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Reconciling the Cretaceous breakup and demise of the Phoenix Plate with East Gondwana orogenesis in New Zealand

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ABSTRACT

Following hundreds of millions of years of subduction in all circum-Pacific margins, the Pacific Plate started to share a mid-ocean ridge connection with continental Antarctica during a Late Cretaceous south Pacific plate reorganization. This reorganization was associated with the cessation of subduction of the remnants of the Phoenix Plate along the Zealandia margin of East Gondwana, but estimates for the age of this cessation from global plate reconstructions (~86 Ma) are significantly younger than those based on overriding plate geological records (105–100 Ma). To find where this discrepancy comes from, we first evaluate whether incorporating the latest available marine magnetic anomaly interpretations change the plate kinematic estimate for the end of convergence. We then identify ways to reconcile the outcome of the reconstruction with geological records of subduction along the Gondwana margin of New Zealand and New Caledonia. We focus on the plate kinematic evolution of the Phoenix Plate from 150 Ma onward, from its original spreading relative to the Pacific Plate, through its break-up during emplacement of the Ontong Java Nui Large Igneous Province into four plates (Manihiki, Hikurangi, Chasca, and Aluk), through to the end of their subduction below East Gondwana, to today. Our updated reconstruction is in line with previous compilations in demonstrating that as much as 800–1100 km of convergence occurred between the Pacific Plate and Zealandia after 100 Ma, which was accommodated until 90–85 Ma. Even more convergence occurred at the New Zealand sector owing to spreading of the Hikurangi Plate relative to the Pacific Plate at the Osbourn Trough, with the most recent age constraints suggesting that spreading may have continued until 79 Ma. The end of subduction below most of East Gondwana coincides with a change in relative plate motion between the Pacific Plate and East Gondwana from westerly to northerly, of which the cause remains unknown. In addition, the arrival of the Hikurangi Plateau in the subduction zone occurred independent from, and did not likely cause, the change in Pacific Plate motion. Finally, our plate reconstruction suggests that the previously identified geochemical change in the New Zealand arc around 105–100 Ma that was considered evidence of subduction cessation, may have been caused by Aluk-Hikurangi ridge subduction instead. The final stages of convergence before subduction cessation must have been accommodated by subduction without or with less accretion. This is common in oceanic subduction zones but makes dating the cessation of subduction from geological records alone challenging.

1. Introduction

During the Late Cretaceous, an important tectonic change occurred in the southern Pacific realm. For hundreds of millions of years, including most of the Mesozoic, the Panthalassa (or Paleo-Pacific) Ocean was surrounded by subduction zones that consumed oceanic lithosphere of the Farallon (NE), Izanagi (NW), and Phoenix (S) plates (e.g., Engebretson et al., 1985; Seton et al., 2012; Wright et al., 2016; Müller et al., 2019; Torsvik et al., 2019; Boschman et al., 2021a). During the Cretaceous, however, subduction ended along the Zealandia sector of the East Gondwana continental margin (e.g., Bradshaw, 1989; Luyendyk, 1995; Davy et al., 2008; Matthews et al., 2012). Sections of the suture of the Mesozoic subduction zone are located along the northern margin of the Chatham Rise and along the Thurston Island sector of

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Antarctica, which are presently separated from each other by the Pacific-Antarctic Ridge (Fig. 1). This implies that when Pacific-Antarctic spreading started, the ridge did not simply replace the former East Gondwana subduction zone. Instead, it cut through the subduction zone suture and formed partly intra-oceanic and partly intra-continental within East Gondwana lithosphere (Larter et al., 2002; Wobbe et al., 2012). Around the time of subduction cessation, several oceanic plates that formed after breakup of the Phoenix Plate, as well as part of Zealandia, merged with the Pacific Plate. Since then, the Pacific Plate has been diverging from West Antarctica, accommodated by oceanic spreading at the Pacific-Antarctic Ridge (Fig. 1B). But despite its importance in the plate tectonic history of the Panthalassa/Pacific domain, the southwest Pacific-East Gondwana plate reorganization is surrounded with uncertainty.

The uncertainty surrounding the southwest Pacific-East Gondwana plate reorganization results from a discrepancy in the age of subduction cessation between different studies. On the one hand, geologists studying the magmatism and deformation in the orogen located at the overriding plate margin of New Zealand have found no conclusive evidence that shows that subduction must have continued beyond 105–100 Ma (e.g., Bradshaw, 1989; Luyendyk, 1995; Mortimer et al., 2019; Crampton et al., 2019). A 105–100 Ma age estimate for subduction cessation is commonly inferred from a change in deformation within New Zealand from largely compression to a regime dominated by extension (e.g., Bradshaw, 1989; Luyendyk, 1995; Crampton et al., 2019), coeval changes in the geochemical signature of magmatism (Muir et al., 1997; Waight et al., 1998; Tulloch and Kimbrough, 2003; Tulloch et al., 2009; Van der Meer et al., 2016, 2017, 2018), and angular unconformities in the New Zealand forearc (Laird and Bradshaw, 2004; Crampton et al., 2019; Gardiner et al., 2021, 2022). On the other hand, global plate reconstructions suggest that convergence across the Zealandia margin of New Zealand continued until at least the end of spreading in the Osbourn Trough, of which estimates vary from ~101 Ma to 79 Ma, based on dredge samples and tentative marine magnetic anomaly identification (Billen and Stock, 2000; Worthington et al., 2006; Seton et al., 2012; Zhang and Li, 2016; Mortimer et al., 2019), with widely-used global plate reconstructions (Seton et al., 2012; Matthews et al., 2016; Müller et al., 2019) inferring an 86 Ma age that follows Worthington et al. (2006). This age, however, is based on interpretations of the New Zealand geological record that is disputed by many geologists that study New Zealand (e.g., Crampton et al., 2019; Mortimer et al., 2019). Reconciling the geological and plate kinematic estimates of the age of subduction cessation therefore requires using kinematic data from the oceanic and continental domain that are independent from interpreted ages of subduction cessation to avoid circular reasoning in making reconstruction choices.

