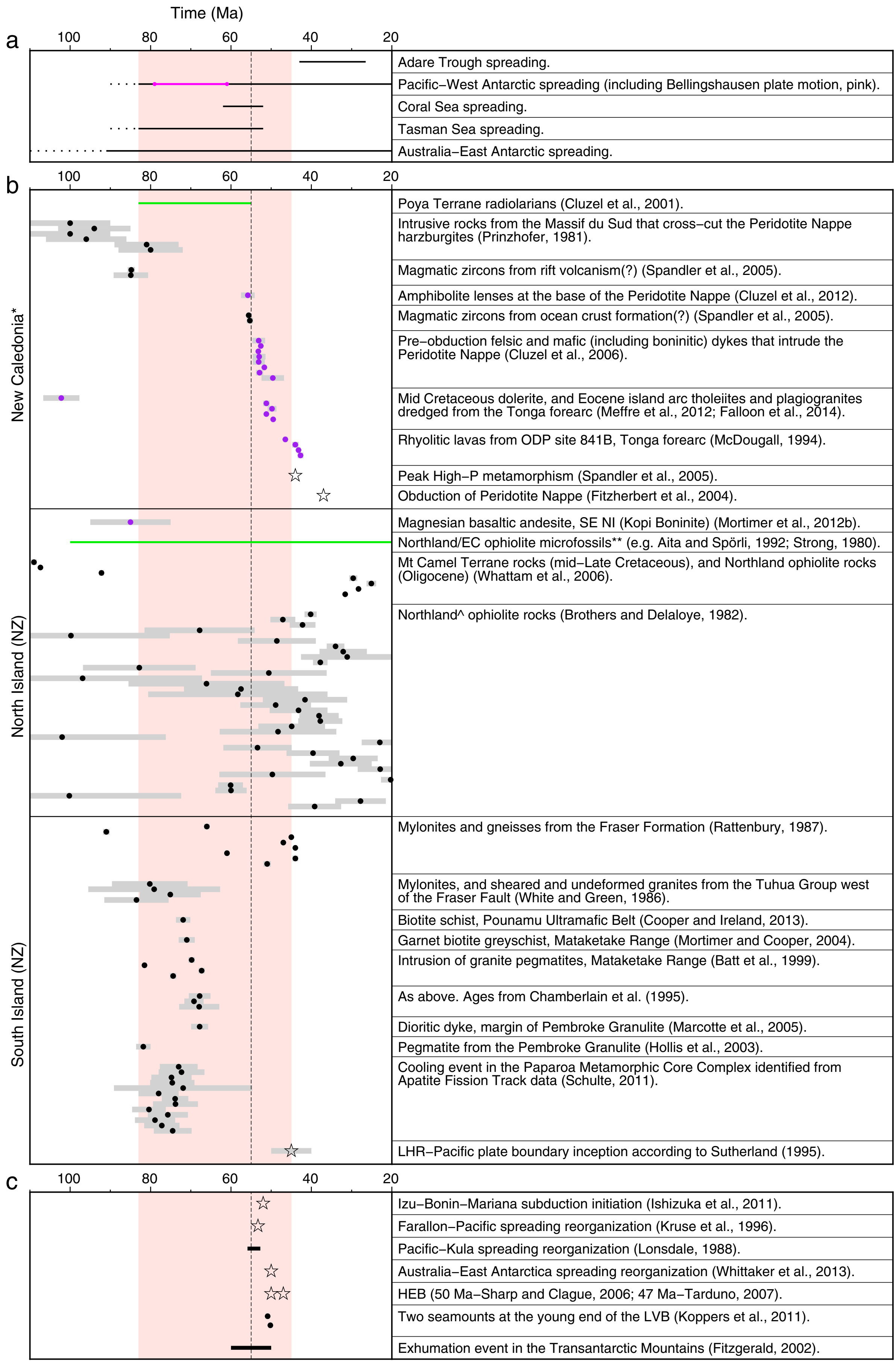
Time (Ma)
a
100
80
60
40
20
Adare Trough spreading.
Pacific–West Antarctic spreading (including Bellingshausen plate motion, pink).
Coral Sea spreading.
Tasman Sea spreading.
Australia–East Antarctic spreading.
b
New Caledonia*
Poya Terrane radiolarians (Cluzel et al., 2001).
Intrusive rocks from the Massif du Sud that cross-cut the Peridotite Nappe harzburgites (Prinzhofer, 1981).
Magmatic zircons from rift volcanism(?) (Spandler et al., 2005).
Amphibolite lenses at the base of the Peridotite Nappe (Cluzel et al., 2012).
Magmatic zircons from ocean crust formation(?) (Spandler et al., 2005).
Pre-obduction felsic and mafic (including boninitic) dykes that intrude the Peridotite Nappe (Cluzel et al., 2006).
Mid Cretaceous dolerite, and Eocene island arc tholeiites and plagiogranites dredged from the Tonga forearc (Meffre et al., 2012; Falloon et al., 2014).
Rhyolitic lavas from ODP site 841B, Tonga forearc (McDougall, 1994).
Peak High–P metamorphism (Spandler et al., 2005).
Obduction of Peridotite Nappe (Fitzherbert et al., 2004).
North Island (NZ)
Magnesian basaltic andesite, SE NI (Kopi Boninite) (Mortimer et al., 2012b).
Northland/EC ophiolite microfossils** (e.g. Aita and Spörli, 1992; Strong, 1980).
Mt Camel Terrane rocks (mid–Late Cretaceous), and Northland ophiolite rocks (Oligocene) (Whattam et al., 2006).
Northland^ ophiolite rocks (Brothers and Delaloye, 1982).
South Island (NZ)
Mylonites and gneisses from the Fraser Formation (Rattenbury, 1987).
Mylonites, and sheared and undeformed granites from the Tuhua Group west of the Fraser Fault (White and Green, 1986).
Biotite schist, Pounamu Ultramafic Belt (Cooper and Ireland, 2013).
Garnet biotite greyschist, Mataketake Range (Mortimer and Cooper, 2004).
Intrusion of granite pegmatites, Mataketake Range (Batt et al., 1999).
As above. Ages from Chamberlain et al. (1995).
Dioritic dyke, margin of Pembroke Granulite (Marcotte et al., 2005).
Pegmatite from the Pembroke Granulite (Hollis et al., 2003).
Cooling event in the Paparoa Metamorphic Core Complex identified from Apatite Fission Track data (Schulte, 2011).
LHR–Pacific plate boundary inception according to Sutherland (1995).
c
Izu–Bonin–Mariana subduction initiation (Ishizuka et al., 2011).
Farallon–Pacific spreading reorganization (Kruse et al., 1996).
Pacific–Kula spreading reorganization (Lonsdale, 1988).
Australia–East Antarctica spreading reorganization (Whittaker et al., 2013).
HEB (50 Ma–Sharp and Clague, 2006; 47 Ma–Tarduno, 2007).
Two seamounts at the young end of the LVB (Koppers et al., 2011).
Exhumation event in the Transantarctic Mountains (Fitzgerald, 2002).

Table 1
Summary of geologic and kinematic observations from the southwest Pacific, including Paleocene/Eocene events from the Pacific ocean basin.

Timing	Observation	Reference(s)
General Southwest Pacific Observations		
105–100 Ma or 86 Ma	Eastern Gondwanaland subduction ended.	105–100 Ma: Laird and Bradshaw (2004) 86 Ma: Vry et al. (2004)
Cenomanian (~100 Ma) or older	Extension in the Fairway-Aotea and New Caledonia basins begins east of the LHR (preceding Tasman Sea spreading). Extension possibly linked with the eastern Gondwanaland plate reorganization at ~105 Ma.	Collot et al. (2009)
102.2 ± 4.5 Ma	U–Pb age of a dolerite dredged from the Tonga forearc.	Meffre et al. (2012), Falloon et al. (2014)
~91 Ma	Seafloor spreading between Australia and East Antarctica begins between the Naturaliste Plateau and Bruce Rise. Initiation of spreading was diachronous and began later in the Bight Basin–Wilkes Land Sector (83–71 Ma), Otway Basin–Terre Adelie sector (68–63 Ma), and the Sorell Basin (58–49 Ma).	Direen et al. (2012)
~90 Ma	Spreading between the Pacific and West Antarctica begins. Oldest magnetic anomaly produced by spreading between the Chatham Rise (Pacific) and Marie Byrd Land: chron 34y (83 Ma). Oldest magnetic anomaly produced by spreading between the Campbell Plateau (Pacific) and Marie Byrd Land: chron 33n (79.5–73.6 Ma).	Cande et al. (1995), Larter et al. (2002), Wobbe et al. (2012)
~83–52 Ma	Tasman Sea spreading. Rifting may date back to ~90 Ma.	Gaina et al. (1998)
Cretaceous (>83 Ma)	Bounty Trough rifting between the Campbell Plateau and Chatham Rise.	Carter et al. (1994), Eagles et al. (2004a) Grobys et al. (2007)
62–52 Ma	Coral Sea spreading.	Gaina et al. (1999)
55.9 Ma	Decrease in spreading rate in the Tasman Sea accompanied by a counterclockwise change in spreading direction, leading up to spreading cessation.	Gaina et al. (1998)
51.2 Ma, 49.5 Ma	$^{40}Ar/^{39}Ar$ ages of island arc-type tholeiites dredged from the Tonga forearc. Similar ages of ~40–46 Ma come from the Tongan island of 'Eua (Duncan et al., 1985).	Meffre et al. (2012)
51.2 Ma, 49.8 Ma	U–Pb ages of plagiogranites dredged from the Tonga forearc and trench.	Meffre et al. (2012)
46.5 ± 0.5 Ma, 44.0 ± 0.8 Ma, 43.2 ± 0.5 Ma, 42.7 ± 0.7 Ma	K–Ar ages of rhyolitic lavas from ODP Site 841-B from the Tonga forearc.	McDougall (1994)
New Caledonia		
~100–80 ± 10 Ma	Period of magmatic activity, producing intrusions into the Peridotite Nappe. Age of dolerite veins that intrude the Peridotite Nappe: 100 ± 10 and 94 ± 9 Ma. Age of gabbros from the Bogota Peninsula: 100 ± 10 and 96 ± 10 Ma. Age for the Mouirange pluton that cross-cuts the Peridotite Nappe: 81 ± 8 and 80 ± 8 Ma.	Prinzhofer (1981)
~85–55 Ma	Possible opening of the South Loyalty Basin E/NE of New Caledonia.	e.g. Cluzel et al. (2001), Spandler et al. (2005)
84.8 ± 0.9 Ma, 84.9 ± 4.3 Ma	Magmatic zircon ages from Pouebo Terrane pelitic schists. Rift volcanism is interpreted to be the sediment source.	Spandler et al. (2005)
Campanian to latest Paleocene or earliest Eocene	Age of microfossils interbedded in and overlying the Poya Terrane.	Cluzel et al. (2001)
55.8 ± 1.7 Ma	Age of amphibolite lenses found at the base of the Peridotite Nappe that record high temperature metamorphic conditions.	Cluzel et al. (2012a)
55.6 ± 0.5 Ma, 55.3 ± 0.8 Ma	Magmatic zircon ages from a Pouebo Terrane gneiss with a volcaniclastic origin and schist with an igneous origin. Ocean spreading environment inferred.	Spandler et al. (2005)
~55 Ma	Subduction initiation event inferred from dyke intrusion, formation of amphibolite lenses, and correlation with Izu–Bonin–Mariana subduction initiation at ~52 Ma (Ishizuka et al., 2011).	Crawford et al. (2003), Cluzel et al. (2006), Cluzel et al. (2012a,b)
53 Ma	Intrusion of pre-obduction dykes into the Peridotite Nappe. (Range of ages: 53.1 ± 1.6–49.6 ± 2.8 Ma)	Cluzel et al. (2006)
44 Ma	Poya Terrane obduction and eclogites of the Pouebo Terrane experience peak metamorphic conditions.	Spandler et al. (2005)
37 Ma	Peridotite Nappe obduction.	Ghent et al. (1994), Fitzherbert et al. (2004)
New Zealand – North Island		
Cretaceous to Oligocene	Age of microfossils interbedded in or overlying the Northland and East Coast allochthons.	Katz (1976), Strong (1976), Strong (1980), Larsen and Spörli (1989), Hollis and Hanson (1991), Spörli and Aita (1994), Cluzel et al. (2010b)
100–42 Ma or 36–25 Ma	Radiometic ages of Northland Ophiolite rocks.	100–42: Brothers and Delaloye (1982) 36–25: Whattam et al. (2006)
85 ± 10 Ma	Eruption of a magnesian basaltic andesite (Kopi Boninite, Moore, 1980). Now preserved in southeastern North Island.	Mortimer et al. (2012)
Campanian–latest Paleocene/ earliest Eocene	Possible formation of ocean floor to the north of New Zealand. Referred to as the Tangihua/Matakaoa Basin by Cluzel et al. (2010b).	Nicholson et al. (2000a,b), Whattam et al. (2006), Cluzel et al. (2010b),
New Zealand – South Island		
110 Ma	Paparoa Metamorphic Core Complex extension begins.	Tulloch and Kimbrough (1989)
100 Ma	Major angular unconformity, expresses change from compressional to extensional tectonic regime.	Laird and Bradshaw (2004)
Late Cretaceous (<88.4 ± 1.2 Ma)	Transpression at the Straight River Shear Zone in Fiordland.	King et al. (2008)
80–72 Ma	Thermal event in the Paparoa Metamorphic Core Complex linked with burial and re-exhumation, unrelated to initial metamorphic core complex development.	Schulte (2011)
82–67 Ma	High-grade metamorphism at the Mataketake Range (71 ± 2 Ma), coincident with pegmatite intrusions (82–67 Ma).	Chamberlain et al. (1995), Batt et al. (1999), Mortimer and Cooper (2004)
80 Ma (Tuhua Group near Fraser Fm), 91–45 Ma (ages from Fraser Fm)	Deformation event involving mylonitisation within and near the Fraser Complex.	White and Green (1986), Rattenbury (1987)
Latest Cretaceous or early	Dextral transpressional event at the Anita Shear Zone in Fiordland. Followed mid	Klepeis et al. (1999)

(continued on next page)

Table 1 *(continued)*

Timing	Observation	Reference(s)
Cenozoic	Cretaceous extension and preceded modern Alpine Fault transpression.	
71.9 ± 1.8 Ma	Metamorphism in the Pounamu Ultramafic Belt (Alpine Schist). Possibly associated with an Alpine Schist deformation event due to temporal correlation with metamorphism in the Mataketake Range.	Cooper and Ireland (2013)
45 ± 5 Ma	Propagation of plate boundary into New Zealand and rifting between the Challenger and Campbell plateaus.	Sutherland (1995)
Antarctica		
105–80 Ma	Postulated main phase of extension between East and West Antarctica that ended with Pacific–Campbell Plateau breakup.	Lawver and Gahagan (1994)
98–95 Ma	Shear zone deformation evidenced by mylonitisation in the Colbeck Trough.	Siddoway et al. (2004)
79–61 Ma	Independent motion of the Bellingshausen plate adjacent to Marie Bird Land.	Cande et al. (1995), Larter et al. (2002), Eagles et al. (2004b)
61–53 Ma	Possible early extension in the Adare Trough.	Cande and Stock (2004)
60–55 Ma	Main phase of uplift in the Transantarctic Mountains begins.	Gleadow and Fitzgerald (1987), Fitzgerald (1992), Fitzgerald (1994), Fitzgerald (2002), Elliot (2013)
60–50 Ma	Exhumation event in the Transantarctic Mountains beginning at ~60 Ma in northern and southern Victoria Land and at ~50 Ma further south.	Fitzgerald (2002)
43–26 Ma	Spreading in the Adare Trough.	Cande et al. (2000), Granot et al. (2013)
Paleocene/Eocene aged events		
55.9–52.7 Ma	Spreading reorganization at the Pacific–Kula spreading ridge identified from fracture zone and abyssal hill trends.	Lonsdale (1988)
~53.3 Ma	Spreading reorganization at the Pacific–Farallon spreading ridge identified from fracture zone trends.	Kruse et al. (1996)
52 Ma	Izu–Bonin–Mariana subduction initiation.	Ishizuka et al. (2011)
~53–46 Ma	Change from NW–SE to NNE–SSW spreading at the Australian–East Antarctic spreading ridge.	Whittaker et al. (2013)
50.9 ± 0.5 Ma, 50.2 ± 0.5 Ma	Maximum ages for two seamounts at the younger end of the Louisville seamount chain bend, suggesting that bend formation began slightly earlier.	Koppers et al. (2011)
50 or 47 Ma	Hawaiian–Emperor seamount chain bend.	50 Ma: Sharp and Clague (2006)

and Delaloye (1982). Brothers and Delaloye (1982) obtained a clustering of late Eocene to Oligocene ages (Brothers and Delaloye, 1982), and only sporadic Cretaceous to Paleocene ages, many of which were associated with large uncertainties ranging from ± 10 to ± 29 Myr for all but one of these samples (Brothers and Delaloye, 1982). They calculated an isochron age of 33.6 ± 2.1 Ma for the North Island ophiolites. Obduction may have resulted from nucleation or margin approach of a ~ northeast-dipping subduction zone, or crustal delamination (or "flaking") associated with initiation of a west-dipping subduction zone (Malpas et al., 1992; Herzer et al., 2009; Rait, 2000; Whattam et al., 2005; Whattam et al.,

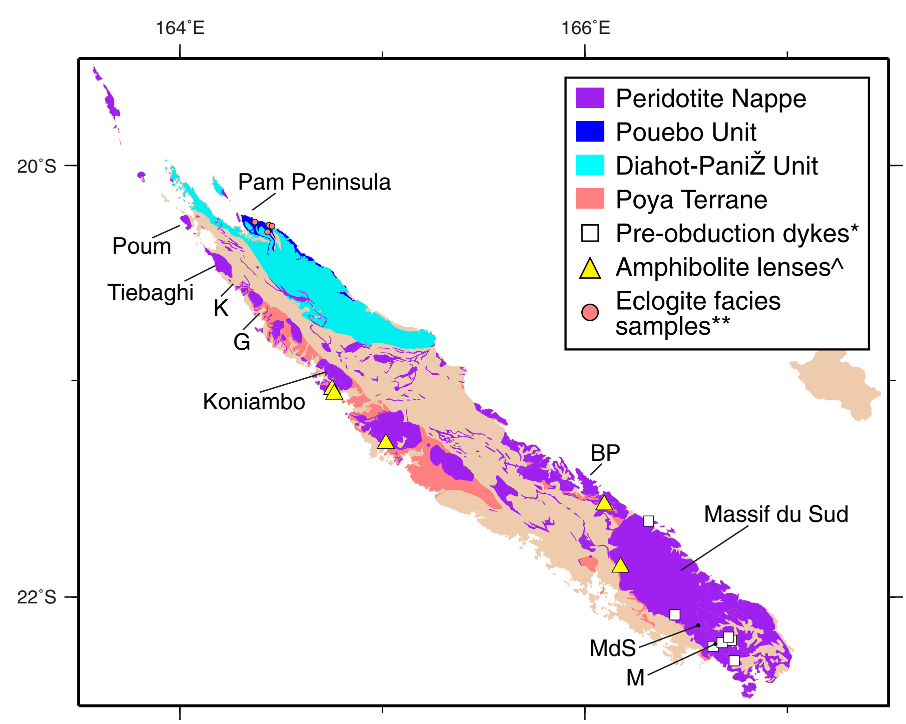

Fig. 4. Geologic map of New Caledonia showing the main geologic terranes/units that were obducted in the Cenozoic (outlines from Maurizot and Vendé-Leclerc, 2009), and sample sites described in the main text. *Pre-obduction dykes are from Cluzel et al. (2006). ^Amphibolite lenses from Cluzel et al. (2012a). **Eclogite facies samples from Spandler et al. (2005). BP, Bogota Peninsula; G, Gomen; K, Koumac; M, Mouirange; MdS, Montagne des Sources.

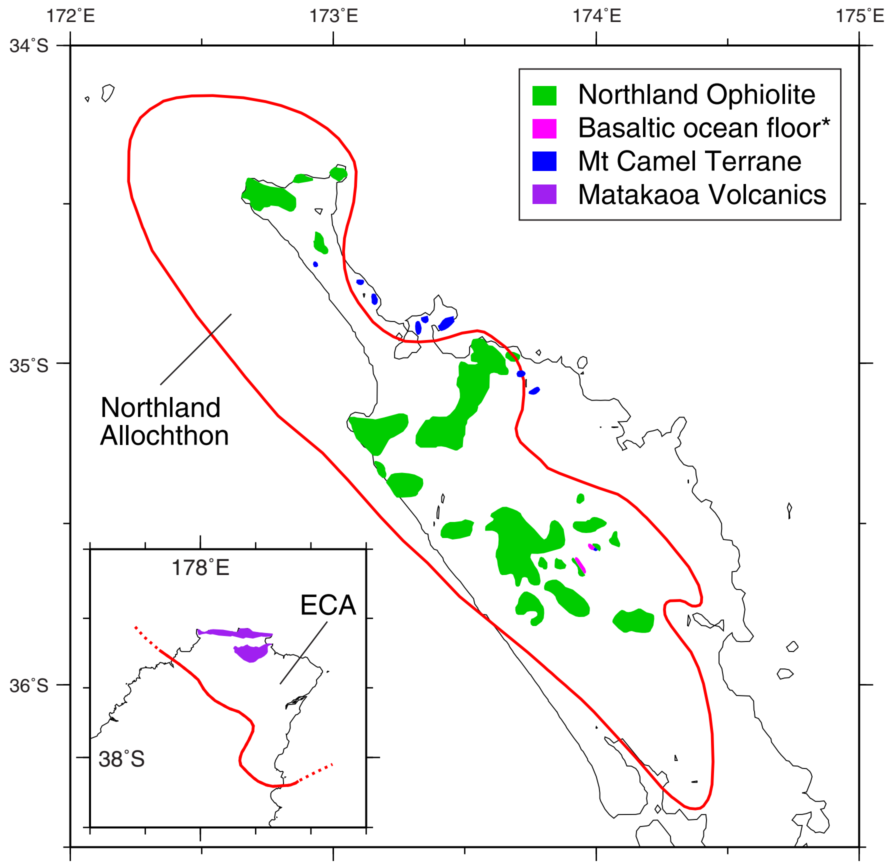

Fig. 5. Northland and East Cape (inset) regions of North Island, New Zealand (see Fig. 1 for their locations). Outlines of the surface extent of the Northland and East Coast (ECA) allochthons are from Adams et al. (2013), all other outlines are from Whattam et al. (2006). *The basaltic ocean floor rocks were interpreted by Whattam et al. (2006) and in their model represent the floor of an ocean basin that was once located to the north of New Zealand.