To do so, we analyse the end of East Gondwana subduction along the Zealandia margin by reassessing both the plate kinematic and orogenic perspectives. First, we evaluate whether the age for the end of convergence suggested by global plate models changes by using the latest, and most detailed published marine magnetic anomaly-based isochrons, and by using the range of estimates for the arrest of Osbourn Trough spreading based on magnetic anomalies or dredge samples. Our reconstruction includes the evolution and fragmentation of the Phoenix Plate into its several daughter plates. We use the recent study of Torsvik et al. (2019) who revisited and modified absolute Pacific Plate models and updated earlier global plate reconstructions. We consider relative motions across the East Gondwana continental margin as a function of absolute plate motion models to evaluate when convergence may have ended, and which process may have been responsible for this cessation. Furthermore, we review aspects of the architecture and evolution of the Cretaceous New Zealand orogen, and attempt to reconcile the timing of the end of subduction with the available geological evidence. We will use our results as a basis for the reconstruction of the demise of the Phoenix Plate's daughters, which resulted from their capture by the Pacific Plate after cessation of subduction along the Gondwana margin and the enigmatic transition to the Pacific-Antarctic spreading ridge.

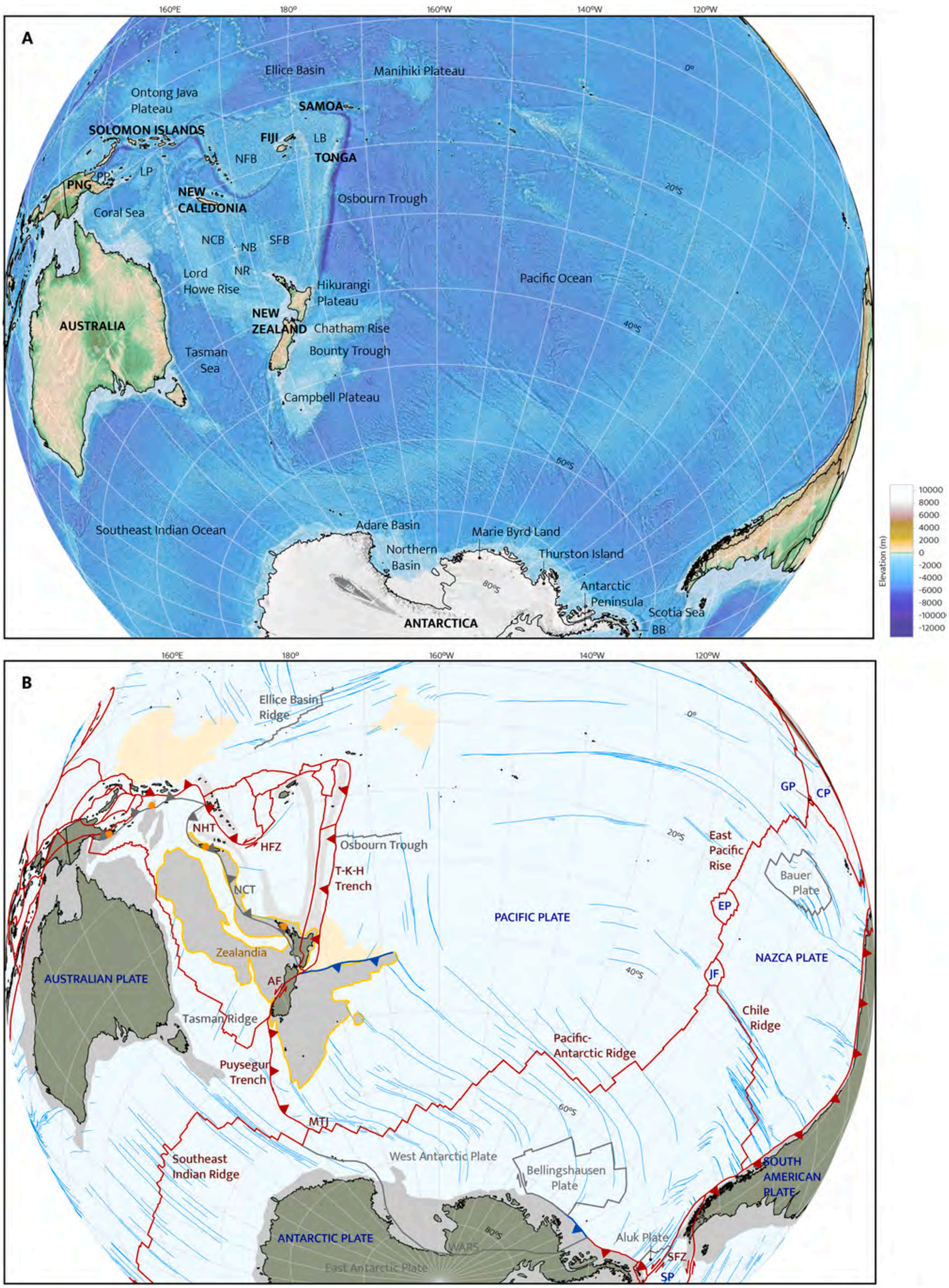
2. Plate tectonic setting

The south Pacific Ocean today is underlain by the Pacific, Antarctic, and Nazca plates, separated by trenches from the South American, Antarctic, and Australian plates (Fig. 1). The oceanic plates of the Pacific Ocean are separated from each other by mid-ocean ridges: the Pacific-Antarctic Ridge, the East Pacific Rise between the Pacific and Nazca plates, and the Chile Ridge between the Antarctic and Nazca plates (Fig. 1). Three microplates are present along the East Pacific Rise: The Juan Fernandez, Easter, and Galapagos microplates.