2006; Mortimer et al., 2007; Schellart, 2007; Booden et al., 2011). These mechanisms are both consistent with the emplacement from the northeast (Ballance and Spörli, 1979; Spörli, 1982; Hanson, 1991; Rait et al., 1991).

2.1.3. Evidence for the presence or absence of volcanic arcs

There is little direct evidence that active Late Cretaceous to mid Eocene volcanic arcs existed in the southwest Pacific. This is perhaps contrary to what may be expected if subduction was prevalent in the region to the north of New Zealand and oceanward of the LHR as is incorporated in many published plate reconstructions (Crawford et al., 2003; Sdrolias et al., 2003; Schellart et al., 2006; Whattam, 2009; Cluzel et al., 2012b). Arc products from the well-established late Paleozoic to mid Cretaceous phase of subduction beneath Eastern Gondwanaland are abundant throughout eastern Australia and Zealandia (Paris, 1981; Campbell, 1984; Ballance and Campbell, 1993; Mortimer et al., 1999; Roser et al., 2002; Adams et al., 2009), and evidence of post-Eocene subduction is also clearly preserved, for instance in the Lau-Colville, New Hebrides and Tonga–Kermadec ridges.

Mortimer et al. (2012) obtained an age of 85 ± 10 Ma for a magnesian basaltic andesite (Kopi Boninite, Moore, 1980) from the Pahaoa Group, Wairarapa district in southeastern North Island, New Zealand (Fig. 1). This boninitic rock is interpreted as evidence for subduction reinitiation along northern Zealandia some time during the period 95–75 Ma (Mortimer et al., 2012). Apart from this boninite, direct evidence for subduction during this Late Cretaceous–Eocene timeframe is restricted to the latest Paleocene–earliest Eocene near New Caledonia.

Extensive analysis of rocks from the New Caledonia Peridotite Nappe and underlying obducted ocean floor terranes suggests that there was a subduction initiation event at ~55 Ma to the east of New Caledonia (Eissen et al., 1998; Cluzel et al., 2001, 2006, 2012a,b; Ulrich et al., 2010). Intrusion of boninitic dykes into the Peridotite Nappe (Cluzel et al., 2006) provide evidence for subduction inception (Spandler et al., 2005), as boninites are often associated with the early stages of subduction and high temperatures (e.g. Stern and Bloomer, 1992). These boninitic dykes, along with other pre-obduction felsic and mafic dykes with supra-subduction zone affinities that intrude the Peridotite Nappe (Fig. 4, white squares) were interpreted to have formed in a single magmatic event with an average age of ~53 Ma. The ~56 Ma amphibolite lenses found beneath the serpentinite sole of the Peridotite Nappe and above the Poya Terrane formed under high temperature metamorphic conditions slightly earlier than the dyke intrusions (Cluzel et al., 2012a) (Fig. 4, yellow triangles). They recrystallized at 850–980 °C and 0.5 GPa, at odds with the high pressure-low temperature metamorphic conditions of the eclogite and blueschist facies rocks of the Pouebo and Diahot Terranes (Aitchison et al., 1995; Clarke et al., 1997; Fitzherbert et al., 2003) (Fig. 4), that are typical of metamorphic rocks associated with subduction, and the largely unmetamorphosed Poya Terrane (Cluzel et al., 2012a). This suggested to Cluzel et al. (2012a) that subduction inception involved young and warm lithosphere and may have occurred at or near a spreading ridge. Eclogites of the Pouebo Terrane experienced peak metamorphic conditions later, at ~44 Ma (Spandler et al., 2005), on the order of 600 °C and 2.4 GPa (Clarke et al., 1997). The proximal occurrence of abyssal- and supra-subduction peridotites, and contemporaneous ridge and arc volcanism has been proposed as evidence to support the hypothesis of subduction inception near a spreading ridge (Ulrich et al., 2010). Ulrich et al. (2010) studied the geochemical trends of the Peridotite Nappe, in particular lherzolites and highly depleted harzburgites from four massifs across New Caledonia (Fig. 4 — Poum, Tiebaghi, Koniambo and Massif du Sud) and attributed a change in melting style, from dry melting in a ridge environment to hydrous melting in a supra-subduction environment, to a shift from extension to convergence, and specifically subduction inception at or near a spreading ridge. In the models of Ulrich et al. (2010) and Cluzel et al. (2012a,b) subduction initiated at the spreading ridge in the South Loyalty Basin, believed to be a back-arc basin that opened to the east of New Caledonia during the Late Cretaceous to Paleocene. In their models changes in plate motion caused compression in the basin and its closure due to subduction.

Tholeiites (back-arc basin-type and island arc-type) dredged from the Tonga forearc and trench that yield ages of 52–49 Ma (Fig. 1), have been used to better understand the evolution of the South Loyalty Basin and region to the east of New Caledonia (Meffre et al., 2012). Formation of these rocks is associated with subduction and is consistent with two alternative scenarios, either they formed in the arc-back-arc transition region of the Loyalty subduction zone discussed by Cluzel et al. (2012a) and this region later became the forearc region of the west-dipping Tonga–Kermadec subduction zone, or alternatively these rocks could have formed in the forearc region of the Tonga–Kermadec subduction zone during incipient subduction (Meffre et al., 2012). A ~102 Ma dolerite was dredged from the same location in the Tonga forearc and displays geochemical similarities with the Poya Terrane, rather than mid-ocean ridge basalts of the Pacific plate (Falloon et al., 2014).

2.1.4. Extensional basins

The southwest Pacific is often described as having undergone widespread extension and formation of marginal or back-arc basins during the Late Cretaceous to mid Eocene (Aitchison et al., 1995; Eissen et al., 1998; Nicholson et al., 2000b; Crawford et al., 2003; Spandler et al., 2005; Whattam et al., 2005; Whattam et al., 2008; Cluzel et al., 2010a). The Poya Terrane (Fig. 4) and parts of the Northland and East Coast allochthons (Fig. 5) are commonly considered to have formed the basement of oceanic extensional basins to the east of the LHR. While it is possible that they may represent slices of the Pacific plate, this is considered unlikely as the Pacific seafloor to the east of the Lord Howe Rise at the time of obduction in New Caledonia and New Zealand is mid Cretaceous in age (Seton et al., 2012). This Pacific seafloor formed at a spreading ridge system that became inactive during the Late Cretaceous (Chandler et al., 2012; Seton et al., 2012), and therefore cannot account for the Paleocene–Eocene microfossils interbedded in the obducted basalts.

The Poya Terrane and the Northland and East Coast allochthons may hold important clues for deciphering the tectonic setting of the southwest Pacific during the Late Cretaceous to mid Eocene, as if they formed during this period they indicate extensional plate boundary activity occurred to the east of the LHR and possibly subduction if they formed as back-arc basins. Most aspects of their formation are debated, including their tectonic setting, age and emplacement mechanism. Determining whether or not these basins formed specifically due to back-arc spreading, rather than simple marginal basin extension unrelated to subduction is facilitated by careful geochemical analyses of rock samples. For instance, depletion of Nb and Ta is commonly used to infer a subduction zone signature during formation (e.g. Stolz et al., 1996) and differentiate a rock sample from being a normal mid-ocean ridge basalt that did not form in close proximity to a subduction zone, but rather in a marginal basin that formed due to divergence between two plates. Enrichment in LILEs may also indicate subduction zone processes during melting. However, these characteristics are non-unique. Nb and Ta depletion may indicate formation in a back-arc basin or melting from a mantle previously modified by subduction processes, and LILE enrichment may have several causes including, but not limited to, fluid enrichment from slab dehydration and hydrothermal alteration (e.g. Cluzel et al., 2001, 2010a,b; Ulrich et al., 2010). Therefore combining multiple geochemical indicators is crucial for isolating the tectonic setting of a rock sample.

Back-arc basin basalt-like tholeiites comprise ~5% of the Poya Terrane (Eissen et al., 1998; Cluzel et al., 2001) and their geochemical signature shares similarities to lavas erupted in well studied active back-arc basins. Cluzel et al. (2001) noted that this back-arc basin geochemical signature may alternatively be due to melting of mantle previously affected by subduction such as pre-mid Cretaceous long-lived subduction beneath Eastern Gondwanaland. They suggest that if formation of the Poya Terrane did occur in a back-arc basin then it was an atypical back-arc basin setting due to the predominance of enriched mid-ocean ridge basalts that suggest a deep mantle source. The basin that was once floored by the Poya Terrane is commonly referred to as the South Loyalty Basin (Cluzel et al., 2001, 2012a,b; Schellart et al., 2006; Ulrich et al., 2010; Meffre et al., 2012; Lagabrielle et al., 2013), as any remnant basin crust would likely be preserved south of the Loyalty Islands (Cluzel et al., 2001), although alternatively the "South Loyalty Basin" is the name given to a Cretaceous back-arc basin that opened from 140–120 Ma to the east of the Norfolk Ridge (Seton et al., 2012).

The age of the South Loyalty Basin, comprising the Poya Terrane, is uncertain, along with the driving mechanism for its formation. Radiometric dating from Prinzhofer (1981) suggest a mid-early Late Cretaceous age, while fossil dating suggests formation during the Late Cretaceous to Paleocene (Cluzel et al., 2001). Magmatic zircons from Pouebo Terrane eclogite melange rocks sampled from the Pam Peninsula provide further constraints on timing of formation of the Poya Terrane and yield ages of 84.8 ± 0.9, 84.9 ± 4.3, 55.6 ± 0.5 and 55.3 ± 0.8 Ma (Spandler et al., 2005) (Fig. 4, light red dots). Magmatic zircons form during melt crystallization, as opposed to metamorphic zircons that record subsequent metamorphism, and can therefore reveal information about the nature and timing of tectonic environments. The older samples of Spandler et al. (2005) were pelitic schists (metamorphosed pelitic sedimentary rocks) and they suggested that the source for the sediments was rift volcanism rather than subduction volcanism as an associated arc has not been located. The younger two samples were a garnet–quartz–phengite gneiss and a chlorite–epidote–phengite schist. The gneiss (55.3 ± 0.8 Ma) did not have an igneous origin but rather formed from a sedimentary rock. Spandler et al. (2005) proposed that the rock had a volcaniclastic origin, either related to rifting or subduction zone volcanism, or it may have been a quartz-rich variety of the ~85 Ma samples. The chlorite–epidote–phengite schist sample (55.6 ± 0.5 Ma) had a mafic intrusive igneous origin and Spandler et al. (2004) identified the protolith as a plagioclase-rich cumulate (troctolite or leucogranite). The Mg, Ni, Sc and V concentrations were typical of ocean crust, while the Al_2O_3, LILE, U and S concentrations and U/Th ratio were anomalously high for ocean crust. Spandler et al. (2004) directly related the rock to ocean crust formation. Based on the dating of Prinzhofer (1981) and Spandler et al. (2005) there appears to be a clustering of radiometrically derived ages preceding ophiolite obduction, a group at 100–80 Ma and two ages at 55 Ma. Fossil dating suggests formation from ~85–55 Ma (Cluzel et al., 2001).

Mid-ocean ridge, back-arc basin and island arc tholeiite affinities have been identified for the Northland and East Coast allochthons, and these ocean basin-derived rock assemblages have likely been influenced by supra-subduction zone processes (e.g. Hopper and Smith, 1996; Thompson et al., 1997; Nicholson et al., 2000a; Whattam et al., 2004, 2005; Cluzel et al., 2010b). For instance, LILE enrichment was identified in all the basalt and diorite samples analysed from the Matakaoa Volcanics by Cluzel et al. (2010b), and this signature was related to supra-subduction zone processes. The Matakaoa Volcanics also host Late Cretaceous to early Eocene volcanogenic massive sulfide (VMS) deposits that have similar mineral assemblages to VMS deposits from arc/back-arc settings in the western Pacific (Brathwaite et al., 2008). Geochemical analyses are not straightforward however, for instance depletion of Nb and Ta was used to discriminate the back-arc basin signature from the normal mid-ocean ridge basalt signature in rocks from the Matakaoa Vocanics (Fig. 5), however the mid-ocean ridge basalts also show minor Nb depletion (Cluzel et al., 2010b). There is also uncertainty over the age of the rocks with supra-subduction affinities, and a pre-Miocene arc is yet to be identified off northern New Zealand (Herzer et al., 2009).

The wide spread of fossil ages (e.g. Katz, 1976; Strong, 1976, 1980; Larsen and Spörli, 1989; Hollis and Hanson, 1991; Spörli and Aita, 1994; Cluzel et al., 2010b) and radiometric ages from Brothers and Delaloye (1982) have been used to infer the presence of a single ocean basin (Nicholson et al., 2007; Cluzel et al., 2010b) that was obducted in the Oligocene or earliest Miocene (Malpas et al., 1992; Rait, 2000; Mortimer et al., 2003). However, it was suggested by several early studies that based on such a wide spread of fossil ages there may have been multiple volcanic events (Katz, 1976; Strong, 1976; Strong, 1980). More recent radiometric age dating has highlighted a possible bimodal distribution of ages interpreted to represent two separate groups of rocks, and this has been supported by identification of petrological differences between these groups (Whattam et al., 2004, 2005, 2006). Tholeiites from the Northland Allochthon yielded $^{40}Ar/^{39}Ar$ ages of 29.6 ± 1 to 25.1 ± 1.2 Ma (with an additional younger age of 18.7 ± 0.3 Ma) and a U-Pb age of 28.3 ± 0.2 (Whattam et al., 2005), and a comparable U-Pb age of 31.6 ± 0.2 Ma was obtained for an ophiolite gabbro sample (Whattam et al., 2006). Based on this clustering of younger ages Whattam et al. (2006) suggested that the Northland Ophiolite formed between 32–26 Ma. Older mid-Late Cretaceous rocks ($^{40}Ar/^{39}Ar$: 109 ± 0.7, 107.4 ± 0.2 and 92.2 ± 0.6 Ma) were interpreted as arc related based on La–Y–Nb and Zr–Nb–Y tectonic discrimination diagrams, and linked with the Mt Camel Cretaceous basement terrane (Whattam et al., 2004, 2005) (Fig. 5). Whattam et al. (2006) later suggested that a Late Cretaceous–Paleocene basin may also have existed to the north of North Island, however ages for their "Northland basaltic ocean floor terrane" are only constrained by fossil data from two locations (Fig. 5), and it is pointed out by the authors that Late Cretaceous to Paleocene fossil ages come from only a small volume of basalts.

To the west of New Caledonia, between the Norfolk Ridge and LHR exists a series of N–S trending basins, including the Fairway-Aotea and New Caledonia basins, that also formed as a result of the widespread extensional regime that accompanied Eastern Gondwanaland fragmentation. Extension in these basins precedes Tasman Sea spreading and was possibly associated with the Eastern Gondwanaland reorganization event in the mid Cretaceous (Collot et al., 2009). Extension in the New Caledonia Basin was previously thought to have occurred during the Late Cretaceous to Paleocene (Lafoy et al., 2005), however this timing was revised by the work of Collot et al. (2009). For instance, Collot et al. (2009) identified that an 85 Ma lineament called the Barcoo–Elizabeth Fracture Lineament displaces the Fairway and New Caledonia basins and must therefore post-date their formation. Based on a thinner sedimentary cover in the New Caledonia Basin, Collot et al. (2009) also suggested that it might be younger than the Fairway Basin. Sutherland et al. (2010) also investigated subsidence events to the west of New Caledonia. They suggest that the New Caledonia Trough (a physiographic feature distinct from a series of underlying basins, including the New Caledonia Basin and Fairway-Aotea basins) had a two-stage subsidence history. Based on their analysis of borehole and seismic data Cretaceous subsidence in the New Caledonia Trough was followed by widespread subsidence in the Eocene–Oligocene. In the model of Sutherland et al. (2010) the Eocene–Oligocene subsidence event was associated with lithospheric delamination, and the delamination event initiated modern Tonga–Kermadec subduction.

2.2. Observations from southern New Zealand

Following the cessation of Eastern Gondwanaland subduction and convergent margin tectonics in the mid or Late Cretaceous New Zealand is commonly described as being tectonically quiet until the mid Eocene, ~45 Ma (Kamp, 1986; Sutherland, 1999). By 45 Ma Tonga–Kermadec subduction had initiated to the north of the North Island as constrained by dated arc rocks (Duncan et al., 1985; Bloomer et al., 1995), and at 45 ± 5 Ma rifting between the Campbell and Challenger plateaus occurred (Sutherland, 1995) to form the Resolution Ridge rift margin and its conjugate along the western Campbell Plateau, the Emerald Basin and the Southeast Tasman Ocean Crust (Fig. 1). Chron 18 (~40.1 Ma) has been identified adjacent to the Southeast Tasman Ocean Crust margin (Wood et al., 1996) and chron 24 (~53.3 Ma) is located to the southwest of the Southeast Tasman Ocean Crust (Lawver and Gahagan, 1994) suggesting rifting propagated in a northeasterly direction (Wood et al., 1996) following the cessation of Tasman Sea spreading (Sutherland, 1995; Barker et al., 2008). 45 Ma is commonly cited as the age of inception of the Pacific–Australian plate boundary (Sutherland, 1995), at the location of the Alpine Fault.

This period of purported tectonic quiescence from the Late Cretaceous to mid Eocene coincides with the opening of the Tasman Sea. Extension associated with Tasman Sea spreading has been linked with extension along the West Coast region of the South Island of New Zealand (Laird, 1994) and core complex re-exhumation in southern New Zealand (Schulte, 2011). Furthermore there are observations of some minor deformation near the present-day location of the Alpine Fault during this time that may signify limited motion precursory to modern Alpine Fault motion, although this is only speculative (e.g. White and Green, 1986; Mortimer and Cooper, 2004).

2.2.1. Activity near the present-day location of the Alpine Fault

During the Late Cretaceous rocks from the Straight River and Anita shear zones in Fiordland (Fig. 6) record dextral shear (White and Green, 1986; Klepeis et al., 1999; King et al., 2008). The Straight River Shear Zone in central Fiordland is a 10 km wide transpressional shear zone that presently parallels the Alpine Fault, which is located 10–30 km offshore at this latitude (King et al., 2008). Based on cross-cutting relationships it initiated after 88.4 ± 1.2 Ma, as this is the U-Pb age assigned to a pegmatite dyke that cuts through extensional fabric and indicates the end of mid Cretaceous extension (King et al., 2008). This age also extended the known period of extension in Fiordland (King et al., 2008). Rutile ages of 73.9 ± 0.5 and 65.8 ± 0.5 Ma from a western Fiordland orthogneiss represent cooling below 400–450 °C (Flowers et al., 2005) and have been interpreted by King et al. (2008) to possibly represent a lower limit for the age of the Straight River Shear Zone. The 4 km wide Anita Shear Zone (Hill, 1995) is located in northern Fiordland near where the Alpine Fault extends offshore, and similarly to the Straight River Shear Zone it parallels the Alpine Fault. Two major amphibolite facies deformational events have affected the Anita Shear Zone and are described by Klepeis et al. (1999). The first event ("D_2"), which occurred in the mid Cretaceous involved crustal thinning that lead to decompression and exhumation, and coincided with normal faulting in central Fiordland and the Paparoa Metamorphic Core Complex further north. This extensional event was followed by a dextral transpressional deformation event ("D_3") that involved up to more than 70% shear zone perpendicular crustal shortening in some areas, based on kinematic vorticity analyses, and may also have involved exhumation. This event was post-mid Cretaceous (as it followed D_2) and pre-late Cenozoic (as this is when deformation at the Anita Shear Zone is associated with modern Alpine Fault transpression). Klepeis et al. (1999) estimated that it probably occurred during the latest Cretaceous or early Cenozoic. Also in northern Fiordland, ~5 km east of the Anita Shear Zone Marcotte et al. (2005) obtained an age of 67.8 ± 2.1 Ma for a dioritic dyke that was emplaced along the margin of the Pembroke Granulite and Milford Gneiss, and near this sample Hollis et al. (2003) obtained an age of 81.8 ± 1.8 Ma for a pegmatite from the Pembroke Granulite (Fig. 6, cyan dot). In both studies the intrusions were linked with regional extension that led to dyke emplacement in the West Coast region (discussed in the previous section).