In the west, the Pacific Plate is currently subducting below the Australian Plate at the Tonga-Kermadec-Hikurangi subduction zone. To the west of this subduction zone is a series of Cenozoic back-arc basins (e.g., Lau Basin, South Fiji Basin, Norfolk Basin; e.g., Yan and Kroenke, 1993; Sdrolias et al., 2003; Herzer et al., 2011), bounded in the west by the extended continental crust of Zealandia that underlies the Lord Howe Rise and Norfolk Ridge (e.g., Mortimer et al., 2017). Zealandia is separated from the Australian continent by the Upper Cretaceous-Paleogene Tasman Sea and New Caledonia basins (e.g., Gaina et al., 1998; Grobys et al., 2008). The Norfolk Ridge was overthrust from the east during the Oligocene (c. 30 Ma) by the Paleocene New Caledonia ophiolite (Cluzel and Meffre, 2002; Cluzel et al., 2012). Ophiolite obduction occurred during cessation of a northeast-dipping intra-oceanic subduction zone that formed around 60 Ma. At this time other ophiolites also formed that were emplaced during the Late Oligocene onto Northland, New Zealand, and northward towards the Louisiade Plateau and the eastern Papuan Peninsula (Fig. 1) (Whattam et al., 2006; Cluzel et al., 2012; Van de Lagemaat et al., 2018a; Maurizot et al., 2020a; McCarthy et al., 2022).

At its northern end, southwest of Samoa, the Tonga-Kermadec-Hikurangi subduction turns sharply to the west (Fig. 1). Here the plate boundary changes to a SW-trending, diffuse transform system around the Fiji Islands, southwest of which it continues as the Hunter fracture zone that connects to the New Hebrides Trench. At this trench the Australian Plate is subducting below the North Fiji back-arc basin that hosts spreading ridges with the Pacific Plate. The southern end of the Tonga-Kermadec-Hikurangi subduction zone connects via the right-lateral Alpine Fault to the Puysegur Trench where subduction of the Australian Plate below the Pacific Plate is occurring (e.g., Collot et al., 1995; House et al., 2002; Gurnis et al., 2019). The plate boundary ends at the Macquarie Triple Junction, where the Australian, Pacific, and Antarctic plates meet, and where the Macquarie microplate formed c. 7 Ma (Cande and Stock, 2004a; Choi et al., 2017). Kinematic reconstructions of Cenozoic tectonic history of the SW Pacific realm differ in the timing and distribution of convergence over the New Caledonia and Tonga-Kermadec subduction zones (Hall, 2002; Schellart et al., 2006; Whattam et al., 2008; Van de Lagemaat et al., 2018a), but mostly agree on the pre-late Cretaceous position of Zealandia against the Australian continent, and on the location of the subduction zone along the eastern Zealandia margin that consumed the Phoenix Plate and its daughters (Fig. 2).

The southern boundary of the Pacific Plate is the Pacific-Antarctic Ridge (Fig. 1B). This plate boundary formed c. 89 Ma, based on the extrapolation of spreading rates from the oldest identified marine magnetic anomaly (C34y; 83.7 Ma) towards the continental margin (Wobbe et al., 2012). This age is in correspondence with the 83.9 $\pm$ 0.1 Ma age of the Erik seamount, obtained from Ar/Ar dating of K-feldspar of a trachyte sample, which provides a minimum age of the oceanic crust (Mortimer et al., 2019). The Pacific-Antarctic Ridge accommodated the divergence of the Campbell Plateau (part of the Zealandia continent, located on the Pacific Plate) from Marie Byrd Land (located on the West Antarctic Plate) (e.g., Wobbe et al., 2012). Before break-up, the Campbell Plateau and West Antarctica formed part of the upper plate adjacent to the Mesozoic active margin of East Gondwana (Fig. 2) (e.g., Larter



(caption on next page)

Fig. 1. Present-day geographic and tectonic maps of the South Pacific region.
A) Geographic map.
BB: Bransfield Basin; LB: Lau Basin; LP: Louisiade Plateau; NB: Norfolk Basin; NCB: New Caledonia Basin; NFB: North Fiji Basin; NR: Norfolk Ridge; PNG: Papua New Guinea; PP: Papuan Peninsula; SFB: South Fiji Basin. Background image is ETOPO1 1 Arc-Minute Global Relief Model. (Amante and Eakins, 2009; NOAA National Geophysical Data Center, 2009).
B) Tectonic map. Current tectonic plate names in blue. Current plate boundaries (Bird, 2003) and plate boundary names in red, former plate names and plate boundaries in grey, suture of East Gondwana subduction zone in blue. Continental crust of Zealandia is outlined in yellow. Northland, New Caledonia, Louisiade, and Papuan Peninsula ophiolites are indicated by orange dots. Ontong Java Nui Large Igneous Provinces in light yellow. Lightblue lines are digitalized fracture zones, obtained from the GSFML database (Matthews et al., 2011; Wessel et al., 2015)
AF: Alpine Fault; CP: Cocos Plate; EP: Easter Plate; GP: Galapagos Plate; HFZ: Hunter Fracture Zone; JF: Juan Fernandez Plate; MS: Manihiki Scarp; MTJ: Macquarie Triple Junction; NHT: New Hebrides Trench; NCT: New Caledonia Trench; SFZ: Shackleton Fracture Zone; SP: Scotia Plate; T-K-H: Tonga-Kermadec-Hikurangi; WARS: West Antarctic Rift System.

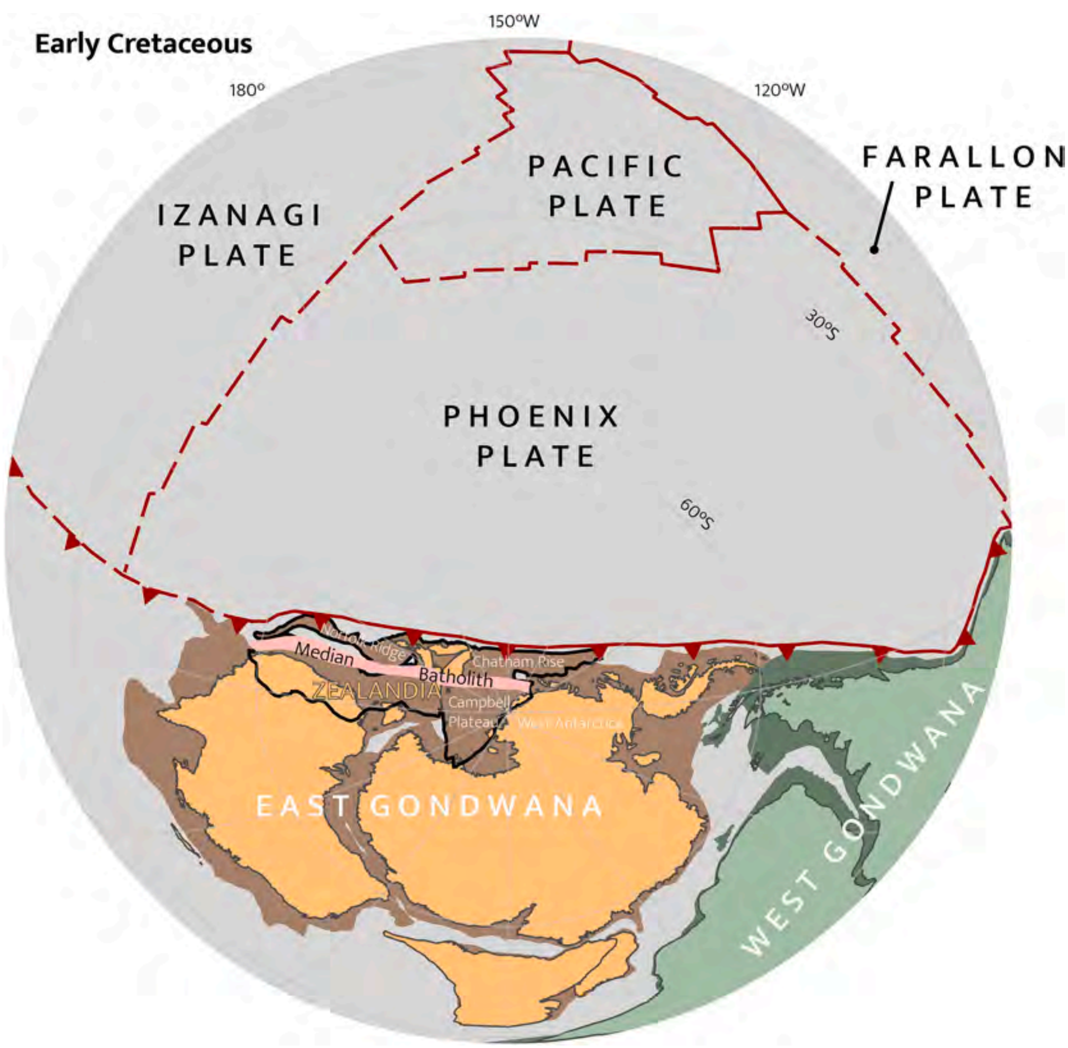


Fig. 2. Early Cretaceous (c. 140 Ma) reconstruction of the paleo-Pacific/Panthalassa realm showing the approximate extent of the Phoenix Plate. Continental crust of future Zealandia is outlined in black, highlighting the Zealandia margin of East Gondwana. Plate boundaries are only shown in the Panthalassa domain, dashed where the location of the plate boundary is estimated.

et al., 2002). This margin was contiguous with the active margins of the Antarctic Peninsula and South America, where subduction remains active today. Presently, subduction of a small remnant of the last of the Phoenix Plate's daughters, the Aluk Plate (Herron and Tucholke, 1976), is ongoing below the northern part of the Antarctic Peninsula (Fig. 1) (e.g. Eagles, 2004). The Aluk Plate is often also referred to as Phoenix Plate, but we prefer the name Aluk Plate to make the distinction with the original parent Phoenix Plate. Subduction below Antarctica progressively ceased with the arrival of different segments of the Aluk-Antarctica Ridge. A small segment of this ridge remains in the Southeast Pacific Ocean, which became extinct c. 3.3 Ma (Eagles, 2004), effectively merging the Aluk Plate with the Antarctic Plate. Subduction below the Antarctic Peninsula is currently accommodated by opening of the Bransfield Basin within the upper Antarctic Plate (Fig. 1; Galindo-Zaldivar et al., 2004). The eastern boundary of the Aluk Plate is the Shackleton Fracture Zone, separating it from the West Scotia Sea. Opening of the Scotia Sea oceanic basins was not related to plate motions of the paleo-Pacific realm (Van de Lagemaat et al., 2021) and the Shackleton Fracture Zone is thus the eastern boundary of our reconstruction. To the north of the Shackleton Fracture Zone, the Antarctic

Plate, Chile Ridge, and Nazca Plate are subducting below South America.

Mesozoic subduction of Phoenix Plate lithosphere was accommodated along the Antarctica and Zealandia margins of East Gondwana (Fig. 2). Breakup of these continents from each other and from Australia led to oceanic spreading around 84 Ma in both the Tasman Sea and South Pacific Ocean (Gaina et al., 1998; Wobbe et al., 2012; Mortimer et al., 2019), but continental rifting between Zealandia and Antarctica and between Zealandia and Australia has been considered to date back to c. 105–100 Ma (Bradshaw, 1989; Luyendyk, 1995; Laird and Bradshaw, 2004). Earliest extension between Australia and Antarctica started at c. 136 Ma (Whittaker et al., 2013).