Dextral transpression is also recorded further north along the trace of the Alpine Fault at this time, in the Fraser Complex and to the west of the Fraser Fault (Fig. 6). The Fraser Complex is bounded to the southeast by the Alpine Fault and the northwest by the Fraser Fault, and comprises gneisses, granitoids and dykes, which have been deformed by

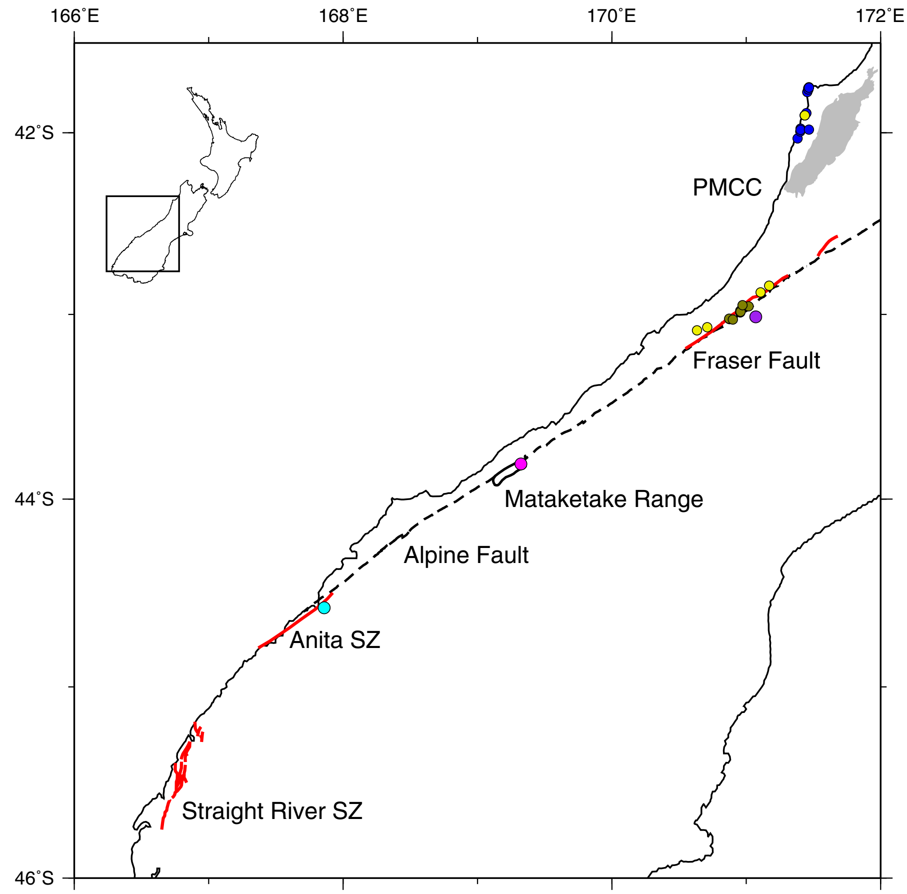

Fig. 6. Map of South Island, New Zealand, showing faults and sample sites described in the main text. In the region of the Paparoa Metamorphic Core Complex, Schulte's (2011) samples are blue, and a granite gneiss sample from White and Green (1986) is yellow. In the region of the Fraser Fault samples of mylonitized rocks are from Rattenbury (1987) — brown, and White and Green (1986) — yellow. To the east of the Fraser Fault the location of a biotite schist sample (Cooper and Ireland, 2013) is purple. In the Mataketake Range the schist sample of Mortimer and Cooper (2004) is pink. Samples from Chamberlain et al. (1995) and Batt et al. (1999) are from a similar part of the Mataketake Range. The diorite dyke sample from Marcotte et al. (2005) near the Anita Shear Zone is cyan. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

mylonitisation during dextral shear in some areas (e.g. White and Green, 1986; Rattenbury, 1987, 1991). White and Green (1986) dated rocks from mylonitised zones in the Fraser Complex and in the nearby Tuhua Group in order to study the development of the Alpine Fault (Fig. 6, yellow dots). They identified a regional deformation event from rocks in the Tuhua Group at ~80 Ma by combining structural observations with zircon and sphene fission track dating of mylonites, sheared granites and orthogneisses. Rattenbury (1987) obtained K–Ar dates for mylonites and gneisses within the Fraser Formation (Fig. 6, brown dots) and found whole rock ages in the range 91–45 Ma, and a biotite age cluster at 61–44 Ma. White and Green (1986) suggested that the deformation event at ~80 Ma represents an early stage in the development of the Alpine Fault, and Schellart et al. (2006) interpreted this deformation to represent oblique dextral convergence along a plate boundary that separated the Challenger Plateau (LHR) from the Campbell Plateau (Pacific plate).

Located between Fiordland and the Fraser Complex, rocks from the Mataketake Range immediately to the east of the Alpine Fault (part of the Haast Schist belt) (Fig. 6), and therefore on the Pacific Plate, record latest Cretaceous high-grade metamorphism and high temperature mineral growth (Mortimer and Cooper, 2004). Mortimer and Cooper (2004) obtained a U–Pb monazite age of 71 ± 2 Ma for a Mataketake schist sample (Fig. 6, pink dot), and noted that this age of metamorphism coincides with the intrusion of proximal anatectic granite pegmatites (Chamberlain et al., 1995; Batt et al., 1999). Chamberlain et al. (1995) computed U-Pb ages of 69.2 ± 2.4 (monazite age), 67.9 ± 5 and 67.8 ± 2.7 (monazite age) Ma for the pegmatites and suggested crystallization occurred at ~68 Ma, and Batt et al. (1999) suggested intrusion occurred over the protracted period 82–67 Ma, based on U–Pb ages of 81.5 ± 0.5, 74.4 ± 0.8, 69.8 ± 0.8 and 67.3 ± 0.8 Ma. Similarly to White and Green (1986) who investigated mylonitisation in the Fraser Complex, Mortimer and Cooper (2004) speculated that an explanation for the high-grade metamorphism in the Mataketake Range at this time could be movement at a plate boundary precursory to the late Cenozoic Australian-Pacific plate boundary.

Cooper and Ireland (2013) dated zircons from a biotite schist from the Pounamu Ultramafic Belt (part of the Alpine Schist) at Hokitika River east of the Alpine Fault (Fig. 6, purple dot), and identified magmatic and metamorphic ages. They obtained a metamorphic age of 71.9 ± 1.8 Ma and highlighted that this timing coincides with the pegmatite intrusion and metamorphism in the Mataketake Range described above (Chamberlain et al., 1995; Batt et al., 1999; Mortimer and Cooper, 2004). Cooper and Ireland (2013) have proposed that at 70 Ma there was a deformation event that affected the Alpine Schist.

Schulte (2011) investigated the thermal history of the footwall of the Paparoa Metamorphic Core Complex (Fig. 6), and identified a Late Cretaceous thermal event at 80–72 Ma (average age ~75 Ma) using Apatite Fission Track data (Fig. 6, blue dots). He linked this event with burial and re-exhumation that was unrelated to initial core complex development in the mid Cretaceous. Schulte (2011) speculated that Tasman Sea rifting may have triggered this activity at the Paparoa Metamorphic Core Complex. He also noted that this event was contemporaneous with activity in the extensional corridor described by Laird (1994). This corridor of extension or transtension along the Tasman Sea-LHR margin of New Zealand has been termed the “West Coast Rift” (Laird, 1981) and also referred to as the “Taranaki Rift-West Coast Rift lineament”

(King and Thrasher, 1996). It has been interpreted as the failed arm of a Late Cretaceous rift system that extended from a triple junction to the southwest of the South Island of New Zealand (Laird, 1981) or a rift transform system (King and Thrasher, 1996). To the west of the Paparoa Metamorphic Core Complex, near the sample locations of Schulte (2011), a zircon fission track age of 81.9 ± 5.8 Ma was determined for a sample of Constant/Charleston gneiss (White and Green, 1986) (Fig. 6, yellow dot).

2.3. Observations from the West Antarctic Rift System

Seafloor magnetic anomalies constrain the spreading history between the Pacific plate and West Antarctica, however determining the history of deformation between East and West Antarctica is more complex, particularly due to ice coverage and remoteness. Their relative motion history is however essential for constraining Pacific plate motion and a variety of observations and interpretations have been presented.

2.3.1. Constraining East–West Antarctica motion

Periods of relative motion between East and West Antarctica have driven intra-continental deformation and formation of the West Antarctic Rift System (WARS). The WARS separates East and West Antarctica and extends beneath the Ross Sea, Ross Ice Shelf and West Antarctic Ice Shelf (Fig. 7). Extension in the Ross Sea has led to the formation of the Victoria Land Basin, Northern Basin, Central Trough and Eastern Basin; a series of north to south trending basins (Fig. 7). The Transantarctic Mountains form the western margin of the WARS, and together they comprise one of the largest extensional provinces on Earth (Luyendyk et al., 2001), comparable in area to the Basin and Range Province in North America (Behrendt et al., 1991). Early extension occurred in the Jurassic coincident with initial Gondwanaland break-up and Ferrar magmatism that produced basaltic sills and dykes (Elliot, 1992). Rifting has continued since this time (Behrendt and Cooper, 1991), with known episodes of Cretaceous and Cenozoic extension. Evidence for extension comes from geologic and thermochronologic studies that identified cooling events in the Transantarctic Mountains linked to extension derived exhumation and denudation (e.g. Stump and Fitzgerald, 1992; Richard et al., 1994; Balestrieri et al., 1999; Balestrieri and Bigazzi, 2001; Luyendyk et al., 2001, 2003; Lisker, 2002; Rossetti et al., 2003; Siddoway et al., 2004; Storti et al., 2008), plate reconstruction models that require East–West Antarctic motion to prevent gaps between continental blocks in the southwest Pacific (e.g. Cande et al., 2000; Cande and Stock, 2004), and paleomagnetic data, namely a comparison of ~100 Ma West Antarctic paleomagnetic poles with a single synthetic pole from East Antarctica reveals ~10° of pole separation, requiring there to have been post–100 Ma relative motion (DiVenere et al., 1994; c.f. Luyendyk et al., 1996). The exact timing, magnitude and nature of East–West Antarctic extension, however, are poorly constrained for most of their history of relative motion (Siddoway, 2008), particularly before chron 24 time (~53.3 Ma) (Cande and Stock, 2004). For instance, in the Ross Sea sediments recording Paleocene or early Eocene extension remain to be directly cored and sampled, and therefore motion is indirectly inferred (Eagles et al., 2009). After chron 18 time (40.1 Ma) magnetic anomalies from the northwestern Ross Sea constrain spreading between East and West Antarctica (Cande et al., 2000; Granot et al., 2013).

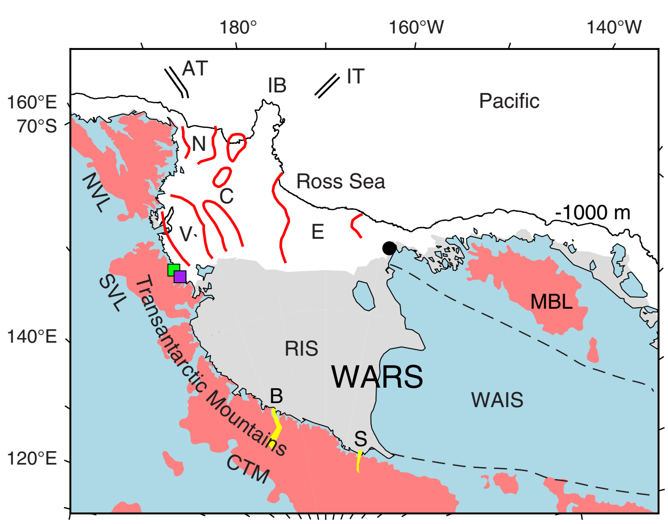

Fig. 7. Map of the southwest Pacific margin of Antarctica showing the West Antarctic Rift System (WARS), that extends beneath the Ross Sea, Ross Ice Sheet (RIS) and the West Antarctic Ice Sheet (WAIS) — dashed lines are a guide to its eastern extent. AT, Adare Trough; B, Beardmore Glacier; C, Central Trough; CTM, central Transantarctic Mountains; E, Eastern Basin; IB, Iselin Bank; IT, Iselin Trough; MBL, Marie Byrd Land; N, Northern Basin; NVL, northern Victoria Land; S, Scott Glacier; SVL, southern Victoria Land; V, Victoria Land Basin. Purple and green squares show the locations of Mt Doorly and Mt England, respectively. Filled black circle shows the location of DSDP site 270. Regions above 500 m in Antarctic are shaded light red. Topographic/bathymetric contours are from ETOPO1 (Amante and Eakins, 2009). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

The history of extension in the WARS has been divided into three major episodes (Fitzgerald, 2002) largely inferred from the cooling history of the Transantarctic Mountains, eastern Ross Sea and western Marie Byrd Land, derived from thermochronologic data (e.g. Stump and Fitzgerald, 1992; Richard et al., 1994; Balestrieri et al., 1999; Balestrieri and Bigazzi, 2001; Luyendyk et al., 2001; Lisker, 2002; Luyendyk et al., 2003; Rossetti et al., 2003; Siddoway et al., 2004; Storti et al., 2008). These episodes occurred in the early Cretaceous (~125–110 Ma), mid-Late Cretaceous (~105–80 Ma) and Eocene (~60–55 Ma). Alternatively, in the Cretaceous there may have been a single protracted event from ~120–80 Ma, centred on ~95 Ma (Decesari et al., 2007; Elliot, 2013). The main phase of extension occurred in the mid-Late Cretaceous (Lawver and Gahagan, 1994; Fitzgerald, 2002) and this phase ended contemporaneously with breakup between Marie Byrd Land and the Campbell Plateau (Lawver and Gahagan, 1994; Siddoway et al., 2004; Siddoway, 2008). During this time uplift and denudation are interpreted in parts of the Scott Glacier region and northern Victoria Land (Balestrieri et al., 1999; Balestrieri and Bigazzi, 2001) (Fig. 6). In the Scott Glacier region the amount of denudation reaches >1100 m (Stump and Fitzgerald, 1992; Fitzgerald and Stump, 1997). The main phase of uplift in the Transantarctic Mountains occurred in the Paleocene, initiating at ~60–55 Ma, and has been related to a pulse of extension at the Transantarctic Mountains front (Fitzgerald, 2002). Uplift is first recorded in northern and southern Victoria Land and later in the central Transantarctic Mountains (Fitzgerald, 2002). In southern Victoria Land, at Mt England (Fig. 7, green square) an estimated 5.5–4.5 km of uplift has occurred since 60 Ma, and at Mt Doorly (Fig. 7, purple square) an estimated 5.8–4.8 km of uplift has occurred since 55 Ma (Fitzgerald, 1992). Gleadow and Fitzgerald (1987) estimated similar amounts of uplift at Mt Doorly. During the application of fission track analysis a "break in slope" of the apatite age versus elevation profile indicates uplift and denudation occurred, however Fitzgerald (1992) points out that this age is a minimum and initial uplift may have occurred earlier. For his southern Victoria Land samples a break in slope occurs at 55 Ma for Mt England, and 50 Ma at Mt Doorly, however he suggests that uplift occurred ~5 Myr earlier and hence at 60 and 55 Ma. In the central Transantarctic Mountains, to the west of Beardmore Glacier ~7 km of uplift has occurred since ~50 Ma (Fitzgerald, 1994) (Fig. 7).

The most well-constrained period of East–West Antarctic extension occurred between ~43 and 26.5 Ma and is associated with ~180 km of spreading in the Adare Trough (Cande et al., 2000; Granot et al., 2013). The Adare Trough, in the northwestern Ross Sea, is aligned with the western Ross Sea basins and positioned to the north of the Northern Basin (Fig. 7). This period is particularly important in the development of the WARS as it is well-constrained by marine magnetic anomalies in the Adare Trough and Northern Basin (Cande et al., 2000; Granot et al., 2013). This spreading episode follows uplift that initiated at ~60–55 Ma in the Transantarctic Mountains, and the two may be related (Cande et al., 2000; Fitzgerald, 2002). Fitzgerald (2002) suggested that early extension in the Adare Trough may have propagated southwards into the Ross Embayment triggering uplift in the Transantarctic Mountains front.

Following the cessation of spreading in the Adare Trough at ~26.5 Ma Granot et al. (2010) have identified additional periods of extension and deformation. They describe a middle Miocene (17–13 Ma) change in East–West Antarctica motion associated with minimal extension in the Adare Trough, yet 5 km of extension in Northern Victoria Land and 10–15 km of extension in the Terror Rift. This event may have preceded the cessation of East–West Antarctica motion (Granot et al., 2010).

2.4. Summary of key geologic observations

2.4.1. New Caledonia to northern New Zealand region

Geologic observations to constrain plate boundary activity in the region north of New Zealand during the Late Cretaceous to Eocene are sparse and many lack tight age constraints. Subduction related rocks are minimal and restricted to the mid Cretaceous and latest Paleocene–Eocene (Cluzel et al., 2006; Meffre et al., 2012; Mortimer et al., 2012; Falloon et al., 2014). In comparison, back-arc basin opening during this time is better supported. Opening of the South Loyalty Basin during the mid or Late Cretaceous until the Paleocene is supported by microfossil dating of cherts interbedded in the obducted Poya Terrane (Cluzel et al., 2001), whereas there is a dichotomy in radiometric ages to date basin spreading with clusters at ~100–80 Ma and ~55 Ma (Prinzhofer, 1981; Spandler et al., 2005). Paleontological and radiometric dating of a possible back-arc basin to the north of the North Island reveal a wide array of ages spanning more than 70 Myr from the mid Cretaceous through to the late Eocene, the implications of which have long been scrutinized (Katz, 1976; Strong, 1976; Strong, 1980; Whattam et al., 2006). Recently dredged ~102 and 52–49 Ma subduction related rocks from the Tonga forearc have back-arc basin geochemical affinities (Meffre et al., 2012; Falloon et al., 2014).

2.4.2. Southern New Zealand

In New Zealand's South Island minor deformation is recorded during the Late Cretaceous in locations neighbouring the present-day Alpine Fault. Shear zone activity and minor magmatic intrusions in Fiordland (White and Green, 1986; Klepeis et al., 1999; Hollis et al., 2003; Marcotte et al., 2005; King et al., 2008), high-grade metamorphism and pegmatite intrusions at the Mataketake Range (Chamberlain et al., 1995; Batt et al., 1999; Mortimer and Cooper, 2004), dextral shear and mylonitisation in the Fraser Fault region (White and Green, 1986; Rattenbury, 1987, 1991) and metamorphism of the Pounamu Ultramafic Belt (Cooper and Ireland, 2013) mainly occurred between 85 and 65 Ma. Some of this activity has been speculatively linked with precursory movement on the Alpine Fault (White and Green, 1986; Mortimer and Cooper, 2004). Extension in the West Coast region and a thermal event in the Paparoa Metamorphic Core Complex also occurred during this time and have been linked with Tasman Sea spreading (Laird, 1981; Laird, 1994; King and Thrasher, 1996; Schulte, 2011).

2.4.3. West Antarctic Rift System

A major phase of uplift in the Transantarctic Mountains, western margin of the WARS, began during the Paleocene at 60–55 Ma, with 4.5 to 7 km of uplift estimated for locations in Victoria Land and the Central Transantarctic Mountains using thermochronologic data (Fitzgerald, 1992, 1994, 2002). This uplift purportedly reflects a major expisode of extension at the Transantarctic Mountain Front (Fitzgerald, 2002). The most well-constrained period of extension in the WARS occurred slightly later during the Eocene–Oligocene, at which time there was spreading in the Adare Trough. Between ~43 and 26.5 Ma there was ~180 km of spreading, constrained by marine magnetic anomalies in the Adare Trough and adjacent Northern Basin (Cande et al., 2000; Granot et al., 2013).

3. Part B: Kinematic constraints on plate boundary activity

In the following sections we investigate the kinematic consequences of using different plate circuits in the southwest Pacific for the period from 74 to 45 Ma. These results will later be discussed in light of the geologic data reviewed above, in order to better understand the broad evolution of southwest Pacific plate boundaries during this timeframe. We have chosen to focus on times younger than 74 Ma as chron 33y (~74 Ma) is the oldest magnetic anomaly that has been widely identified in the southwest Pacific (Sutherland, 2008).

3.1. Relative plate motion chains and circuits

The relative motion between two plates can be tightly constrained if they are separated by a mid-ocean ridge. Fracture zones and magnetic anomalies within the new lithosphere can be identified from marine and satellite-derived geophysical data, and these features provide direct constraints on the direction and rate of spreading. Conversely, when two plates are separated by a subduction zone or transform boundary, determining their relative motion is much more complex. Uncertainties in the amount of lithosphere that has been subducted or the amount of strike-slip motion that has occurred lead to uncertainties when constructing a relative motion history for the pair of adjacent plates. Motion between plates that are separated by zones of diffuse and poorly understood deformation, such as the WARS between East and West Antarctica, is also difficult to quantify.

Plate motion "chains" and relative plate motion "circuits" are important tools for investigating both relative and absolute plate motions on a regional or global scale. When a series of adjacent plates are separated by boundaries where motion is well-constrained, then the relative motion between any two plates in that chain of plates can be determined. Furthermore, if that plate motion chain can be linked back to Africa then the absolute motion of that plate relative to published mantle reference frames can be estimated. Absolute reference frames are typically defined through the absolute motion of Africa, since motion of this plate has been relatively slow since Pangea breakup as it has largely been surrounded by ridges (Burke and Torsvik, 2004). Its motion can be tied to the Earth's spin axis using hotspot tracks produced by plumes sourced from the deep mantle and paleomagnetic data (Müller et al., 1993; Torsvik et al., 2008). Absolute motion of plates can also be estimated using a subduction zone reference frame in which the positions of slabs in the mantle are used to longitudinally correct the position of Africa (van der Meer et al., 2012), or a reference frame based on shallow, asthenosphere-sourced hotspot trails (Cuffaro and Doglioni, 2007).

A relative plate motion circuit consists of a chain of adjacent plates that form a closed loop. They are useful for investigating relative plate motions, particularly where motion at one of the plate boundaries is poorly constrained. Where two plates are separated by a convergent boundary, an alternative mechanism to study the relative motion is to sum the motions across a series of spreading ridges that link the two plates together — their plate circuit. For example, relative motion between India and Eurasia can be determined via a plate circuit that sums motions between India, Somalia, Nubia, North America and Eurasia (Molnar and Stock, 2009). See Seton et al. (2012) and Torsvik et al. (2008) for more details on the construction of relative and absolute plate motion models, respectively.