A prominent record of Mesozoic subduction is present in New Zealand. The Eastern Province consists of Permian intra-oceanic arc sequences and a long-lived Mesozoic accretionary wedge (Fig. 3) (Mortimer, 2004; Mortimer et al., 2014). It is possible that the Eastern Province hosts the records of two subduction systems, one along the Gondwana margin and one intra-oceanic (Adams et al., 2007; Van de Lagemaat et al., 2018b; Campbell et al., 2020), but these have been juxtaposed since at least the latest Jurassic (Tulloch et al., 1999), i.e., throughout the window of interest of this paper. The western and eastern provinces are separated by the Median Batholith that represents a long-lived Paleozoic to Mesozoic magmatic arc (Figs. 2 and 3) (Mortimer, 2004). The accretionary wedge of the Eastern Province consists of ocean plate stratigraphy (OPS; Isozaki et al., 1990) comprising pillow lavas, oceanic pelagic and hemipelagic sediments, and trench fill clastics (Caples, Waipapa and Torlesse terranes; Mortimer et al., 2014). These OPS sequences accreted to the Gondwana margin from Permian to Early Cretaceous times and were intruded by magmatic arc plutons and overlain by forearc basin clastics (Adams et al., 1998; Adams et al., 2013; Mortimer, 2004; Boschman et al., 2021a). The geology of New Caledonia shares broad similarities with that of New Zealand: The Boghen Terrane of New Caledonia has been correlated to the Torlesse Complex of New Zealand; both Jurassic-Cretaceous accretionary complexes, and the Teremba Terrane of New Caledonia to the Murihiku Terrane of New Zealand; both forearc terranes consisting of late Permian to Jurassic island-arc derived strata (Cluzel and Meffre, 2002; Maurizot et al., 2020b). Cretaceous sedimentary sequences that overlie the Torlesse accretionary complex from 100 Ma onwards in New Zealand (Laird and Bradshaw, 2004; Crampton et al., 2019) provide important arguments for interpreting the end of subduction: they are widely seen as signaling a transition from a subduction margin to a passive margin (e.g., Field and Uruski, 1997; Laird and Bradshaw, 2004; Crampton et al., 2019). However, others have considered these Late Cretaceous sequences to be accretionary shelf and slope basin fill that accumulated during outbuilding of the accretionary wedge and that subduction continued until c. 84 Ma (Mazengarb and Harris, 1994; Kamp, 1999, 2000; Gardiner and Hall, 2021). Deposition of subduction-related volcaniclastic greywackes continued until c. 90 Ma in New Caledonia (Cluzel et al., 2010; Maurizot et al., 2020b).

The oceanic lithosphere of the modern Pacific Plate contains three prominent oceanic plateaus interpreted to have formed as a single ~120 Ma Large Igneous Province (LIP): the conceptual Ontong Java Nui LIP (Taylor, 2006; Chandler et al., 2012). The three oceanic plateaus that are thought to have once formed as Ontong Java Nui are currently separated by post-120 Ma Cretaceous oceanic basins. These oceanic plateaus are the Ontong Java Plateau, located to the north of the Solomon Islands; the Manihiki Plateau, located to the northeast of Samoa; and the Hikurangi Plateau, located offshore the North Island of New Zealand (Fig. 1). The Manihiki Plateau is separated from the Ontong Java Plateau by the Ellice Basin, and the Hikurangi Plateau is separated from the Manihiki Plateau by the Osbourn Trough (Fig. 1).

3. Reconstruction approach, plate circuits, and reference frames

Quantitative constraints on the convergence history between the plates of the Panthalassa realm and the Zealandia margin of East Gondwana follows from the kinematic reconstruction of the South Pacific region. The reconstruction presented here includes a compilation of the most recent kinematic data and the new Pacific reference frame of Torsvik et al. (2019). For the analysis in this paper, we focus on the history of the South Pacific region back to the Early Cretaceous. Our reconstruction is made in GPlates, a freely available plate reconstruction software (www.gplates.org; Boyden et al., 2011; Müller et al., 2018).

We restore spreading along the different mid-ocean ridges that existed in the southern Panthalassa realm based on published marine magnetic anomaly data of ocean floor presently underlying the south Pacific Ocean (Fig. 4), reviewed in section 4. The ages of the polarity chrons in our reconstruction are updated to the timescale of Ogg (2020). We incorporate all rotation poles as published, even though on short time intervals (<1 Myr) these are likely subject to some noise (Iaffaldano et al., 2012). Our conclusions, however, are not affected by the short time-scale noise and we prefer to see the effect of all interpreted isochrons rather than an arbitrary selection of these.

In the absence of polarity reversals during the Cretaceous Normal Superchron (121.4–83.7 Ma), the restoration of oceanic basins for this time interval is based on previously published radiometric data from dredged and cored samples as well as published interpretations of seafloor fabric (Fig. 4; see section 4). Magnetic anomaly picks and fracture zone data were obtained from the Global Seafloor Fabric and Magnetic Lineation (GSFML) Database (Matthews et al., 2011; Seton et al., 2014; Wessel et al., 2015). We restore intra-continental deformation within East Gondwana applying a reconstruction hierarchy that uses quantitative kinematic constraints on continental extension, transform motion,