3.2. Alternative plate circuits for the southwest Pacific

During the Late Cretaceous to mid Eocene period, the southwest Pacific comprised three spreading ridges: the Pacific–West Antarctic ridge, the Australian–East Antarctic ridge and the Australian–LHR ridge. At least one, and possibly two other plate boundaries accommodated relative plate motions during this time — within the WARS and between the LHR and the Pacific (Fig. 8). Previous studies have suggested deformation occurred within the LHR at this time leading to opening of the New

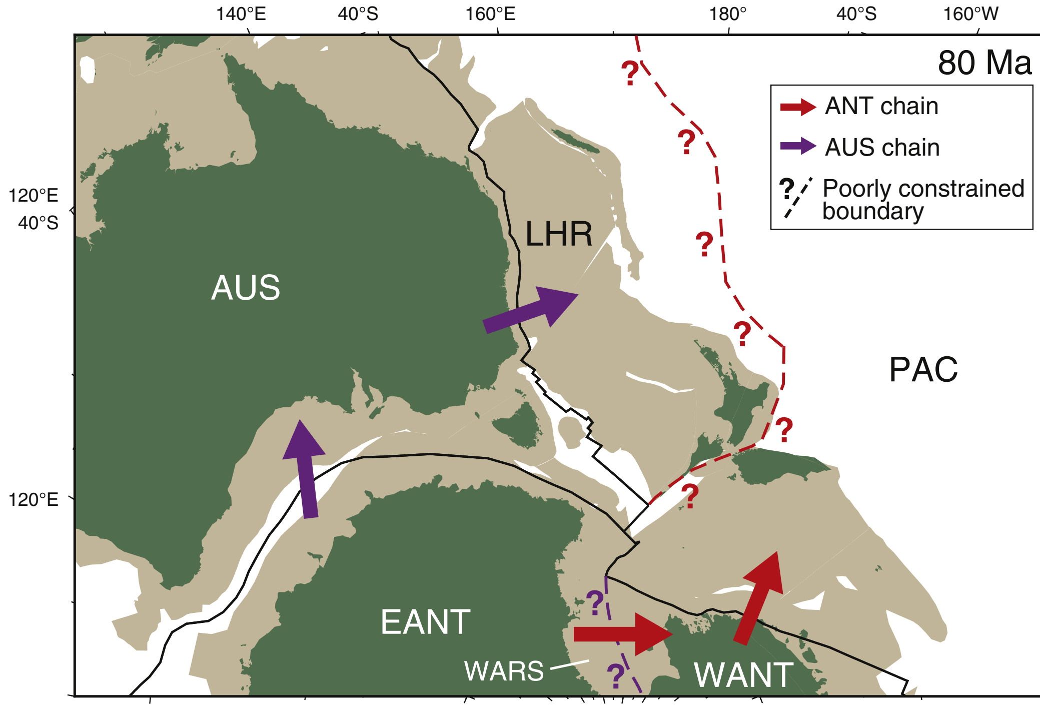

Fig. 8. Reconstruction of the southwest Pacific at 80 Ma (Seton et al., 2012). Poorly constrained Late Cretaceous–mid Eocene plate boundaries are dashed. During this time motion between East and West Antarctica in the WARS is poorly constrained, and it is uncertain when a boundary existed between the Pacific and LHR. Based on the assumption that there was no boundary between the Pacific and LHR, and due to poorly constrained motion in the WARS, an Australian plate motion chain can be used to link the Pacific plate to Africa and an absolute reference frame (purple path). Conversely, based on the assumption that a Pacific–LHR plate boundary existed and motion at this boundary is unknown, an alternative Antarctic plate motion chain can be used to reconstruct the Pacific plate (red path). Both plate motion chains connect East Antarctica to Africa via spreading at the southwest Indian ridge. AUS, Australia; EANT, East Antarctica; LHR, Lord Howe Rise; PAC, Pacific; WANT, West Antarctica; WARS, West Antarctic Rift System. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Caledonia Basin in the latest Cretaceous to early Paleocene (Lafoy et al., 2005; Schellart et al., 2006). However, more recent geophysical data suggests opening occurred prior to 85 Ma, along with the Fairway-Aotea Basin (Collot et al., 2009), and thus predates our timeframe of interest. Thus, our discussion includes a total of 5 major plates — Pacific, West Antarctic, East Antarctic, Australian and LHR plates. Together these plates comprise a regional southwest Pacific plate circuit (Fig. 9a–b). Furthermore, they can be combined to produce two separate plate motion chains that link motion of the Pacific plate to an absolute reference frame (Fig. 9c).

Motion between the LHR and Australia, and between the Pacific and West Antarctica, is well-constrained by seafloor spreading magnetic anomalies and fracture zones. The history of relative motion at the Australian–East Antarctic ridge is more strongly debated, as will be outlined below. Motion at the remaining two plate boundaries in the southwest Pacific plate circuit is poorly constrained: East–West Antarctica and Pacific–LHR (Figs. 8–9). These uncertainties largely arise from the shortage and ambiguity of geologic and geophysical data from onshore regions in New Zealand, New Caledonia and Antarctica that make it difficult to identify plate boundary activity. The

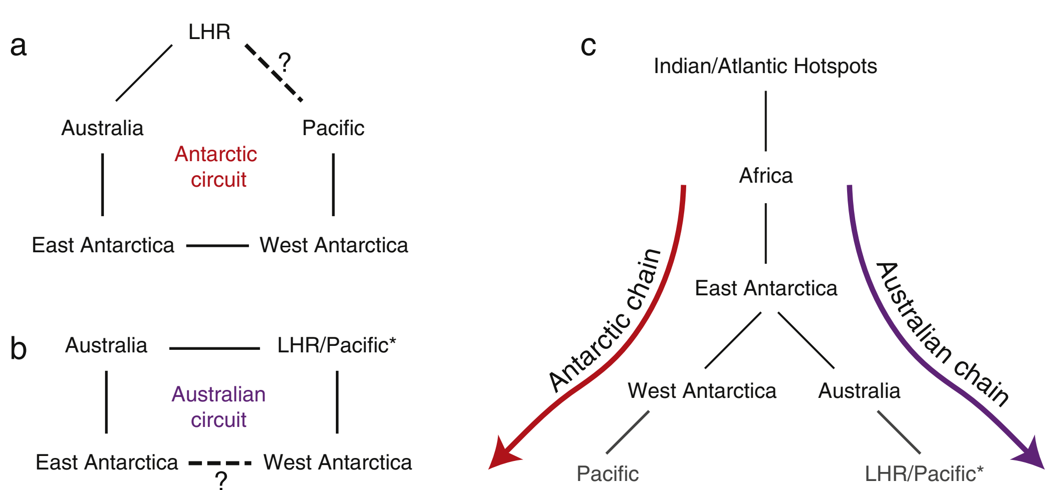

Fig. 9. Alternative southwest Pacific relative plate motion circuits (a–b), and alternative plate motion chains that can be used to reconstruct the Pacific plate relative to an absolute reference frame (c). Where the motion between two plates is poorly constrained they are connected with a dashed line (a–b). *In the Australian circuit and chain it is assumed that there is no plate boundary between the LHR and Pacific plate.

quantity and quality of geophysical data available from offshore areas also presents a problem for determining tightly constrained relative motion histories. Due to its remoteness and inaccessibility there is a paucity of geophysical data in Antarctica and the Ross Sea that can be used to constrain motion in the WARS. Additionally, ice coverage reduces the quality of gravity data derived from satellite altimetry (e.g. McAdoo and Laxon, 1997). Data coverage in parts of the southwest Pacific is also sparse and several back-arc basins have been entirely or partially subducted, thereby completely eliminating the seafloor spreading record. Uncertainty in the timing and magnitude of East–West Antarctic motion, and whether the LHR and Pacific plate were separated by one or several plate boundaries fuel the debate over how Pacific motion is best quantified and how it should be reconstructed using plate circuits during the Late Cretaceous to mid Eocene.

3.2.1. Constraints on southwest Pacific spreading histories

East Antarctica–Australia. Differing interpretations of marine geophysical data and correlations between conjugate tectonic structures have resulted in alternative spreading histories for the Southeast Indian Ridge, particularly for times older than ~50 Ma (Royer and Rollet, 1997; Tikku and Cande, 1999, 2000; Whittaker et al., 2007, 2013). Initiation of seafloor spreading between Australia and Antarctica was diachronous, with the breakup inferred to occur around 91–87 Ma between Naturaliste Plateau and Bruce Rise, 83–71 Ma in the Bight Basin-Wilkes Land Sector, 68–63 Ma in the Otway Basin-Terre Adelie sector, and 58–49 Ma in the Sorell Basin (Direen et al., 2012). Initial seafloor spreading proceeded at slow spreading rates, and the lack of clear fracture zones within the seafloor formed prior to chron 18o (40.1 Ma) has lead to different interpretations of the spreading history during the Late Cretaceous–Early Paleocene (Royer and Rollet, 1997; Tikku and Cande, 1997; Tikku and Cande, 1999; Whittaker et al., 2007, 2013).

Reconstructions proposed by Royer and Rollet (1997) produce a good fit between the South Tasman Rise and Cape Adare, but lead to misfits of magnetic anomalies in the Bight Basin, and an unreasonably large overlap between the Broken Ridge and Kerguelen Plateau in the western part of the Australia–Antarctica plate boundary system. Tikku and Cande (1999, 2000) proposed an alternative reconstruction that produces a better fit between Kerguelen and Broken Ridge, but also results in large overlaps between Tasmania and Cape Adare in the Late Cretaceous. Tikku and Cande (2000) invoked large Late Cretaceous strike-slip motion between Tasmania and Australia to resolve these overlaps, but geologic and geophysical observations discount this possibility. Whittaker et al. (2007) invoked a major change in the direction of Australia–Antarctica relative motions prior to chron 20o, but these rotation parameters result in tectonically improbable episodes of extension and compression between Broken Ridge and Kerguelen. Rotation parameters proposed by Whittaker et al. (2013) reconcile available constraints for the entire Australia–Antarctica plate-boundary system through the Late Cretaceous–Early Paleocene. A general limitation on all these reconstructions is the uncertainty over whether the earliest identified magnetic anomalies in the Australia–Antarctic Basin are truly isochrons (e.g. Tikku and Cande, 1999; Direen et al., 2007). This uncertainty is significant for chron 34y which corresponds to peridotite ridges within the continent–ocean transition. Younger anomalies are interpreted to lie within oceanic crust from interpretation of geophysical data (Leitchenkov et al., 2007; Close et al., 2009) and a new compilation of magnetic data for the Antarctic margin (Golynsky et al., 2013).

Pacific–West Antarctica. Despite the sparse data coverage in the region, multiple ship, airborne and satellite datasets have been combined to produce plate reconstruction models for the evolution of the Pacific margin of West Antarctica (e.g. Stock and Molnar, 1987; Cande et al., 1995; Larter et al., 2002; Eagles et al., 2004a; Wobbe et al., 2012). Pacific–West Antarctic breakup began between the Chatham Rise and Marie Byrd Land at ~90 Ma to form the Bellingshausen Sea (Larter et al., 2002). Chron 34y (83 Ma) is the oldest magnetic isochron that has been identified from spreading in this sector (Larter et al., 2002; Wobbe et al., 2012), and based on the backwards extrapolating the chrons 33–34 spreading rate an age of 90 Ma was obtained for breakup (Larter et al., 2002). Breakup apparently propagated westward, with spreading between the Campbell Plateau and Marie Byrd Land occurring later to form the Amundsen Sea (Larter et al., 2002). The oldest magnetic anomaly identified in this sector is chron 33n (73.6–79.5 Ma).

Stock and Molnar (1987) identified the independent Late Cretaceous Bellingshausen plate to the north of Marie Byrd Land, between the Marie Byrd Land seamounts and the De Gerlache gravity anomaly, that formed along a section of the Pacific–West Antarctic spreading ridge. The Bellingshausen plate moved as an independent plate between about chrons 33o and 27o times (~79–61 Ma), with its cessation coinciding with a south Pacific plate boundary reorganization (Cande et al., 1995; Larter et al., 2002; Eagles et al., 2004b). Relative motion between the Pacific and Bellingshausen plates at the Pacific–Bellingshausen spreading ridge is constrained by magnetic anomalies, however a zone of deformation existed at the Bellingshausen–Marie Byrd Land plate boundary that may have been up to 670 km wide (Wobbe et al., 2012). As a result of independent Bellingshausen plate motion and formation of a wide zone of deformation, Pacific motion can only be directly tied to West Antarctica between the Campbell Plateau and Marie Byrd Land until chron 27 time (61 Ma), away from the Bellingshausen plate (Stock and Molnar, 1987). Along this part of the Pacific–Marie Byrd Land margin to the east of the Bellingshausen plate the oldest magnetic anomaly picks are chron 33o (80 Ma), however there are a very limited number of picks of this age (Wobbe et al., 2012).

East Antarctica–West Antarctica. Mesozoic–Cenozoic motion between East and West Antarctica has lead to formation of the WARS (Fig. 7), but the details of this motion are poorly constrained. Paleomagnetic data, crustal extension estimates and subsidence analysis indicate between 300 and 700 km of extension since the late Early Cretaceous (DiVenere et al., 1994; Luyendyk et al., 1996; Decesari et al., 2007; Wilson and Luyendyk, 2009). However, much of the extension in the WARS likely occurred during the Early Cretaceous rifting of the East Gondwana margin, and prior to breakup between Antarctica and Zealandia, as supported by geologic observations from Marie Byrd Land (Siddoway, 2008; McFadden et al., 2010; Saito et al., 2013) and from geologic samples recovered by dredging/drilling within the Ross Sea (Siddoway et al., 2004).

The most tightly constrained period of East–West Antarctica extension occurred in the mid Eocene–Oligocene and is associated with ~180 km of spreading in the Adare Trough (Fig. 7) between 43 and 26 Ma (Cande et al., 2000). From an analysis of ship based magnetic, gravity and bathymetry data from the Adare Trough Cande et al. (2000) identified a fossil spreading centre and anomalies 18–9. They interpreted a history of extension throughout the rift system that started slightly earlier than anomaly 18, from about anomaly 20 time (43 Ma). Cande and Stock (2004) extended the period of extension by adding 50 km of additional motion between 53 and 61 Ma to close basins in the northern Ross Sea between Cape Adare and Hallett Ridge (Cande and Stock, 2004). They suggested that any additional motion must have occurred in this region as there is no deformation of the isochrons on the neighbouring Southeast Indian Ridge and Pacific–Antarctic Ridge. During this time period there may also have been counterclockwise rotation of the Iselin Bank in the northern Ross Sea to form the Iselin Trough (a graben structure) (Cande and Stock, 2004) (Fig. 7). This motion is required to align magnetic anomalies and strands of the Emerald Fracture Zone on the Pacific and Antarctic plates (Cande and Stock, 2004).

The pivotal work of Cande et al. (2000) was the first to compute a finite rotation pole for chron 13o (33.6 Ma) but the limited dataset used yielded a large uncertainty in the pole position (~5000 km). Rotations for earlier phases of E–W Antarctica motion (Cande et al., 2000; Cande and Stock, 2004) were estimated using the chron 13o pole position, but incrementally larger rotation angles. Several subsequent studies have predicted more strike-slip to convergent Cenozoic motion in the central and eastern rift system. Davey et al. (2006) computed a pole of rotation for anomaly 18o constrained by extension in the Victoria Land Basin (95 km) and Adare Trough magnetic anomalies (Cande et al., 2000). Their anomaly 18o pole of rotation is located in the central part of the rift system and produces convergence in the eastern rift system, contemporaneous with extension in the Ross Sea (Davey et al., 2006). Most recently, Granot et al. (2013) interpreted aeromagnetic data from the Adare Trough and Northern Basin and computed four rotations from the period 40–26.5 Ma (anomalies 18o, 16y, 13o, 12o). Their kinematic model predicts extension in the Adare Basin (~140 km), dextral transcurrent motion in the region of the Ross Ice Shelf (~35–50 km) and oblique convergence further east in the rift system (~160 km, Antarctic Peninsula) due to a pole of motion in the central part of the rift system. The poles of rotation derived by Granot et al. (2013) from magnetic anomalies agree well with those predicted by Davey et al. (2006) from geologic evidence, and represent the first well-constrained model of Eocene–Oligocene kinematics between East and West Antarctica.

Other studies have estimated East–West Antarctica motion which to some extent rely on plate circuit calculations. Wilson and Luyendyk (2009) produced a model for WARS extension that extends back to ~100 Ma in order to study Antarctic paleotopography and ice sheet growth. Their model incorporates previously published estimates for the amounts and orientations of extension in the rift system, and yields extension in the Ross Sea and dextral transcurrent to dextral extensional motion in the central and eastern rift system during the entire period. They model >700 km of extension in the Ross Sea since 105 Ma, including >600 km during the period preceding Adare Trough spreading. Sutherland (2008) derived pre-mid Eocene (pre-Adare Trough spreading) poles of rotation for West Antarctica motion at 74 and 56 Ma by fitting rift margins south of New Zealand. From 74 to 56 Ma his model predicts very oblique sinistral extension in the Ross Sea, sinistral strike-slip motion in the central part of the rift system and convergence in the eastern part of the rift system. From 56 to 44 Ma extension in the Ross Sea becomes more orthogonal and motion in the central part of the rift system is minimal. Overall Sutherland's (2008) model predicts 300 km of extension in the Ross Sea since 74 Ma.

Australia–LHR. A robust spreading history for the Tasman Sea has benefitted from several comprehensive studies of satellite and ship-derived marine geophysical data including magnetic anomalies and fracture zones (Hayes and Ringis, 1973; Weissel and Hayes, 1977; Shaw, 1979; Gaina et al., 1998). The most recent of these models by Gaina et al. (1998) is well-accepted and widely used. They revised magnetic anomaly and fracture zone interpretations and presented a quantitative model, inclusive of error uncertainty, for the period from 52–74 Ma. For times older than chron 33y (73.6 Ma), their model is qualitative and based on a lack of well defined fracture zone and magnetic anomaly picks they were unable to compute finite rotations with associated uncertainty. Tasman Sea rifting started at ~90 Ma, with seafloor spreading beginning before chron 34y time (83 Ma). Spreading stopped at ~52 Ma. Gaina et al. (1998) incorporated 13 continental blocks in their Tasman Sea model, indicating a complex spreading history, and that the early stages of Tasman Sea opening did not strictly operate as a simple two plate system.

3.3. Alternative Pacific plate circuits and their implications for relative plate motions

Using the plate reconstruction software *GPlates* (Boyden et al., 2011) we investigate how adopting different versions of the southwest Pacific plate circuit (Fig. 9a–b) influences Late Cretaceous to mid Eocene relative motions between the LHR and Pacific, and East and West Antarctic plates. For all our plate circuit tests we adopt the widely used and well accepted spreading history of Gaina et al. (1998) for Australia–LHR motion. For the remaining spreading systems we test alternative spreading histories as shown in Tables 2 and 4. We adopt the rotations used in published plate circuits, spreading histories from recent studies, as well as end-member reconstructions to allow for a wide spread of scenarios to be analysed.

Our workflow allows us to investigate the implied amounts of intra-continental deformation within New Zealand and in the WARS, and convergence or divergence to the north of New Zealand, resulting from the different plate circuits. We refer to the plate circuit that allows motion between the Pacific and LHR as the “Antarctic plate circuit”, and models that incorporate a four-plate system in which the LHR is part of the Pacific plate as the “Australian plate circuit” after Doubrovine et al. (2012).

The Antarctic plate circuit leaves relative motion between the LHR and Pacific unconstrained and assumes a plate boundary separated the two plates (Fig. 9a). This circuit yields a prediction of the motion through New Zealand and across a plate boundary to the east of the LHR. The predicted motion depends on estimates of motion between East and West Antarctic, Pacific and West Antarctica, and between East Antarctica and Australia. We test different estimates of Late Cretaceous–Early Cenozoic motion between East and West Antarctica. We test six Antarctic circuits in total, summarized in Table 2, including the preferred plate circuits of Seton et al. (2012), Sutherland (2008) and Schellart et al. (2006). In addition to previously published East–West Antartica motions, we derive a further set of rotations for the period from 85–45 Ma. These rotations use the chron 18o (40.1 Ma) finite pole of rotation from Granot et al. (2013), with the angle increased for older times (Table 3) to minimize relative motion within New Zealand (c.f. Cande and Stock, 2004; Sutherland, 2008).

In the Australian plate circuit, there is no plate boundary between the Pacific and LHR (Steinberger et al., 2004; Doubrovine et al., 2012) (Fig. 9b). This circuit leaves relative motion between East and West Antarctica unconstrained and yields a prediction of the motion within the WARS. The predicted motion depends on the different interpreted

Table 2
Alternative Antarctic plate circuits and incorporated spreading histories.

Circuit	Reference for poles of rotation		
	EANT–AUS	EANT–WANT	WANT–PAC
AntC1	Whittaker et al. (2013)	Cande et al. (2000)[a]	Larter et al. (2002)
AntC2[b]	Whittaker et al. (2013)	Cande and Stock (2004)	Larter et al. (2002)
AntC3	Whittaker et al. (2013)	Granot et al. (2013)[a]	Larter et al. (2002)
AntC4	Whittaker et al. (2013)	This study and Granot et al. (2013)[c]	Larter et al. (2002)
AntC5[d]	Royer and Rollet (1997); Royer and Sandwell (1989)	Cande et al. (2000)[a]	Cande et al. (1995)
AntC6[e]	Tikku and Cande (2000)	Sutherland (2008)	Larter et al. (2002)

[a] No East–West Antarctica motion before opening of the Adare Trough at ~43–40 Ma.
[b] Preferred plate circuit of Seton et al. (2012).
[c] Angle of opening of Granot et al.'s (2013) anomaly 18o pole of rotation is extended to minimize motion within New Zealand (Table 3), see main text for details.
[d] Preferred plate circuit of Schellart et al. (2006).
[e] Preferred plate circuit of Sutherland (2008).

Table 3
Finite poles of rotation for West Antarctica relative to East Antarctica prior to 40 Ma.

Chron	Age (Ma)	Latitude (°)	Longitude (°)	Angle (°)
18o[a]	40.13	−85.87	−139.51	4.48
24.3o	53.3	−85.87	−139.51	4.48
27o	61.3	−85.87	−139.51	6.2
29o	64.7	−85.87	−139.51	7.0
32.1y	71.1	−85.87	−139.51	7.5
–	100.0	−78.76	−46.5	7.72

[a] From Granot et al. (2013).

spreading histories for the Southeast Indian Ocean (Table 4). For our Australian plate circuit tests we fix the Pacific plate to the LHR from 74 to 45 Ma (Steinberger et al., 2004; Doubrovine et al., 2012). We also test a scenario in which the Pacific Plate is only fixed to the LHR between 83 and 52 Ma to coincide with cessation of Tasman Sea spreading.

We analyse plate motions predicted by different Antarctic and Australian plate circuits by dividing the Late Cretaceous to mid Eocene timeframe into two periods: 74–52 Ma and 52–45 Ma. From about 74 to 52 Ma New Zealand was purportedly tectonically quiescent (Sutherland, 2008) and the period 52 to 45 Ma follows cessation of spreading in the Tasman Sea (Gaina et al., 1998).

To test the Antarctic circuits we compute the relative plate motion paths for the Pacific plate relative to the LHR at seven points along the western edge of the Pacific plate from 74 to 45 Ma (Figs. 10–11). In this scenario the Pacific plate is reconstructed through time relative to a fixed LHR in order to isolate their relative motion history. To test the Australian circuits we compute relative plate motion paths for West Antarctica relative to East Antarctica at 4 points in the West Antarctic Rift System, again from 74 to 45 Ma (Fig. 12). This was repeated for the test case with a cross-over to the Antarctic circuit at 52 Ma (Fig. 13).

In addition to computing 74–45 Ma relative plate motion paths, we also compute relative plate motions (Pacific–LHR and West Antarctica–East Antarctica) individually for each time interval at fixed points along the inferred plate boundaries (Figs. 14–16). For example, in the case of the Australian circuit we selected 4 points in the WARS at present-day and determine where those same points would have travelled over each of the three time periods of investigation, assuming they were on the West Antarctic Plate. For each time interval we compute the motion of the same points. By calculating the motion at the same points for each time interval we are able to directly and clearly compare the direction and amount of motion predicted for each circuit for each period simultaneously.

3.3.1. Antarctic plate circuit results

The Antarctic plate circuits produce overall oblique convergence between the Pacific plate and the LHR during the period 74–45 Ma, except for model AntC6 that produces divergence (Figs. 10–11 and 14). Motion changes from dextral (74–52 Ma) to sinistral (52–45 Ma), except during the period 74–52 Ma when orthogonal convergence is predicted by AntC1–3 and AntC5 to the north of New Zealand, and during the period 52–45 Ma AntC6 predicts orthogonal divergence to the north of New Zealand (Fig. 11). The amount and orientation of motion is clearly time-dependent and differs significantly between the circuits tested, indicating the importance of using well-constrained spreading histories for each plate pair within a circuit. The magnitude of convergence/divergence tends to be minimal to the south of New Zealand and increases in a northerly direction, reflecting a local pole of rotation.

Motion within New Zealand. From 74 to 52 Ma AntC2 and AntC4 (Fig. 14b, black and cyan lines, respectively) produce notably very little relative motion between the Pacific and LHR, <50 km. AntC4 motion, designed to minimize deformation within New Zealand, predicts <50 km of strike-slip motion along the entire length of New Zealand. The motion produced by AntC2 is comparable in magnitude to AntC4, however the orientation of motion becomes more orthogonally convergent in the north. During this period the models that do not incorporate extension between East and West Antarctica produce the most convergence (AntC1,3,5), and it should be noted that the motion associated with AntC5 is only for the period 65–52 Ma, 9 Myr less than the other circuits. AntC6 produces a similar amount of relative motion to AntC1, AntC3 and AntC5, however the motion is strike-slip to obliquely divergent in orientation (Fig. 13b, blue line).

From 52–45 Ma (Fig. 14c) AntC1–4 produce a very similar pattern of relative motion that is dominantly strike-slip. AntC5 produces oblique convergence, while AntC6 produces oblique divergent motion during this period (Fig. 14c, purple and blue lines, respectively).

Table 5 shows the amount of motion, orthogonal to a boundary paralleling the Alpine Fault, for an arbitrary point on the southern Campbell Plateau implied by different models. We assume that if a plate boundary transected New Zealand at this time it would likely have followed the trend of the present-day Alpine Fault. We consider this the most reasonable approximation as it has been proposed that the Alpine Fault formed at a major lithospheric discontinuity that may date back to the Paleozoic (Sutherland et al., 2000). During the period from 74–52 Ma, when New Zealand is thought to have been tectonically quiet, AntC2, 4 and 6 predict very little motion. These circuits incorporate East–West Antarctica motion prior to Adare Trough spreading at ~43–40 Ma. For instance AntC4 predicts only 10 km of extension, compared to AntC3 that predicts 95 km of convergence.

The alternative plate circuits predict different locations of the Campbell Plateau, Chatham Rise and eastern block of the South Island of New Zealand (part of the Pacific plate) at 45 Ma (Fig. 10). The difference is largely related to different East–West Antarctica rotations. AntC1, AntC2, AntC5 and AntC6, which use the East–West Antarctica rotations from Cande et al. (2000) leave a gap between the Campbell Plateau and the western block of the South Island of New Zealand of 80–120 km to the south of where the present-day Marlborough Fault System stems from the Alpine Fault (Fig. 10a–b, e–f). AntC3 and AntC4, using the well-constrained East–West Antarctica motions from Granot et al. (2013), produce alignment along this portion of the margin and overlap further north (Fig. 10c–d). Considering that rifting between the Challenger and Campbell plateaus initiated at 45 ± 5 Ma, some overlap of rigid continental blocks is expected.

Motion north of New Zealand. During the period 74–52 Ma (Fig. 14e) models AntC1–3 and AntC5 predict orthogonal convergence. Circuits AntC4 and AntC6 are the exceptions to this and produce very different patterns of motion. AntC4 produces very little motion (<50 km) that is very oblique in orientation (Fig. 14e, cyan line), while AntC6 on the other hand produces strike-slip to divergent motion (~120 km) (Fig. 14e, blue line). Both AntC4 and AntC6 were designed to minimize relative motion in New Zealand during a period of apparent tectonic quiescence, and imply that little convergence occurs between the LHR and Pacific plates during the same time period.

From 52 to 45 Ma (Fig. 14f) AntC1–5 produce very oblique motion. Model AntC6 predicts up to 100 km of extension, with the Pacific plate moving orthogonally away from the LHR (Fig. 14f, blue line).

Table 5 shows the overall amount of motion orthogonal to a boundary paralleling the LHR. Given there is so much uncertainty over the existence

Table 4
Alternative Australian plate circuits and incorporated spreading histories.

Circuit	EANT–AUS rotations	EANT–WANT rotations used to compute cross-overs
AusC1	Whittaker et al. (2013)	Cande and Stock (2004)
AusC2	Whittaker et al. (2013)	This study and Granot et al. (2013)[a]
AusC3	Tikku and Cande (2000)	Cande and Stock (2004)
AusC4	Tikku and Cande (2000)	This study and Granot et al. (2013)[a]

[a] Angle of opening of Granot et al.'s (2013) anomaly 18o pole of rotation is extended to minimize motion within New Zealand (Table 3), see main text for details.

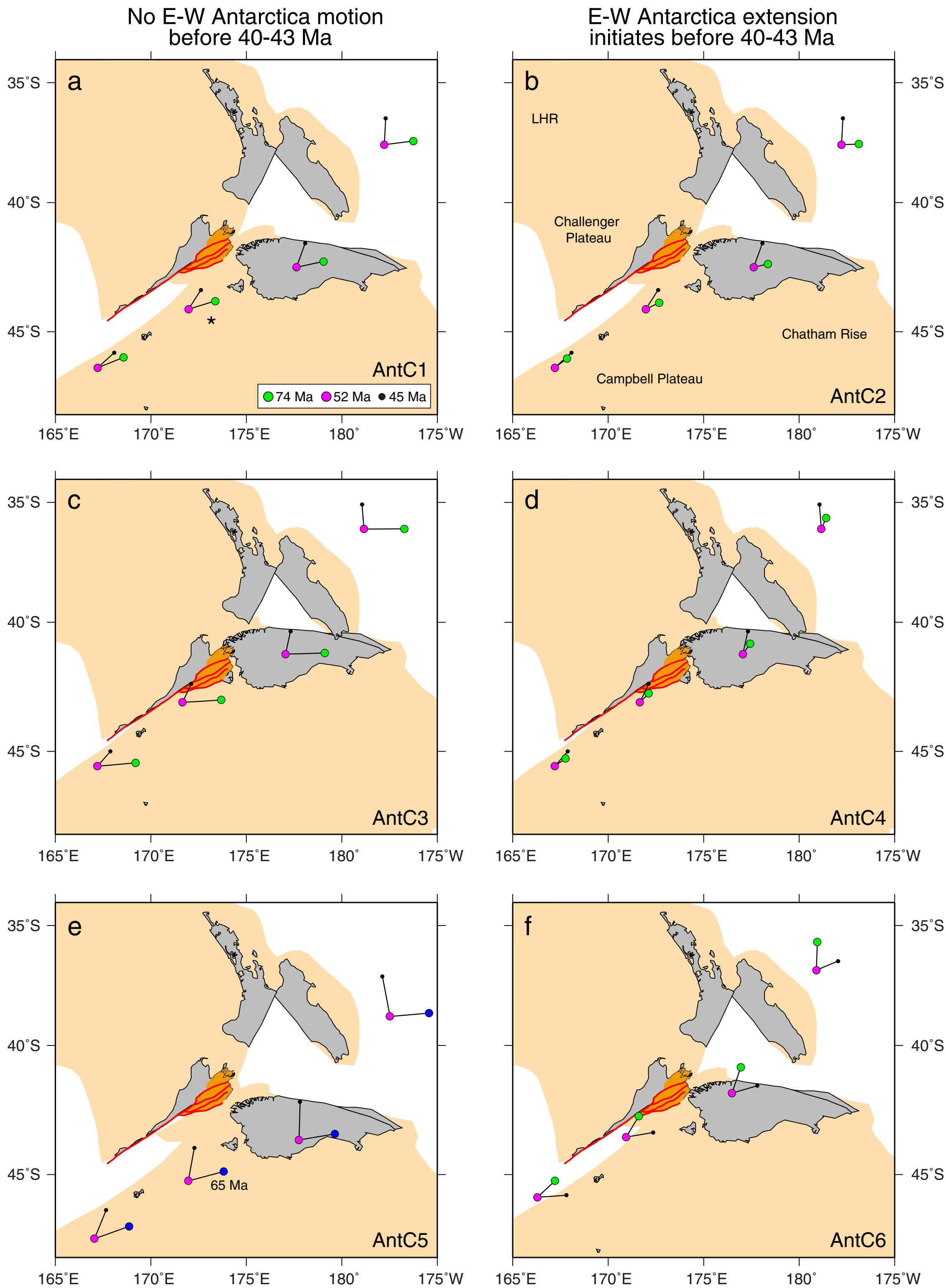

Fig. 10. Southwest Pacific reconstructions at 45 Ma in a fixed LHR reference frame. Motion paths are shown for the Pacific plate based on different Antarctic plate circuits (AntC1–6, see Table 2). Present-day coastlines (gray) and continental crust (yellow) have been reconstructed. The present day traces of the Alpine Fault and Marlborough Fault System are red. The region of the Marlborough Fault system is shaded orange. Although this area has been highly deformed, the eastern block of the South Island at this time has been rigidly reconstructed. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

of this boundary, we consider this the simplest and most reasonable orientation for a LHR-Pacific plate boundary if it existed. Furthermore long-lived subduction beneath Eastern Gondwanaland that ended in the mid Cretaceous is also commonly modeled as paralleling the continental margin (Veevers, 2000; Matthews et al., 2011; Seton et al., 2012). During the period from 74–52 Ma when New Zealand was tectonically quiet AntC4 predicts only 15 km of convergence. On the other hand AntC1–3 predict 135–225 km of convergence, more than 9 times the amount of AntC4. AntC6 also predicts very little motion, with 35 km of divergence.

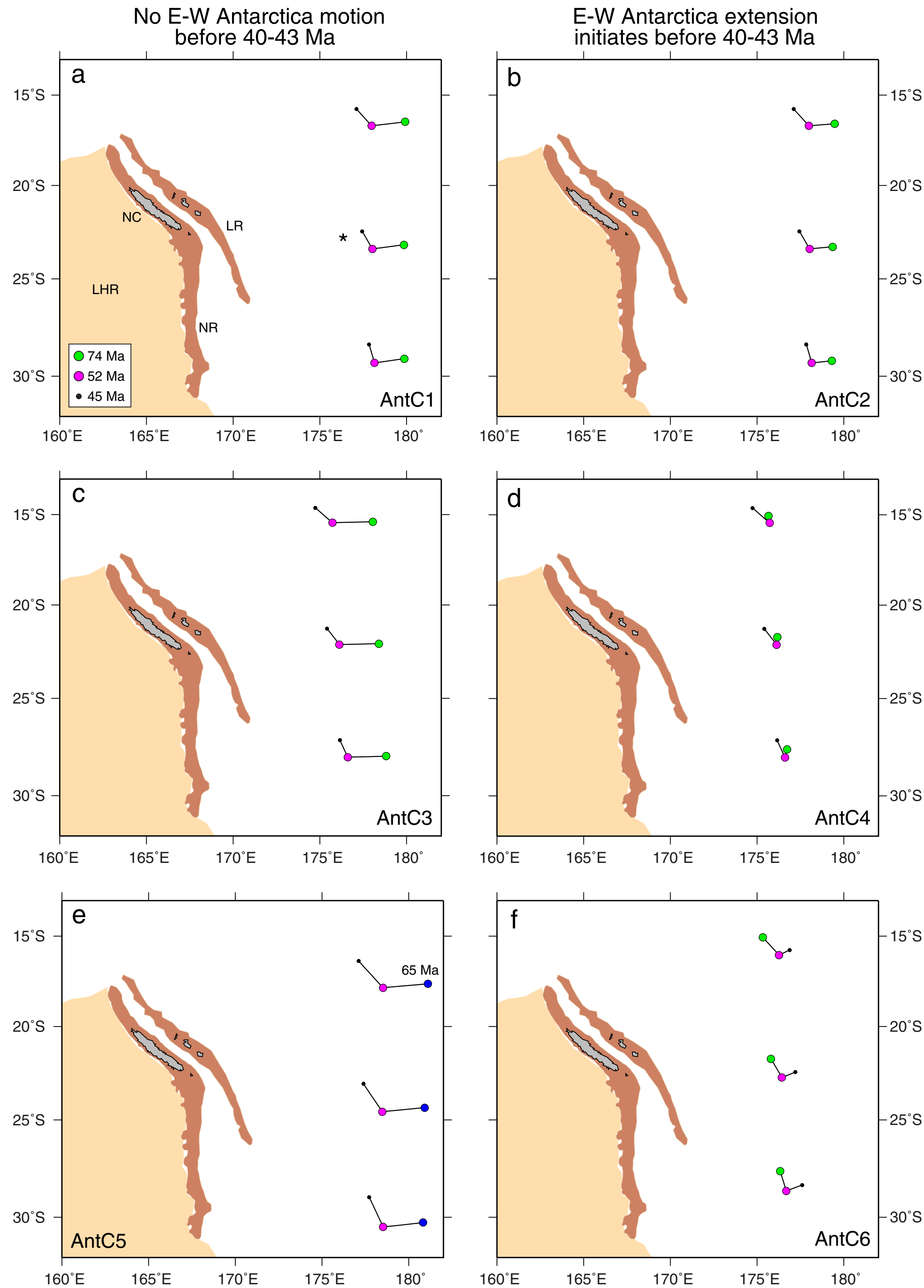

Fig. 11. Southwest Pacific reconstructions at 45 Ma in a fixed LHR reference frame, focusing on the region to the north of New Zealand. Motion paths are shown for the Pacific plate based on different Antarctic plate circuits (AntC1–6, see Table 2). Present-day coastlines (gray), continental crust (yellow) and the Norfolk (NR) and Loyalty (LR) ridges have been reconstructed. NC, New Caledonia. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

3.3.2. Australian plate circuit results

The Australian plate circuits predict a history with episodes of both extension and compression between East and West Antarctica (Figs. 12 and 15). For all models using this circuit the motion within the WARS is predicted to be net extension from 74 to 52 Ma (50–250 km) followed by net compression during the period 52 to 45 Ma (100–200 km).

During the period 74–52 Ma (Fig. 15b) the Australian plate circuits predict extension in the WARS, and the magnitudes and orientations of

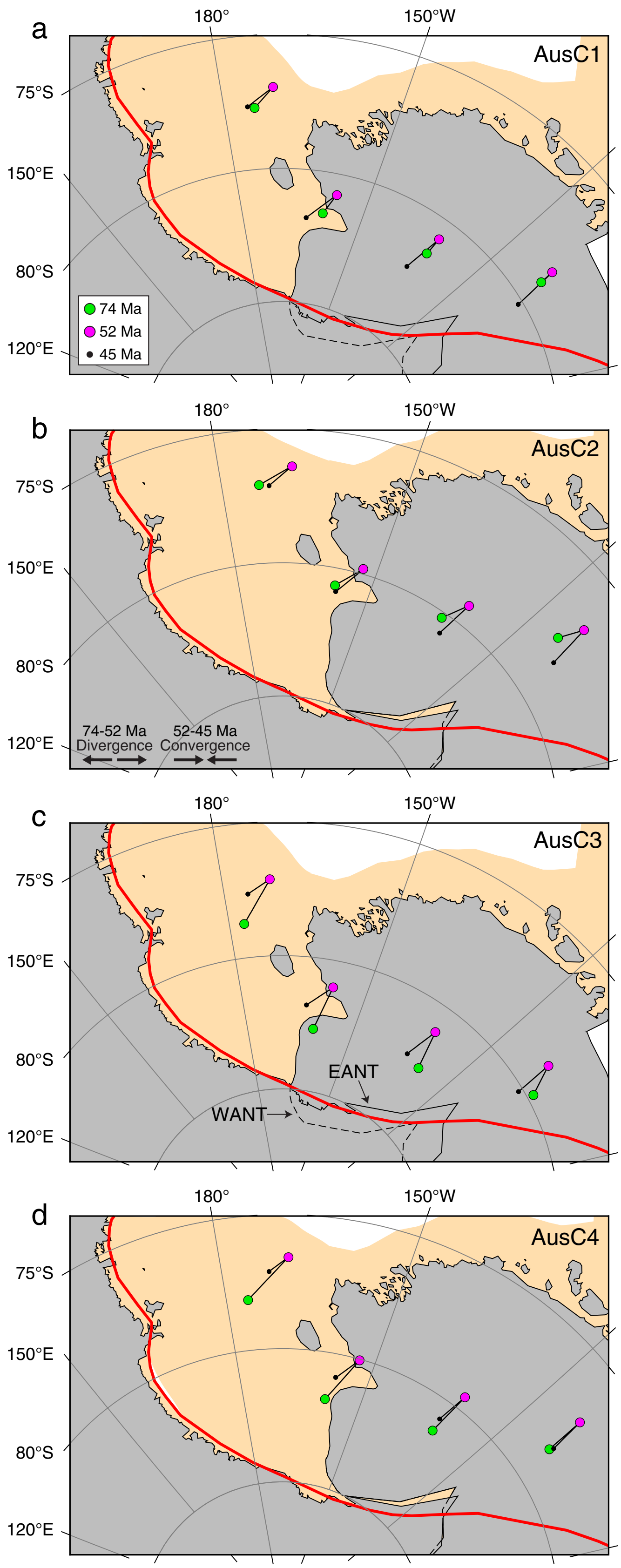

Fig. 12. Antarctica-centred reconstructions at 45 Ma in a fixed East Antarctica reference frame. Motion paths are shown for the West Antarctic plate based on different Australian plate circuits (AusC1–4, see Table 4). Present-day coastlines (gray) and continental crust (yellow) have been reconstructed. The approximate trace of the southern margin of the WARS is red. EANT, East Antarctica; WANT, West Antarctica. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

motion are more variable compared to the earlier and later periods. The models that adopt Tikku and Cande's (2000) Australia–East Antarctica rotations predict 150 to 250 km of extension, while the models that adopt Whittaker et al.'s (2013) rotations predict only 50–150 km of extension. For each of the circuits, extension is larger in the Ross Sea and decreases in an easterly direction away from the Ross Embayment. AusC2 that adopts Whittaker et al.'s (2013) spreading history for Australia–East Antarctica results in orthogonal extension in the Ross Embayment that transitions to more oblique (dextral) motion further east to the south of Marie Byrd Land. When the Australian circuits are only implemented for the period from 74 to 52 Ma, then East–West Antarctica relative motion is reduced by ~25 km (Figs. 13 and 16).

From 52 to 45 Ma AusC1–4 produce very similar patterns of relative motion that are identical in orientation and only differ by <50 km magnitude (Fig. 15). During this interval the amount of motion is smallest in the Ross Sea.