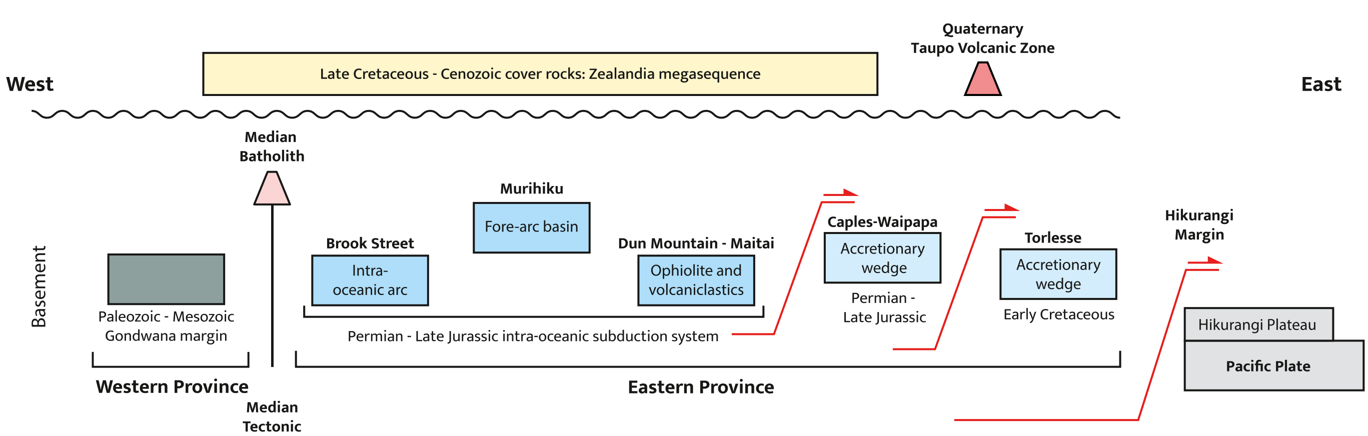


Fig. 3. Schematic cross-section of the present-day geology of the North Island of New Zealand, based on Mortimer et al. (2014).

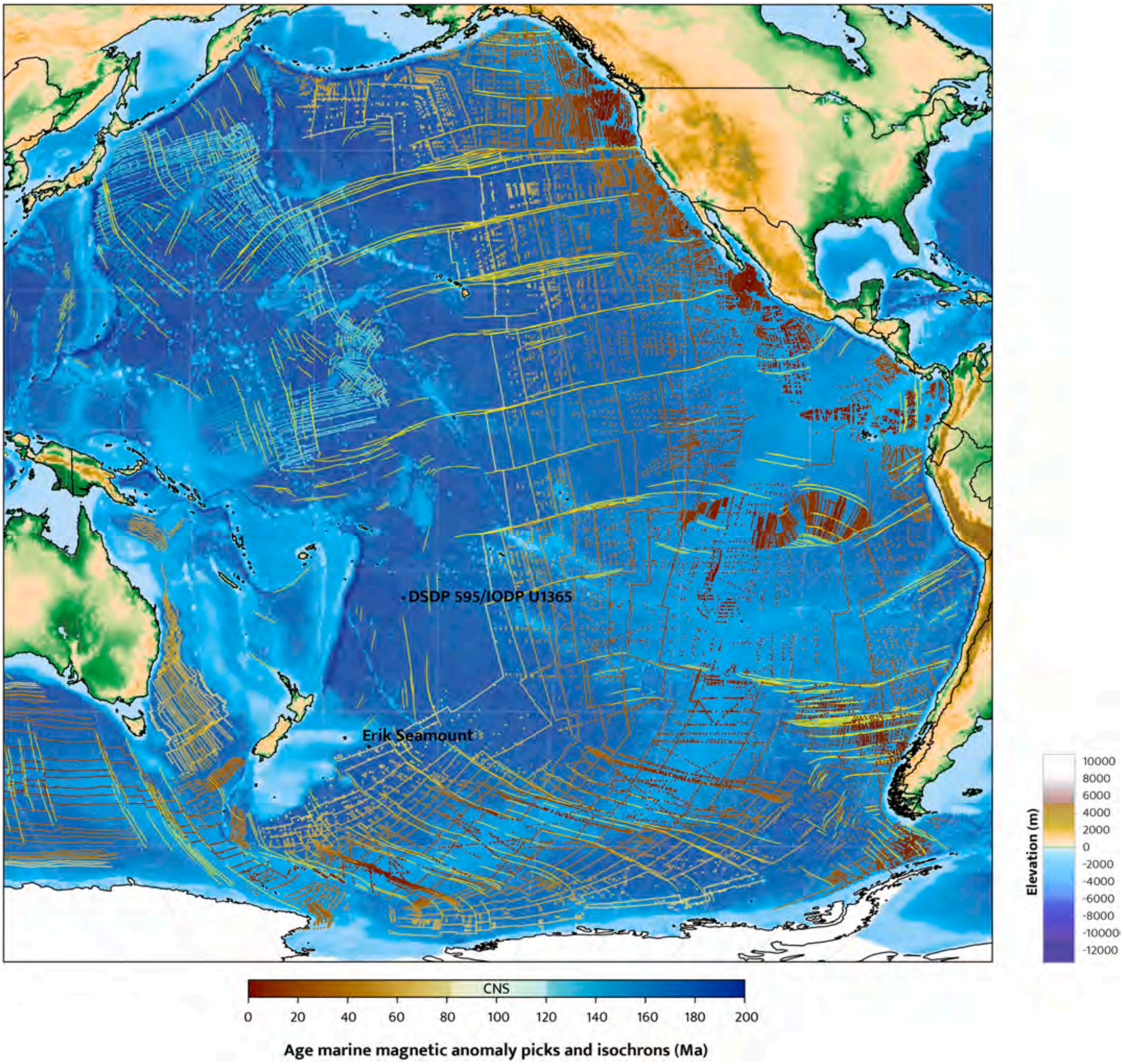


Fig. 4. Geographic map of the Pacific and Southeast Indian oceans, showing the data used for the reconstruction of the oceanic domain. Marine magnetic anomaly picks (colored by age) and fracture zone data (in yellow) were obtained from the GSFML database (Matthews et al., 2011; Seton et al., 2014; Wessel et al., 2015, and references therein). Interpreted isochrons are from Seton et al. (2012), Wright et al. (2016), and from this study. Background image is ETOPO1 1 Arc-Minute Global Relief Model (Amante and Eakins, 2009; NOAA National Geophysical Data Center, 2009). Marine magnetic anomaly picks are colored using a colour bar of Crameri (2018); Crameri et al. (2020). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

or crustal shortening (see Boschman et al., 2014 and Van de Lagemaat et al., 2018a for details).