4. Discussion

4.1. Southwest Pacific plate boundaries during the Late Cretaceous to mid Eocene

An important and unresolved question concerning the tectonic evolution of the southwest Pacific is whether or not there was relative motion and plate boundary activity between the Pacific and LHR from the Late Cretaceous to mid Eocene, prior to the widely accepted onset of Tonga–Kermadec subduction from at least ~45 Ma (e.g. Bloomer et al., 1995), and following long-lived subduction beneath Eastern Gondwanaland from the Carboniferous to at least the mid Cretaceous (e.g. Leitch, 1975; Paris, 1981; Campbell, 1984; McPhie, 1987; Mortimer et al., 1999; Roser et al., 2002; Adams et al., 2009). Many regional plate reconstruction models incorporate subduction during the entire period (Fig. 2, Model 1) (Crawford et al., 2003; Sdrolias et al., 2003, 2004; Schellart et al., 2006; Whattam et al., 2008; Cluzel et al., 2010a, 2012a,b). While the details of these kinematic histories differ between the various studies, they share the implication that the LHR cannot have been part of the Pacific plate. This is contrary to the assumption within global geodynamic studies that have shown that when the Pacific and LHR are assumed to form a single plate before 45 Ma (Fig. 2, Model 2), and they are reconstructed using a plate motion chain through Australia, then both Pacific and Indo-Atlantic hotspot trails can be better approximated (Steinberger et al., 2004; Doubrovine et al., 2012). In particular, the path of the Hawaiian–Emperor seamount chain can be reproduced back to 65 Ma; this includes the enigmatic ~50–47 Ma bend, which is better represented assuming no relative motion between LHR and the Pacific before 45 Ma (Steinberger et al., 2004).

4.1.1. Late Cretaceous–Eocene subduction zone processes east of the LHR

Geologic observations from New Caledonia indicate a subduction initiation event occurred at, or just before 55 Ma to the east of New Caledonia. Pre-obduction dykes, including boninitic dykes were erupted during the period 53.1 ± 1.6 to 49.6 ± 2.8 Ma (Cluzel et al., 2006). Amphibolite lenses dated at ~55.8 ± 1.7 Ma have been identified at the base of the Peridotite Nappe indicating high temperature metamorphism (Cluzel et al., 2012a). Recently, 51.2 and 49.5 Ma island arc-type tholeiites, and 51.2 and 49.8 Ma plagiogranites were dredged from the Tonga forearc and related to the early stages of subduction; either Tonga–Kermadec subduction or a northeast-dipping subduction zone adjacent to New Caledonia (Meffre et al., 2012). At least one plate boundary, a subduction zone, must therefore have separated the Pacific plate from the LHR since 55 Ma.

Direct geologic evidence to support active subduction east of the LHR immediately prior to this ~55 Ma subduction initiation event is limited to a single boninite from southeastern North Island, New Zealand (Kopi Boninite, Moore, 1980), dated at 85 ± 10 Ma (Mortimer et al., 2012). This is in strong contrast to the post-Eocene and pre-mid

Cretaceous phases of well-accepted subduction from which arc or arc related rocks are abundant.

Subduction is commonly associated with arc volcanism driven by slab dehydration and melting. The lack of arc magmatism in the southwest Pacific does not preclude subduction during this period, and this has been discussed by Schellart et al. (2006) and Crawford et al. (2003) and compared to examples in the Mediterranean and Southeast Asia. They suggest that subduction driven by rapid slab rollback does not always produce large volumes of arc magmatism, rather magmatism can be suppressed. For instance at the Hellenic, Calabrian and Betic-Rif arcs, where subduction is dominated by slab rollback, arc volcanism is limited and restricted to only a small number of isolated volcanoes (Schellart et al., 2006), and during opening of the Parece Vela-Shikoku back-arc basin magmatism at the Mariana Arc waned (Crawford et al., 1981). Cluzel et al. (2010a) also hypothesized that in the southwest Pacific strong eastward-directed aesthenospheric flow driven by mid to Late Cretaceous extension in Eastern Gondwanaland, followed by eastward slab rollback during the Late Cretaceous and early Cenozoic, may have hindered normal corner flow and suppressed the development of an arc. Schellart et al. (2006) noted that traces of arc rocks may have been largely destroyed during the complex post mid-Eocene evolution of the region that involved collision and obduction events, subduction polarity reversals, arc splitting, subduction and back-arc basin formation.

Due to the sparsity of data in the southwest Pacific it may be fair to assume that if an arc existed then it has not yet been found. Remnants of the arc may be buried beneath younger structures such as the Tonga Arc (Falloon et al., 2014), extinct Vitiaz Arc (Eissen et al., 1998; Meffre et al., 2012; Lagabrielle et al., 2013) or Solomon Islands (Meffre et al., 2012). A ~102 Ma dolerite dredged from the Tonga forearc has back-arc basin geochemical characteristics and similarities with the Poya Terrane (Meffre et al., 2012; Falloon et al., 2014), making the Tonga Arc region an ideal target for future exploration.

Until more arc products are discovered or the remnant arc is located a mid or Late Cretaceous–Paleocene subduction scenario strongly relies on support from indirect evidence, namely formation of the South Loyalty Basin as a back-arc basin during this timeframe (e.g. Crawford et al., 2003; Schellart et al., 2006; Whattam et al., 2008; Cluzel et al., 2012a,b).

4.1.2. Nature and timing of the South Loyalty Basin opening

Crawford et al. (2003) emphasized the importance of back-arc basin opening for providing evidence for subduction to the east of the LHR. The South Loyalty Basin is often described as a Late Cretaceous–Paleocene back-arc basin (e.g. Cluzel et al., 2001, 2010a, 2012a,b; Crawford et al., 2003; Schellart et al., 2006; Whattam et al., 2008), although it has also been described as a marginal basin (Eissen et al., 1998). Regardless of what process drove opening or whether spreading began as early as the mid Cretaceous, if it opened during this period then a mid-ocean ridge must have existed to the east of New Caledonia, and at least one plate boundary separated the LHR from the Pacific.

A wide spread of ages, from only a few studies, exists for the South Loyalty Basin. These are largely derived from dating the Poya Terrane that formed the floor of the basin, and intrusions into the Peridotite Nappe. Radiometric age dating suggests that a basin was actively opening from the mid Cretaceous (Prinzhofer, 1981), possibly the result of thinned continental crust rifting away from the Gondwanaland margin (Collot et al., 1987). The ~85 Ma formation of magmatic zircons now contained within pelitic schist samples from the Pouebo Eclogite Melange is consistent with this timing if it is assumed that they are related to rift

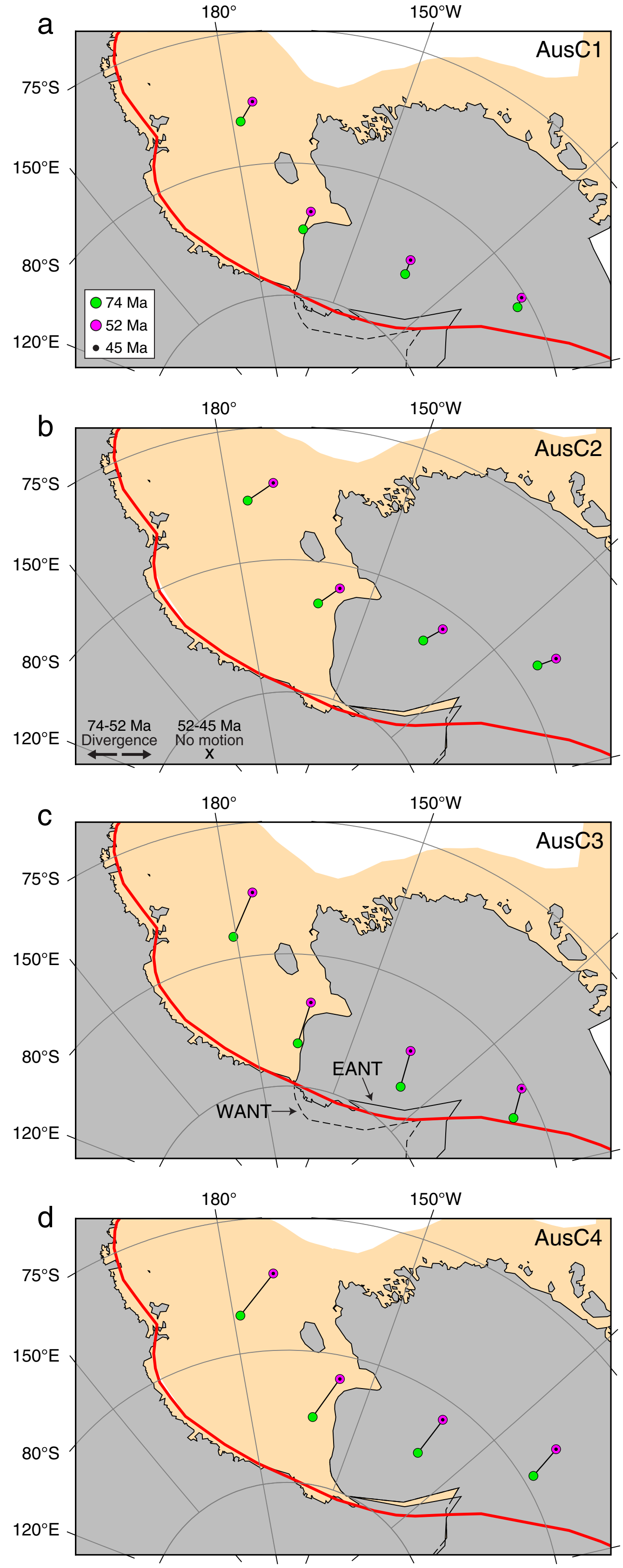

Fig. 13. Antarctica-centred reconstructions at 45 Ma in a fixed East Antarctica reference frame. Motion paths are shown for the West Antarctic plate based on different Australian plate circuits (AusC1–4, see Table 4). In these scenarios the Australian circuit is only adopted from 74–52 Ma, rather than 74–45 Ma. Present-day coastlines (gray) and continental crust (yellow) have been reconstructed. The approximate trace of the southern margin of the WARS is red. EANT, East Antarctica; WANT, West Antarctica. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

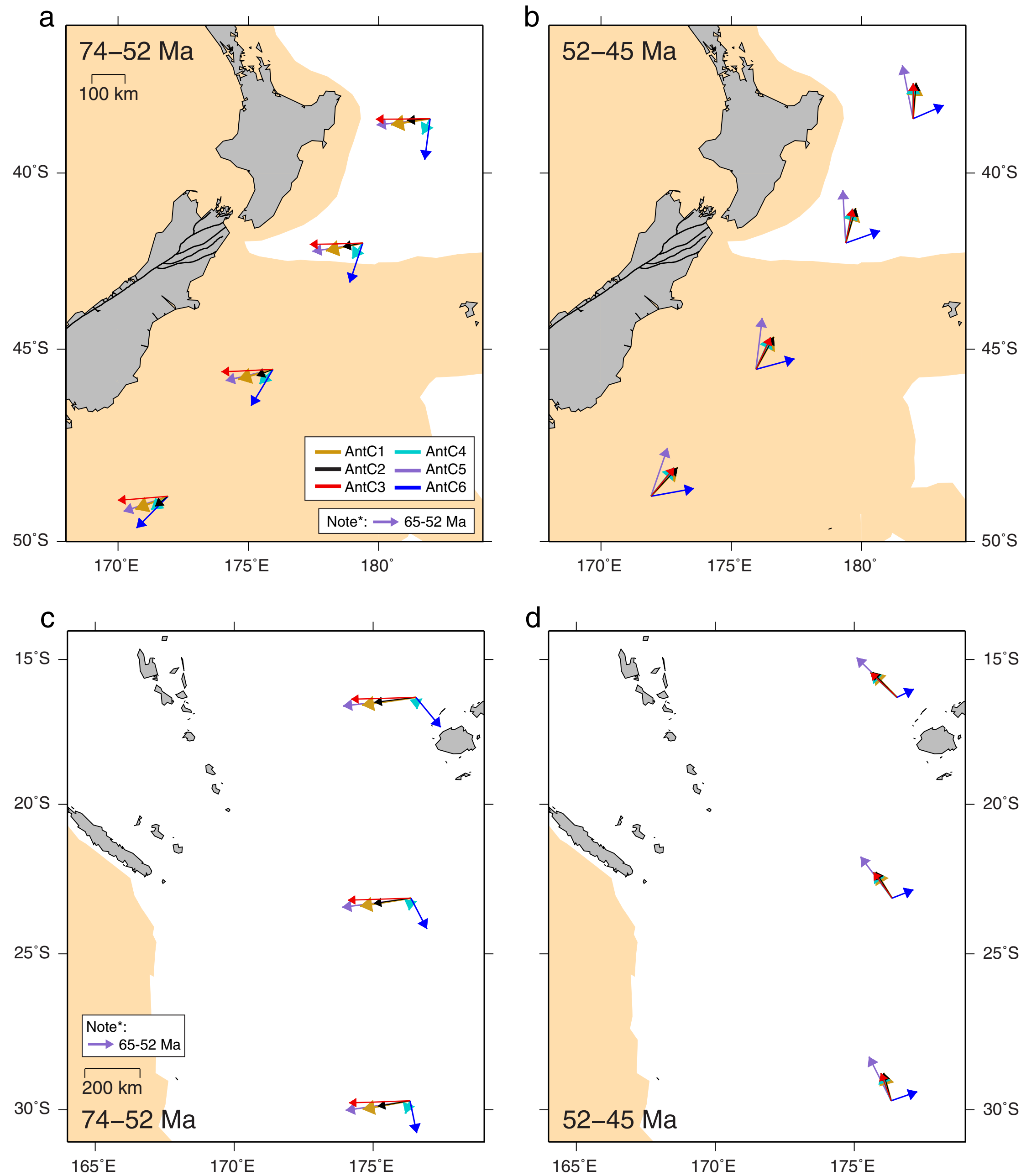

Fig. 14. Present-day maps of New Zealand (a–c) and the New Caledonia region (d–f). The motion of the Pacific plate with respect to the LHR is shown for each of the given time periods, and for each of the Antarctic plate circuits tested (AntC1–6, see Table 2). Arrowhead sizes and line thicknesses are for aesthetic purposes only. The Alpine Fault in New Zealand is also plotted.

magmatism (Spandler et al., 2005). Evidence for basin formation during the Late Cretaceous to Paleocene, leading up to the 55 Ma subduction initiation event, comes exclusively from fossil dating. Campanian to latest Paleocene/earliest Eocene fossils are interbedded in and overlie the Poya Terrane, suggesting that the basin may have formed at this time (Cluzel et al., 2001). If Campanian to Paleocene aged microfossils do in fact indicate timing of basin formation, then this suggests that the basin may have formed over a longer period from the mid Cretaceous to Paleocene, to be consistent with data from Prinzhofer (1981), Spandler et al. (2005), and Cluzel et al. (2001).

Supra-subduction zone geochemical affinities have been identified from samples of the Poya and Pouebo terranes, however the dominance of enriched mid-ocean ridge basalts suggest there was also a deep mantle influence, making the South Loyalty Basin an atypical back-arc basin (Cluzel et al., 2001). It is also not possible to rule-out a scenario in which the supra-subduction affinity is caused by mantle having been previously affected by subduction zone processes (Cluzel et al., 2001).

Subduction, slab roll-back and back-arc extension provide a driving mechanism for South Loyalty Basin formation. In the model of Schellart et al. (2006) subduction is linked with Pacific–LHR convergence (Model 1a). In the models put forth by Crawford et al. (2003) and Whattam et al. (2008) (Models 1b) there is no requirement for convergence between the Pacific and LHR. In this latter scenario there may have been little to no convergence, rather subduction that resulted in trench retreat and back-arc extension was driven by the negative buoyancy of

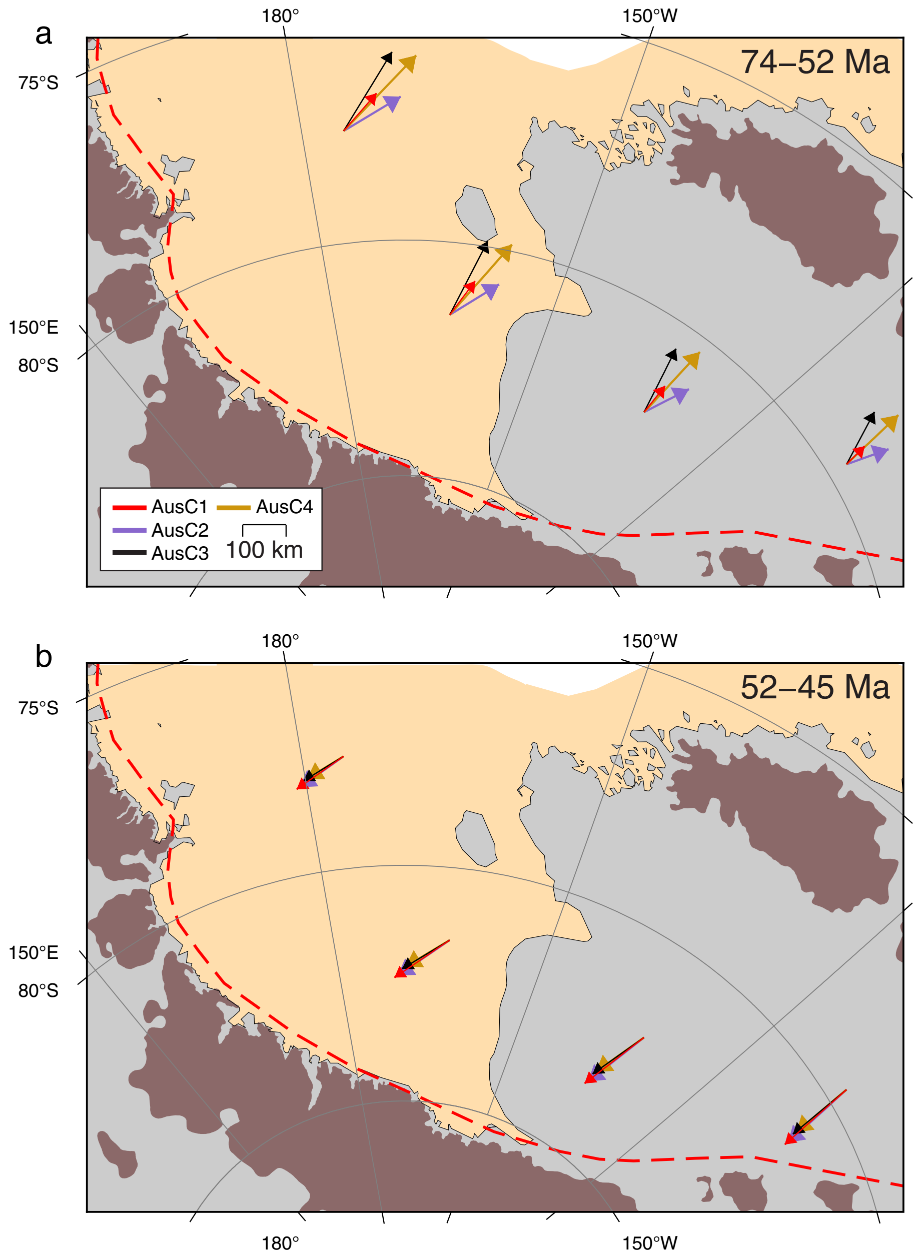

Fig. 15. Present-day map of Antarctica centred on the WARS, with >500 m topography shaded brown (ETOPO1, Amante and Eakins, 2009). The motion of the West Antarctic plate with respect to the East Antarctic plate is shown for each of the given time periods, and for each of the Australian plate circuits tested (AusC1–4, see Table 4). The approximate trace of the southern margin of the WARS is red. Arrowhead sizes and line thicknesses are for aesthetic purposes only. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

the slab. This has been proposed for the Tyrrhenian Basin–Apennines Arc system since 20–15 Ma (e.g. Malinverno and Ryan, 1986; Jolivet et al., 2008) and the Pannonian Basin–Carpathian Arc during the middle Miocene (Royden and Burchfiel, 1989).

Extension and fragmentation along Eastern Gondwanaland may have been driven by collapse of the East Gondwanaland Cordillera from ~90 Ma (Rey and Müller, 2010). In the orogenic collapse model of Rey and Müller (2010) changes in relative motion at the active continental margin, between Eastern Gondwanaland and the Pacific, reduced the buoyancy in the mantle wedge and this led to collapse and boudinage in the overriding plate. In this model a subduction zone is required to act as a free boundary to accommodate the continental thinning (Rey et al., 2001). The extension in the overriding plate forced trench retreat, rather than a significant amount of subduction due to convergence.