Plate convergence can best be quantified when a plate circuit is present that connects the two converging plates through a series of active or fossil spreading ridges (Cox and Hart, 1986). For times after the formation of the Pacific-Antarctic Ridge (Chron C34y, c. 83.7 Ma; see section 4), plate convergence in the region can be reconstructed through a plate circuit that constrains the motion of the Australian Plate relative to the Antarctic Plate based on the record of oceanic spreading at the Southeast Indian Ridge (SEIR), and the motion of the Pacific Plate relative to the Antarctic Plate by restoring spreading at the Pacific-Antarctic Ridge (PAR) (Fig. 5). The Late Cretaceous and Cenozoic opening of marginal and back-arc basins east of Australia are reconstructed relative to the Australian Plate, that adds Zealandia-Australia, and Tonga-Kermadec-Hikurangi trench-Zealandia motion to the plate circuit. In addition, the relative motion of oceanic plates flooring the Pacific Ocean are reconstructed relative to the Pacific Plate (Fig. 5). For the period of activity of the New Caledonia subduction zone in Paleocene to Oligocene time, it is not possible to quantify partitioning of convergence over the Tonga and New Caledonia trenches – only net convergence between Zealandia and the Panthalassa plates can be quantified (Van de Lagemaat et al., 2018a). However, for the interval of interest of this paper, this problem is of no consequence.

For times after the Cretaceous Normal Superchron, i.e. at C34y (post-

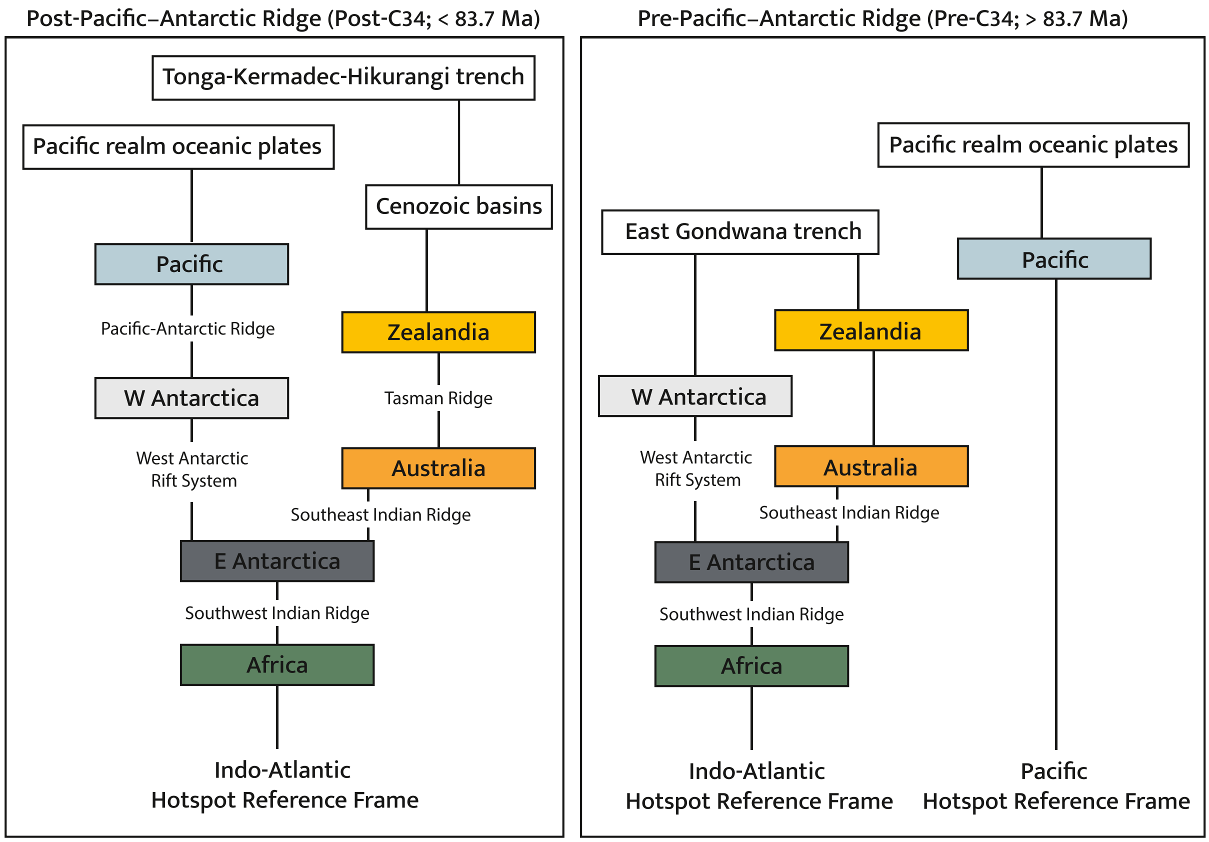


Fig. 5. Plate circuits used in our reconstruction, highlighting the differences in the plate circuit before and after formation of the Pacific-Antarctic Ridge.