Another mechanism for driving back-arc basin formation was recently modeled by Moresi et al. (2014) and applied to the Ordovician–Silurian evolution of the Tasmanides, eastern Australia. Their model may also be applicable to opening of the South Loyalty Basin. The 3-D Cartesian dynamic modeling of Moresi et al. (2014) predicts that when buoyant crust enters and chokes a subduction zone then subduction zone retreat and back-arc extension can occur at the margin along strike from the collision zone. The Hikurangi Plateau had arrived at the New Zealand margin of the Eastern Gondwanaland subduction zone by ~100 Ma (Davy, 1992; Davy and Wood, 1994; Lonsdale, 1997; Hoernle et al., 2010). For instance, 99–87 Ma HIMU-type seamounts on the Hikurangi Plateau have a similar age and composition to igneous complexes in New Zealand's South Island, yet a different composition to its basement lavas (Hoernle et al., 2010). Based on the timing of cessation of spreading at the Osbourn Trough, final collision of the Hikurangi Plateau has

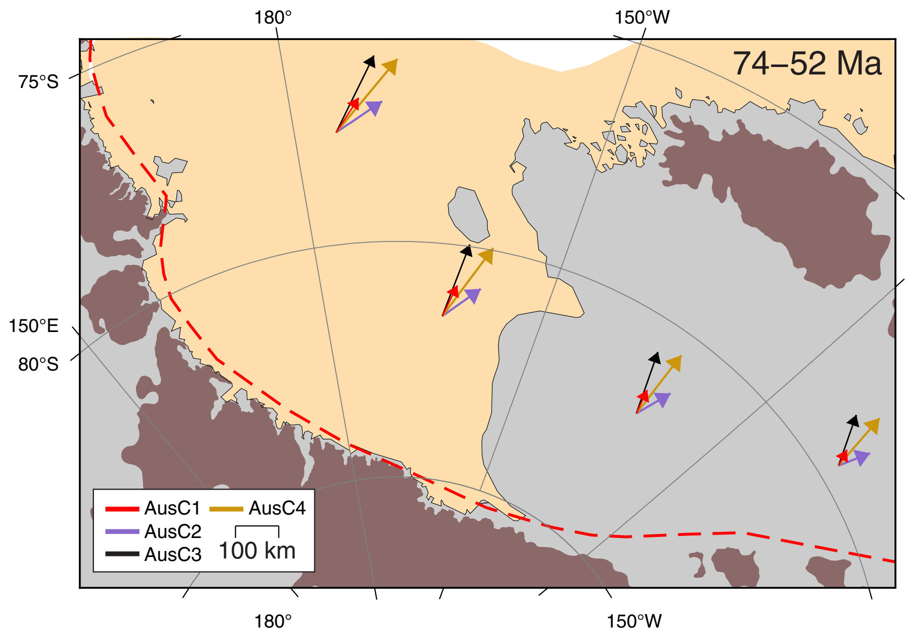

Fig. 16. Present-day map of Antarctica centred on the WARS, with >500 m topography shaded brown (ETOPO1, Amante and Eakins, 2009). The motion of the West Antarctic plate with respect to the East Antarctic plate is shown for each of the given time periods, and for each of the Australian plate circuits tested (AusC1–4, see Table 4). In these scenarios the Australian circuit is only adopted from 74–52 Ma, rather than 74–45 Ma. The approximate trace of the southern margin of the WARS is red. Arrowhead sizes and line thicknesses are for aesthetic purposes only. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

also been dated at ~93–87 Ma (Downey et al., 2007) or ~86 Ma (Worthington et al., 2006). These ages are coincident with or slightly predate suspected initial opening of the South Loyalty Basin. Therefore the mechanism discussed by Moresi et al. (2014) may be a candidate for driving South Loyalty Basin extension.

Reconstruction model 1c and 2 (Fig. 2) are inconsistent with opening of the South Loyalty Basin during the period from the mid or Late Cretaceous to 55 Ma. Model 1c incorporates basin closure and the relative motion of the Loyalty Ridge towards New Caledonia due to east dipping subduction, while in Model 2 there is no plate boundary between the Pacific and LHR.

Table 5
Convergence/divergence between the Pacific and LHR plates using an Antarctic plate circuit.

Circuit	Convergence or divergence[ab] (km)		
	74–52[c] Ma	52–45 Ma	Total
New Zealand sector[d]			
AntC1[e]	50	30	80
AntC2	20	30	50
AntC3[e]	95	40	135
AntC4	10 (d)	40	30
AntC5[e]	65	90	155
AntC6	35 (d)	55 (d)	90 (d)
New Caledonia sector[f]			
AntC1[e]	185	25	210
AntC2	135	25	160
AntC3[e]	225	45	270
AntC4	15	45	60
AntC5[e]	240	60	300
AntC6	30 (d)	85 (d)	115 (d)

[a] Divergence is indicated by a "(d)", all other values are for convergence.
[b] These amounts are qualitative and give an indication of the approximate amounts of predicted convergence/divergence. They do not take into account error associated with the poles of rotation at each plate pair within the Antarctic circuit.
[c] AntC5 values are for 65–52 Ma, not 74–52 Ma.
[d] See Fig. 10 for the motion path that was used to compute convergence/divergence.
[e] No East–West Antarctica motion before opening of the Adare Trough at ~43–40 Ma.
[f] See Fig. 11 for the motion path that was used to compute convergence/divergence.

4.1.3. Other possible basin opening events between LHR and Pacific

Several southwest Pacific plate reconstruction models also incorporate back-arc basin opening to the north of New Zealand during the period from ~83–55 Ma that was continuous with spreading to the east of New Caledonia in the proposed South Loyalty Basin (Crawford et al., 2003; Whattam et al., 2008; Cluzel et al., 2010b). Remnants of this basin are purportedly preserved in the correlative Northland and East Coast allochthons in northern New Zealand, where obducted basalts exhibit supra-subduction zone geochemical characteristics (e.g. Nicholson et al., 2000a, 2007; Whattam et al., 2004; Cluzel et al., 2010b). As is the case for the South Loyalty Basin, a wide spread of ages has been proposed for the allochthon components, based on different dating techniques. Fossils from the Cretaceous to Oligocene are reported to be intercalated in some of the basalts in various parts of the allochthons, however their preservation is poor and abundance is low in some localities (Spörli and Aita, 1994) and furthermore it is not straightforward to determine whether the fossils are truly interbedded (Strong, 1980). Radiometric dating has been conducted by only a handful for studies and reveals a clustering of Oligocene ages (Whattam et al., 2004, 2005), and late Eocene to Oligocene ages (Brothers and Delaloye, 1982), as well as sporadic Cretaceous to Paleocene ages, many of which are associated with large uncertainties of up to ± 29 Myr (Brothers and Delaloye, 1982). This distribution of ages fits with the suggestion of Strong (1980) and Katz (1976) for the Matakaoa Volcanics that there may have been multiple separate volcanic events based on such a wide spread of fossil derived ages, that are unlikely to represent continuous spreading over such a long period of time (Katz, 1976). There is also no evidence for an arc prior to the well-accepted Miocene volcanic arc (Herzer et al., 2009).

4.1.4. History of deformation within New Zealand and implications for plate boundary activity

At 45 ± 5 Ma a major plate boundary is believed to have propagated into New Zealand and accommodated rifting between the Challenger and Campbell plateaus (Sutherland, 1995). The location of rifting may have been controlled by pre-existing zones of weakness in the lithosphere, such as fracture zones and a pre-existing discontinuity in the lithosphere between the Challenger and Campbell plateaus (Sutherland et al.,

2000) and/or a fossil spreading centre at the continental western margin of the Campbell Plateau (Barker et al., 2008). Prior to this time the existence of a major plate boundary is not supported by existing studies (Sutherland, 1995; Steinberger et al., 2004), although it is likely that non-rigid intraplate deformation has occurred and resulted in oroclinal bending at ~85 Ma coinciding with Tasman Sea opening and rifting in the Bounty Trough (Mortimer, 2014).

Deformation likely occurred in New Zealand during early Tasman Sea spreading until ~74 Ma (Sutherland, 2008), however after this time New Zealand is commonly described as tectonically quiescent until the 45 ± 5 Ma mid Eocene rifting event. From our review of tectonic activity in New Zealand during Tasman Sea opening there is only a very small amount of geologic evidence for deformation, such as possible shear zone motion (Klepeis et al., 1999; King et al., 2008), Alpine Schist metamorphism (Mortimer and Cooper, 2004; Cooper and Ireland, 2013), mylonitisation (White and Green, 1986; Rattenbury, 1987) and extension in the West Coast region (Bishop, 1992; Thrasher, 1992; Laird, 1994; Schulte, 2011). This tectonic activity mainly occurred before ~70 Ma (Fig. 3).

For the period ~83–55 Ma geologic observations are consistent with either no plate boundary running through New Zealand or alternatively a plate boundary with only very limited motion amounting to minor deformation.

4.2. Tectonic evolution of the southwest Pacific and plate circuit implications

Based on our review of southwest Pacific geology combined with kinematic testing of alternative plate circuits we are able to describe a history of plate boundary evolution in the region that is consistent with the available geologic constraints (Fig. 17). Where relative plate motions are well-constrained the reconstructions should be relatively insensitive to our choice of plate circuit.

>74 Ma

Chron 33y (~74 Ma) is the oldest magnetic anomaly that has been widely identified in the southwest Pacific (Sutherland, 2008). Increased uncertainties in spreading histories at older times make it difficult to critically assess plate circuit implications and properly combine geologic observations with kinematic predictions. Therefore, prior to 74 Ma we focus on geologic observations to constrain plate boundary activity and plate motions.

South Loyalty Basin seafloor spreading initiates by ~85 Ma (Cluzel et al., 2001) and detaches a fragment of the LHR (Eissen et al., 1998). The basin is a back-arc basin with opening driven by east-directed roll-back of a west-dipping subduction zone. South Loyalty Basin seafloor spreading is contemporaneous with rifting and seafloor spreading between Australia and the LHR (Gaina et al., 1998).

74–52 Ma

By 74 Ma opening of the South Loyalty Basin had been under way for at least ~10 Myr based on radiometric and fossil dating of the Poya Terrane. Spreading in the basin continued until 55 Ma when a subduction initiation event occurred to the east of New Caledonia (e.g. Crawford et al., 2003; Cluzel et al., 2006, 2012a,b; Whattam et al., 2008; Ulrich et al., 2010; Lagabrielle et al., 2013). A phase of basin opening may also have been underway during this period to the north of New Zealand based on radiometric and fossil dating of obducted ocean floor in northern North Island.

A plate boundary through New Zealand at the long-lived Challenger Plateau–Campbell Plateau lithosphere discontinuity (Sutherland et al., 2000) connected the west-dipping subduction zone to spreading in the Tasman Sea that was beginning to dwindle at about the same time as the ~55 Ma subduction initiation event (Gaina et al., 1998). The rules of plate tectonics require plate boundaries to be continuous features, and this scenario allows for continuity of plate boundaries to be preserved. Motion within New Zealand was minor, possibly due to a proximal pole of rotation, until rifting commenced ~45 ± 5 Ma (Sutherland, 1995).

Kinematic results from analysing alternative plate circuits have interesting implications for understanding this period of basin opening and reconciling geologic observations from New Zealand with those from New Caledonia. A scenario of little or no net motion between the Pacific and LHR can be supported by use of an Antarctic plate circuit, despite leaving motion unconstrained at this boundary (Fig. 9a–b). This is consistent with a tectonically quiet New Zealand, yet also permits a situation in which subduction to the east of New Caledonia drives back-arc basin opening. Subduction can occur in the absence of convergence (Doglioni et al., 2006), rather the slab sinks into the mantle driven by its own negative buoyancy. In this situation the amount of back-arc basin opening would be about equal to the amount of trench retreat. This scenario is also consistent with subduction driven by orogenic collapse as described by Rey and Müller (2010), and the subduction zone retreat may also have been influenced by the Hikurangi Plateau choking the New Zealand subduction zone in the mid or Late Cretaceous (c.f. Moresi et al., 2014).

Our plate motion analysis for the Antarctic circuit shows that convergence between the LHR and the Pacific is predicted by some but not all models. A large amount of convergence is predicted by the plate circuit used by Schellart et al. (2006) (Fig. 14, AntC5), whereas plate circuits involving revised kinematics for different plate-pairs yield significantly different results. Our preferred scenario (AntC4) incorporates minimal motion within New Zealand, East–West Antarctica motions consistent with the recently revised kinematics for the WARS, and a recently revised spreading history for Australia–East Antarctica. It predicts less that 50 km of strike-slip motion over the 22 Myr period (<2.5 mm/yr) between the Pacific and LHR in the region to the north of New Zealand (Fig. 14). This includes an element of convergence between the LHR and the Pacific to the east of New Caledonia (up to 15 km) (Table 5). This small amount of motion over such a long period of time excludes a model with large amounts of convergence and associated roll-back (Model 1a). However it is consistent with a model of minimal convergence and subduction (Model 1b).

Although an Australian plate circuit leaves motion between East and West Antarctica unconstrained during this period, reasonable motion is predicted within the West Antarctica Rift System when we (1) incorporate an up-to-date spreading history for Australia–East Antarctica, that matches geophysical observations along the entire margin from Broken Ridge/Kerguelen Plateau to Tasmania and Cape Adare (Whittaker et al., 2013), and (2) use the East–West Antarctica rotations of Granot et al. (2013) for times younger than 40 Ma with an extrapolation to 45 Ma in order to implement the cross-over of the Pacific plate from an Antarctic plate circuit to an Australian plate circuit at 45 Ma. This model, AusC2, is our preferred Australian plate circuit for this period. In this scenario ~150–100 km of motion is predicted over the 22 Myr period that is orthogonal in the Ross Sea region and more dextral further south in the region of the present-day West Antarctic Ice Sheet (Fig. 15). The kinematics of this extension are similar to that predicted by the reconstruction model of Wilson and Luyendyk (2009), though the magnitude of extension predicted by our model is smaller for the same period. Our preferred Australian model also shows a broad continuity of kinematics from those determined by Granot et al. (2013) using geophysical data for the younger period from 40 to 26.5 Ma.

An Australian plate circuit can produce reasonable East–West Antarctica kinematics despite treating the Pacific plate and LHR as a single plate. Adopting the preferred Australian plate circuit (AusC2) does not rule out a tectonic history for the southwest Pacific incorporating plate boundary activity between the Pacific plate and LHR, rather it is consistent with a scenario of little to neglible net relative motion. This plate circuit can support a plate boundary scenario involving subduction and back-arc basin formation in which subduction is not driven by convergence, and back-arc spreading is equal to the amount of trench retreat.

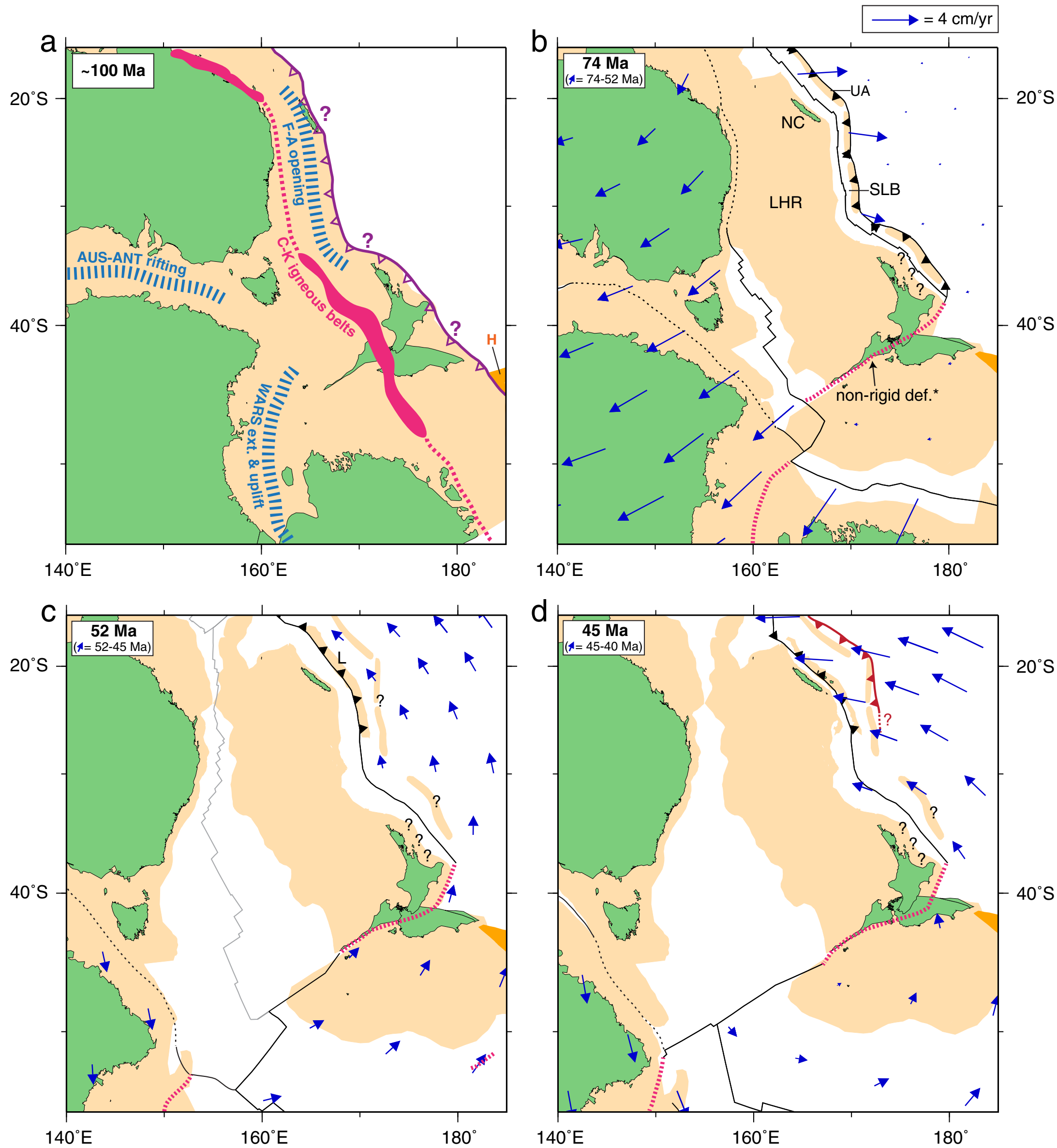

Fig. 17. (a) Mid Cretaceous (~100 Ma) Eastern Gondwanaland margin after Mortimer (2008) and Collot et al. (2009). Subduction may have ended at this time along all or part of the margin associated with a major plate reorganization (for more details see Matthews et al., 2012). There is opening of the Fairway-Aotea Basin at this time (Collot et al., 2009), as well as extension between Australia and Antarctica, and in the West Antarctic Rift System. Also, by this time the Hikurangi Plateau (H) had arrived at the Zealandia margin (e.g. Hoernle et al., 2010). (b–d) Reconstructions depicting our preferred scenario for southwest Pacific evolution, using our preferred Antarctic plate circuit (AntC4), and based on Crawford et al. (2003) and Whattam et al. (2008) (Model 1b, Fig. 2). Plates are reconstructed using a fixed Lord Howe Rise (LHR) reference frame. West-dipping subduction occurs to the east of the LHR from at least ~85 Ma until 55 Ma, and is associated with opening of the South Loyalty Basin (SLB) due to east-directed slab roll-back from 85 to 55 Ma (although opening may have initiated in the mid Cretaceous). There is minimal relative motion between the Pacific plate and LHR during basin opening. Northeast-dipping subduction initiates at the extinct SLB spreading centre at ~55 Ma, possibly associated with construction of the Loyalty Arc (L), and consumes the basin. It should be noted that there is large uncertainty in the reconstruction to the north of New Zealand and more work is needed to better refine this region. Vectors indicate the average velocities for the periods 74–52 Ma (shown in b), 52–45 Ma (shown in c) and 45–40 Ma (shown in d). Continental crust is light tan and reconstructed as rigid blocks. For reference the present-day coastlines are filled green. The Tasman Sea spreading ridge at 52 Ma is gray as this is when spreading ceased. Dashed black lines denote continental rifting (b–d). The locations of the plate boundaries through New Zealand and between East and West Antarctica are particularly uncertain (pink dashed lines, b–d). *Regional-scale non-rigid intraplate deformation has likely occurred within New Zealand according to Mortimer (2014), suggesting the existence of a diffuse plate boundary. This may also have been the case between East and West Antarctica. C-K, Carboniferous–Cretaceous; F-A, Fairway-Aotea Basin; NC, New Caledonia; UA, currently unidentified arc. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

52–45 Ma

There is a consensus that between 45 Ma and present-day one or multiple plate boundaries separated the LHR and Pacific plate. Pacific–LHR relative motion cannot be directly constrained during this period due to intra-plate deformation within New Zealand and the destruction of ocean floor due to episodes of subduction. Instead, the Pacific plate must be reconstructed using a plate motion chain through Antarctica during this interval, and an Antarctic plate circuit (Fig. 9a). From our review of southwest Pacific geologic data there must have been a plate boundary separating the Pacific from the LHR since at least 55 Ma, 10 Myr earlier than has previously been proposed in some studies (Steinberger et al., 2004; Doubrovine et al., 2012). At ~55 Ma subduction initiated to the east of New Caledonia (Crawford et al., 2003; Cluzel et al., 2012a,b; Meffre et al., 2012) and we therefore propose that for the extended period 55-0 Ma an Antarctic plate circuit is more consistent with geologic observations and the Pacific plate should be reconstructed using a plate motion chain through Antarctica. Alternatively, adopting various Australian plate circuits during the period from 52–45 Ma (following Tasman

Sea opening), and thus treating the Pacific and LHR as a single plate, results in at least 100–200 km of compression in the WARS, which cannot be supported by existing geologic observations (Fig. 15). We emphasize that there are fewer constraints on WARS deformation before ~43 Ma, compared to the period of Adare Trough opening from ~43–26 Ma, and as more data are acquired the plate circuit calculations will become more tightly constrained.

4.2.1. Model limitations and future improvements

Scarce and ambiguous geologic data have made it difficult to favour a single model for southwest Pacific evolution, particularly from ~85 to 55 Ma. However, a scenario of west-dipping subduction and back-arc basin formation is able to account for the geologic observations we reviewed and is therefore our preferred model (Model 1b). We emphasize that opening of the South Loyalty Basin throughout the entire period from the Late Cretaceous to Paleocene largely underpins this model, rather than a preserved volcanic arc. In turn, the timing of its formation hinges on the interpretations of Cluzel et al. (2001) that Campanian–Paleocene microfossils are interbedded with Poya Terrane basalts and thus record seafloor formation. A model with no basin opening during this time would have to find an alternative explanation for these interbedded Campanian–Paleocene microfossils. While some doubt surrounds several aspects of our preferred tectonic reconstruction, we identify testable predictions to be addressed in future investigations.

Analysing the mantle structure beneath the southwest Pacific inferred from seismic tomography has allowed for a better understand of the subduction history of the region over the past 45 Myr (Schellart et al., 2009; Schellart and Spakman, 2012). Tomographic models may shed light on the older evolution of the southwest Pacific, and this will require deeper imaging at more southerly latitudes compared to the studies by Schellart et al. (2009) and Schellart and Spakman (2012).

An important and testable element of our preferred model (Model 1b) is the prediction that remnants of the arc associated with west-dipping subduction that drove opening of the South Loyalty Basin should be buried beneath the arc associated with early Tonga–Kermadec and Vitiaz subduction (Eissen et al., 1998; Meffre et al., 2012; Falloon et al., 2014). This reflects an important aspect of the model, that subduction zones exploit pre-existing zones of weakness (Hall et al., 2003; Gurnis et al., 2004). The recovery of an ~102 Ma dolerite from the Tonga forearc that exhibits back-arc basin characteristics and similarities with the Poya Terrane (Meffre et al., 2012; Falloon et al., 2014) is intriguing and strongly encourages renewed exploration in this region. The model also predicts that the Loyalty Ridge is an extinct Eocene arc (Eissen et al., 1998; Cluzel et al., 2001; Crawford et al., 2003; Whattam et al., 2008) that was active from ~55 Ma, again testing this prediction will require deep drilling or carefully targeted dredging of the ridge.

4.3. Model implications for global plate motion studies using plate circuits

Global plate motion studies rely on plate motion chains through the southwest Pacific to link observations from the Indo-Atlantic and Pacific domains. Several studies have directly compared the Antarctic and Australian plate circuits (Fig. 9a–b). These studies investigated the absolute motion of the Pacific plate using paleomagnetic data (Doubrovine and Tarduno, 2008; Somoza and Ghidella, 2012), or through numerical modeling coupled with observations from seamount trails (Steinberger et al., 2004; Doubrovine et al., 2012). Our review of southwest Pacific geology combined with our plate circuit analyses, have implications for interpreting the results of these geodynamic studies, and future studies of absolute and relative Pacific plate motion. We are also able to highlight where future numerical modeling studies can be used to further constrain plate boundary evolution in the region, particularly immediately after the mid Cretaceous plate reorganization event (Veevers, 2000; Matthews et al., 2012).

4.3.1. Previous investigations of the Antarctic and Australian plate circuits

Paleomagnetic data analyses. Doubrovine and Tarduno (2008) used paleomagnetic data to test the alternative plate circuits. They compared Pacific paleomagnetic data, obtained from four Emperor seamounts (Koko — 49.2 Ma, Nintoku — 55.5 Ma, Suiko — 60.9 Ma and Detroit — 75.8 and 80 Ma), with North American and East Antarctic paleomagnetic poles that had been transferred into a Pacific reference frame. The non-Pacific paleomagnetic poles, from Torsvik et al. (2001) and Besse and Courtillot (2002), were transferred to the Pacific plate using the two alternative plate circuits. For each investigated time the choice of plate circuit did not strongly affect the position of the rotated pole, rather the differences in the positions of the transferred poles (~1–2.5°) were smaller than the uncertainty associated with the paleomagnetic data. Doubrovine and Tarduno (2008) concluded that both circuits could be used to investigate Pacific plate motion, however their results did not give preference to one over the other.

Somoza and Ghidella (2012) performed a similar test of the Antarctic and Australian plate circuits, by transferring their Late Cretaceous paleomagnetic pole for South America to Pacific coordinates using each circuit. They found that the alternative circuits gave near identical pole positions that both matched Pacific paleomagnetic data.

Numerical simulations of mantle flow and hotspot trails. Steinberger et al. (2004) and Doubrovine et al. (2012) compared the Australian and Antarctic plate circuits using numerical models of whole mantle convection. These models incorporate plume conduits stemming from the D″ layer that freely advect with the mantle flow field, so that their surface motions can be compared with hotspot chains. Using plate velocities from a plate reconstruction model as surface boundary conditions, present-day mantle density anomalies inferred from the SMEAN seismic tomography model (Becker and Boschi, 2002) are backwards advected between 0 and 70 Ma using the technique of Steinberger and O'Connell (1998); a constant flow field is assumed for times older than 70 Ma to avoid major numerical artifacts (Conrad and Gurnis, 2003).

Steinberger et al. (2004) were interested in predicting four hotspot chains in the Pacific and Indo-Atlantic realms, and ultimately reproducing the bend in the Hawaiian–Emperor chain at ~50 Ma, that usually cannot be recreated from plate reconstruction models alone. They found that by adopting a plate motion model that reconstructs the Pacific plate via a circuit through Australia for times older than 43 Ma they could produce a closer match to the Hawaiian–Emperor chain back to 65 Ma, including the bend period. On the other hand, incorporating the Antarctic circuit for Pacific plate motion resulted in a comparatively broader bend and a poorer fit before 50 Ma.

Doubrovine et al. (2012) used this numerical modeling technique to produce an updated global moving hotspot reference frame. From looking at goodness of fit statistics and RMS misfits they suggested that the absolute reference frame produced using the Australian plate circuit for constraining Pacific plate motion results in hotspot motions that more closely match five Pacific and Indo-Atlantic hotspot chains (Hawaiian–Emperor, Louisville, Tristan, Reunion and New England), compared to a reference frame that is built from an Antarctic plate circuit for constraining Pacific plate motion. Their resultant synthetic hotspot trails were very similar for times <70 Ma. Both plate circuits produced references frames with the poorest fits at 70 and 80 Ma, however only the misfits associated with the Antarctic circuit were deemed formally unacceptable based on goodness of fit statistical criteria.

4.3.2. New implications for global studies

As pointed out by Wessel et al. (2006), sensitivity of results in the global analyses discussed above to the plate circuit used suggests that some aspects of the plate circuits are poorly constrained — for example motion within the WARS, or the timing and kinematics of plate boundary processes east of the LHR. The choice of spreading histories within a plate circuit is crucial and has important implications for distinguishing between plate circuits (Tarduno et al., 2009). Tarduno et al. (2009)

compared the predicted trace of the Hawaiian–Emperor chain resulting from using several Pacific plate motion chains, including two that traverse Australia that adopt different Australia–East Antarctica spreading histories, and one through Antarctica. While the Australian plate motion chains predicted closer matches to the Hawaiian–Emperor chain, the Australian plate motion chain with the revised Australia–East Antarctica spreading history actually converged with the Antarctic plate motion chain results. Their results demonstrate the importance of incorporating the most geologically robust spreading histories before drawing conclusions on the soundness of a plate motion chain or circuit.

Steinberger et al. (2004) supported a model for southwest Pacific evolution in which the LHR and Pacific plate were attached from 83–45 Ma. The work of Doubrovine et al. (2012) later supported this view. These studies highlighted that a combination of plume motion and the careful selection of a plate circuit could result in a model that closely predicts hotspot trails. These were both global studies, and notably their results disagreed with many regional plate reconstruction models that had previously proposed subduction to the east of the LHR (Cluzel et al., 2001; Crawford et al., 2003; Sdrolias et al., 2003, 2004; Schellart et al., 2006). In stark contrast to this, the model of Schellart et al. (2006) proposed that there was convergence between the Pacific plate and LHR, with ~1500 km of subduction of the Pacific plate during the period from 82–45 Ma. These are end member studies and are mutually exclusive (Model 1a versus Model 2). Both studies addressed the use of plate circuits in different ways and found their results were consistent with only one plate circuit; Schellart et al. (2006) with the use of an Antarctic circuit, and Steinberger et al. (2004) and Doubrovine et al. (2012) with the use of an Australian circuit. Our results have shown that, when using well-constrained relative motion histories for Australia–Antarctica and East–West Antarctica, and motion in New Zealand is minimized, during the period from 74–55 Ma both the Australian and Antarctic plate circuits produce similar results when adopted in southwest Pacific reconstruction models, and motions at each plate pair do not violate geologic observations. Although our preferred tectonic history for the southwest Pacific includes opening of the South Loyalty Basin during this period driven by subduction and slab roll-back, as there was likely little net relative motion between the Pacific plate and LHR using an Australian plate circuit will produce similar kinematic results to an Antarctic circuit.

Our regional assessment of how the plate circuit in the southwest Pacific has evolved due to changes in plate boundary activity has important implications for interpreting the results of the numerical models of mantle convection that investigated plume advection in the mantle and absolute Pacific plate motion (Steinberger et al., 2004; Doubrovine et al., 2012). We suggest that in order to satisfy a wide variety of geologic observations from the southwest Pacific and Antarctica an Antarctic plate circuit should be adopted from at least 55–0 Ma rather than 45–0 Ma (Steinberger et al., 2004; Doubrovine et al., 2012). Significantly, this 10 Myr time period includes formation of the bends in the Hawaiian–Emperor and Louisville seamount chains at ~50 Ma, which were used as important temporal and spatial markers in the above studies. Therefore our results have implications for assessing the relative importance of different processes that may have contributed to formation of the Hawaiian–Emperor chain bend, including shallow sub-lithospheric processes such as plume-ridge interaction, as well as deep mantle flow (Tarduno et al., 2009).

Previous studies (e.g. Steinberger et al., 2004; Schellart et al., 2006) have also relied on poorly-constrained East–West Antarctica motion for the 26–43 Ma period from Cande et al. (2000). Better-constrained motion based on new geophysical data (Granot et al., 2013) implies significantly different kinematics. The overall magnitude of Eocene–Oligocene motion remains relatively small, but because the well-constrained pole of rotation is near to the WARS, the implications for the absolute position of the Pacific plate reconstructed relative to African reference frames is magnified, as envisaged by Sutherland (2008).

Our reconstructions, and their implications for the subduction history in the southwest Pacific, also have important consequences for estimates of slab pull forces acting on the Pacific plate through time (Faccenna et al., 2012; Butterworth et al., 2014), and subduction budgets around the circum-Pacific (Doglioni et al., 2007; Sutherland, 2008). Changes in slab-pull in the western Pacific have been linked with plate boundary reorganizations (e.g. Faccenna et al., 2012). Yet determining the cause and effect relationship between widespread plate motion changes and plate boundary readjustments is still a matter of ongoing debate.

4.4. Possibility of prolonged activity at the relict Eastern Gondwanaland plate boundary

Observations to support convergence and arc magmatism prior to 55 Ma to the east of the LHR are limited, however this does not rule-out certain reconstruction scenarios in which there is a plate boundary between the Pacific plate and the LHR. Firstly, a small amount of oblique convergence at a plate boundary is possible, as is predicted by our preferred Antarctic plate circuit (AntC4, see Table 2). Furthermore, subduction and back-arc basin formation can occur in the absence of convergence, driven by sinking of the slab under its own negative buoyancy. Orogenic collapse and forced trench retreat, is a possible mechanism for driving subduction and continental fragmentation with minimal convergence (Dewey, 1988; Rey and Müller, 2010).

Long-lived west-dipping subduction beneath Eastern Gondwanaland is commonly reported to have ended at ~105–100 Ma (Laird and Bradshaw, 2004) and this resulted in dramatically altered motion at the plate boundary, possibly giving way to strike slip motion (Veevers, 2000) and orogenic collapse of the East Gondwanaland Cordillera (Rey and Müller, 2010). Alternatively subduction ended closer to 85 Ma (Vry et al., 2004). Unlike subduction cessation due to major continent–continent collision, which is an obvious mechanism for terminating subduction, subduction cessation to the east of Eastern Gondwanaland likely resulted from, or was at least partially attributed to, thick/buoyant oceanic lithosphere choking the New Zealand portion of the margin (e.g. Bradshaw, 1989; Davy et al., 2008). It is plausible that motion continued at the relict subduction zone plate boundary further north, in the region of New Caledonia.

The mid Cretaceous was a time of major global-scale plate motion changes and therefore the plate boundary to the east of the LHR and New Caledonia may not have “healed” to form a rigid LHR-Pacific plate. This plate boundary, a lithospheric zone of weakness, would make a suitable location for subduction reinitiation at ~90 Ma (Rey and Müller, 2010), or alternatively initiation of Eocene west-dipping Tonga–Kermadec subduction (Gurnis et al., 2004; Sutherland et al., 2010). In the region of New Zealand where the Hikurangi Plateau choked the Eastern Gondwanaland subduction zone the plate boundary may have jumped west in the mid Cretaceous to the lithospheric discontinuity identified by Sutherland et al. (2000), and may account for very minor deformation observed in New Zealand during Tasman Sea opening.

Although this scenario is highly speculative, a better understanding of major plate boundary reorganizations associated with subduction cessation (Matthews et al., 2012) would benefit from numerical modeling investigations and will help with deciphering plate boundary evolution in the southwest Pacific following the reorganization at ~105–100 Ma. Conversely, investigating the possibility of a minor strike-slip boundary in New Zealand solely from empirical field or geophysical data would likely be very difficult. For instance deformation at the Alpine fault since 45 Ma may have destroyed any evidence for earlier plate boundary activity.

4.5. Significance of subduction initiation at ~55 Ma

Subduction initiation at ~55 Ma to the east of New Caledonia appears to be part of a wider plate reorganization that occurred at the beginning of the Eocene in the Pacific region. Yet a driving mechanism for subduction initiation at 55 Ma has not previously been forthcoming (Ulrich et al., 2010). Subduction initiation is thought to require convergence at a pre-existing zone of weakness in the lithosphere associated with a change in the balance of driving forces acting at the boundary (Hall et al., 2003; Gurnis et al., 2004), unless initiation was spontaneous due to collapse of gravitationally unstable lithosphere into the mantle (Stern, 2004). A better understanding of the location, nature and timing of these plate motion changes will therefore lead to a better understanding of what drove southwest Pacific subduction initiation at 55 Ma, and how this event was linked with other events in the Pacific region.

Elsewhere in the southwest Pacific plate circuit, in the Tasman Sea, a decrease in spreading rate and a counterclockwise change in the direction of spreading occurred at chron 25y time (55.9 Ma), shortly prior to the end of spreading at about 52 Ma (Gaina et al., 1998). At ~55 Ma an exhumation event began at the Transantarctic Mountains front (Gleadow and Fitzgerald, 1987; Fitzgerald, 1992) and may have been related to early extension in the Adare Trough (Fitzgerald, 2002). It is recognized first in northern and southern Victoria Land at ~55 Ma and apparently propagated southwards reaching the central Transantarctic Mountains at ~50 Ma (Fitzgerald, 2002). A reorganization at the Australian–East Antarctic spreading ridge occurred between chrons 24o and 21y (~53–46 Ma), with a change from NW–SE to NNE–SSW spreading (Whittaker et al., 2013).

Further afield in the Pacific ocean basin seafloor fabric records spreading reorganizations at the Pacific–Kula and Pacific–Farallon spreading ridges. Between the end of chron 25 and the beginning of chron 24.1 (55.9–52.7) there was a change in motion at the Pacific–Kula spreading ridge inferred from fracture zone bends and changes in abyssal hill orientations (Lonsdale, 1988), and fracture zone bends observed in the central Pacific formed from changes in spreading at the Pacific–Farallon spreading ridge at chron 24 time (~53.3 Ma) (Kruse et al., 1996).

Previously Izu–Bonin–Mariana subduction initiation in the southwest Pacific was dated at ~45 Ma, along with Tonga–Kermadec subduction initiation in the southwest Pacific (e.g. Stern and Bloomer, 1992; Bloomer et al., 1995), and this similarity in timing has long been of interest for studies of subduction initiation and major plate boundary reorganization events (e.g. Gurnis et al., 2004; Stern, 2004). However, a revised age of ~52 Ma for Izu–Bonin–Mariana subduction in the western Pacific has been proposed based on dating forearc basalts from the Bonin Ridge (Ishizuka et al., 2011). This closely coincides with the 55 Ma subduction initiation event in the southwest Pacific. Therefore, confirming if subduction initiation at 55 Ma was associated with a short-lived northeast-dipping subduction zone in the South Loyalty Basin, rather than early initiation of west-dipping Tonga–Kermadec subduction is important as it has implications for determining if and how Paleocene–Eocene plate boundary changes in the west and southwest Pacific were related to each other. A better understanding of the timing of major events in the region will help constrain the sequence and driving mechanisms of those events.

4.5.1. Earlier plate motion changes between chrons 27 and 26 (~61.2–57.9 Ma)

If subduction initiation at ~55 Ma was driven by plate motion changes, and forced convergence at a lithospheric zone of weakness (Gurnis et al., 2004), then it is important to consider the history of plate motions in the southwest Pacific leading up to the initiation in order to identify or eliminate possible driving mechanisms. A regional southwest Pacific plate reorganization has been identified at chron 27o to chron 26o time (~61.2–57.9 Ma) (Cande et al., 1995; Eagles et al., 2004a) and correlates with events adjacent to West Antarctica (Cande et al., 1995) and in the Tasman Sea (Gaina et al., 1998). Notably, the cessation of independent motion of the Bellingshausen Plate off West Antarctica occurred at ~61 Ma (Cande et al., 1995). Cande et al. (1995) also describe a slow down in the rate, and counterclockwise change in the direction of spreading at the Pacific–Antarctic ridge, and associated changes in fracture zone trends. Between chrons 27o and 26o there was spontaneous formation of the Pitman fracture zone, a right-stepping fracture zone between two left-stepping fracture zones. In the Tasman Sea Gaina et al. (1998) identified a counterclockwise change in spreading at chron 27o time. In the Coral Sea region further north Gaina et al. (1999) identified a major plate reorganization soon after at chron 26o time (~57.9 Ma). During this reorganization the preceding extensional regime that had produced Coral Sea, Louisiade Trough and Cato Trough seafloor spreading ended and strike-slip motion predominated at two arms of the triple junction that separated the Australian plate, Louisiade Plateau and Mellish Rise continental block. Whether these southwest Pacific-wide plate boundary readjustments are related to subduction initiation or any of the early Eocene plate motion changes outlined above is uncertain and beyond the scope of this investigation. However these observations are important for future investigations into subduction initiation, particularly for comparisons with numerical modeling output.

5. Conclusions

We combined geologic observations from the southwest Pacific with kinematic data to determine how the region evolved from the Late Cretaceous to mid Eocene (~85–45 Ma), including the nature and timing of plate boundary activity. This allowed us to place tighter constraints on the time-dependent evolution of the southwest Pacific regional plate circuit so that motion between the plate pairs is consistent with geologic observations and known tectonic regimes.

Based on our analysis of currently available geologic data combined with a kinematic analysis of alternative plate circuits we favour a model for southwest Pacific evolution (Fig. 17), after Whattam et al. (2008) and Crawford et al. (2003), in which subduction has been active to the east of the LHR and north of New Zealand since at least 85 Ma (Fig. 2, Model 1b). From at least 85 Ma, and possibly 100 Ma, until 55 Ma the South Loyalty Basin opened to the east of New Caledonia associated with west-directed slab roll-back. At ~55 Ma northeast dipping subduction initiated in the South Loyalty Basin and consumed the basin over the period ~55–45 Ma.

Between at least 74 Ma and the initiation of northeast dipping subduction at ~55 Ma there was little to no net relative motion between the Pacific plate and LHR. In our preferred tectonic scenario motion is minimized at a plate boundary through New Zealand where little tectonic activity is recorded during this time period (Sutherland, 1995; Sutherland, 2008). From 74 to 55 Ma both the Australian and Antarctic plate circuits, which use an updated Australia–East Antarctica spreading history (Whittaker et al., 2013), use East–West Antarctica motions consistent with recently revised kinematics for the WARS (Granot et al., 2013), and infer motion in the WARS prior to spreading in the Adare Trough at ~43 Ma, produce similar results when adopted in southwest Pacific reconstruction models. Although we proposed that at least two plate boundaries separated the Pacific plate from the LHR, as there was little to no net relative motion between these plates then the Australian plate circuit, which treats the Pacific plate and LHR as essentially a single plate, will produce similar kinematic results to the Antarctic circuit.

Previous studies have invoked Pacific–LHR plate boundary inception at 45 Ma (e.g. Steinberger et al., 2004; Mortimer et al., 2007; Sutherland et al., 2010; Doubrovine et al., 2012), however we revise this by at least 10 Myr, as subduction initiation occurred at 55 Ma to the east of New Caledonia. For the period from 55 Ma until present-day we advocate using an Antarctic plate circuit for southwest Pacific reconstructions. The Australian plate circuit models that were tested (i.e. assuming no LHR–Pacific motion) predict significant net compression in the WARS during the period 52–45 Ma (Fig. 15c), which is not supported by available geologic observations — thus, the Australian plate circuit is unlikely to be valid during this time period and there was net relative motion between the Pacific plate and LHR.

Our review has emphasized the need for ongoing acquisition of geologic and geophysical data from the southwest Pacific, as much of its evolution is currently poorly constrained due to scarce and ambiguous data. Our preferred model is able to satisfy the available observations, yet must be reassessed when more data become available. The continual formation of the South Loyalty Basin from the mid or Late Cretaceous until the latest Paleocene is a key observation in support of this model, particulary in the absence of strong evidence for arc volcanism throughout this time, so better constraints on its formation are essential. Our preferred scenario makes testable predictions that can be addressed by future work. For instance, it is predicted that the unidentified arc associated with west-dipping subduction that opened the South Loyalty Basin is buried beneath the Tonga–Kermadec Arc and possibly extinct Vitiaz Arc (Eissen et al., 1998; Meffre et al., 2012; Falloon et al., 2014), and the Loyalty Ridge is an extinct arc associated with northeast-dipping subduction that ended with obduction of the Poya Terrane and Peridotite Nappe (e.g. Eissen et al., 1998; Cluzel et al., 2001; Crawford et al., 2003; Whattam et al., 2008).

More data are needed within each spreading system in the southwest Pacific to better constrain plate kinematics prior to 74 Ma during breakup and early seafloor spreading, and quantify deformation within continental crust within Zealandia. A better understanding of what processes led to the major reorganization of plate boundaries at about 105–100 Ma (Veevers, 2000; Matthews et al., 2012), and how this event shaped early fragmentation of the region (Rey and Müller, 2010), is necessary for tying the Late Cretaceous to present-day evolution of the southwest Pacific to the earlier Mesozoic history of long-lived subduction beneath Eastern Gondwanaland.

The rotations for our southwest Pacific kinematic model, along with coastlines, and arc and continental block polygons can be downloaded from the following location: ftp://ftp.earthbyte.org/papers/Matthews_etal_southwest_Pacific.

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