83.7 Ma), we use the 'Antarctic' plate circuit Zealandia – Australia – East Antarctica – West Antarctica – Pacific (Fig. 5). Due to uncertainties in relative motion between West Antarctica and East Antarctica before 45 Ma, some studies use a plate circuit for these times that ties the Lord Howe Rise to the Pacific Plate directly before c. 45 Ma instead (i.e., the Australian circuit) using magnetic anomalies in the Tasman Sea basin (e. g., Steinberger et al., 2004; Torsvik et al., 2019). In the Australian circuit it is assumed that there is no plate boundary between the Pacific Plate and Zealandia between 83 and 45 Ma. However, geological data from New Caledonia provides evidence for the existence of a subduction zone between the Pacific Plate and the Norfolk Ridge between c. 60 and 30 Ma (e.g., Cluzel et al., 2012; Van de Lagemaat et al., 2018a; Maurizot et al., 2020b), which means that the Pacific Plate should not be reconstructed relative to Zealandia after 60 Ma. Combined with recently improved constraints on deformation within Antarctica (e.g., Granot et al., 2013; Granot and Dyment, 2018) leads us to prefer the Antarctic circuit for our reconstruction, similar to Seton et al. (2012), Matthews et al. (2015), and Müller et al. (2019). There is a c. 150 km difference in location of the Pacific relative to the Gondwana plates between the Antarctic and Australian circuits at chron C34y (83.7 Ma).

Before the onset of Pacific-Antarctic Ridge spreading, the plate circuit is broken, as the Panthalassa and Gondwana plates are connected through a subduction zone only (Seton et al., 2012; Wright et al., 2016). The reconstruction of pre-chron C34y (83.7 Ma) relative motions across the East Gondwana margin then relies on placing the Gondwana continents and the Panthalassa plates in mantle reference frames that were developed for each of the two systems separately (Fig. 5). For this reason, we put our reconstruction in a mantle reference frame for the entire reconstruction period. The Gondwana continents are part of the Indo-Atlantic realm, whose relative motions are constrained by the reconstruction of the Indian and Atlantic Oceans. Several mantle reference frames are available for the Indo-Atlantic realm, from different approaches and iterations. We will illustrate the sensitivity of the choice of reference frame for convergence across the Zealandia margin, using the moving hotspot reference frames of O'Neill et al. (2005), Torsvik et al. (2008), and Doubrovine et al. (2012), and the semi-quantitative slab-fitted reference frame of Van Der Meer et al. (2010). These reference frames are given in African coordinates, requiring reconstructing the eastern Gondwana continents circuit to the African Plate. For the period after the Cretaceous Normal Superchron, we also use the Indo-Atlantic reference frame for the Panthalassa domain, as it is connected to the plate circuit. For the period before chron C34y (83.7 Ma), when the plate circuit is broken, we use the Pacific reference frame of Torsvik et al. (2019), who updated a fixed hotspot frame that constrains absolute Pacific Plate motion back to 150 Ma. We incorporate the 'Earthbyte Model R' of Torsvik et al. (2019), which corresponds to the Antarctic circuit as explained above.

4. Review of kinematic data

4.1. Post-Cretaceous Quiet Zone plate reconstruction of ocean basins, and East Gondwana fit

The onset of spreading at the Pacific-Antarctic Ridge marks a major break in the plate tectonic history of the Panthalassa-Pacific realm, as it formed the first passive margin that connected the oceanic domain to the Indo-Atlantic plates after hundreds of millions of years (e.g., Molnar et al., 1975; Seton et al., 2012; Wright et al., 2016; Müller et al., 2019). The oldest magnetic anomaly that records spreading between the Campbell Plateau and Marie Byrd Land (West Antarctica) is chron C33 (79.9 Ma; Wobbe et al., 2012), the oldest crust having formed after the end of chron C34y, i.e., after 83.7 Ma. Farther east, however, the marine magnetic anomaly of chron C34y (83.7 Ma) was identified just south of Chatham Rise and its conjugate margin off the coast of Thurston Island (Larter et al., 2002; Eagles et al., 2004a; Wobbe et al., 2012). There is no evidence for the existence of a plate boundary between the oceanic crust that formed south of the Chatham Rise and the Campbell Plateau and oceanic crust of the Pacific Plate, and it is therefore assumed that the Chatham Rise and Campbell Plateau have been part of the Pacific Plate since the formation of the Pacific-Antarctic Ridge (Molnar et al., 1975; Luyendyk, 1995). The set of marine magnetic anomalies of chron C34y (83.7 Ma) that formed south of Chatham Rise and off the coast of Thurston Island is therefore the oldest marine magnetic anomaly constraint for Pacific-West Antarctica spreading (Wobbe et al., 2012; Wright et al., 2016). As these marine magnetic anomalies are located close to the continental margins of Chatham Rise and West Antarctica, it is thought that true seafloor spreading started shortly before the end of the Cretaceous Quite Zone (Wobbe et al., 2012). Based on the extrapolation of seafloor spreading rates, Wobbe et al. (2012) suggested that the first oceanic crust between Chatham Rise and Thurston Island (West Antarctica) formed around 84 Ma, which is in accord with the minimum age for the oceanic crust between Chatham Rise and West Antarctica, based on a 83.9 ± 0.1 Ma Ar/Ar age of K-feldspar in a trachyte sample from Erik Seamount (Mortimer et al., 2019), while rifting is thought to have started around 89 Ma (Wobbe et al., 2012). The oldest oceanic crust between Chatham Rise and West Antarctica may have formed during extension in the Bounty Trough (between 92 and 84 Ma; Grobys et al., 2008), before Chatham Rise was captured by the Pacific Plate. The timing of the capture of Chatham Rise by the Pacific remains uncertain, although it must have occurred in the 90–83.7 Ma interval: the location of the Pacific Plate is constrained at either end of this time interval: 90 Ma is the youngest age in the Pacific hotspot reference frame of Torsvik et al. (2019) and 83.7 Ma (i.e., chron C34y) is the oldest marine magnetic anomaly constraint (Wright et al., 2016). Between those times (90–83.7 Ma), the Pacific Plate may have started to diverge from West Antarctica, but we reconstruct the start of Pacific-Antarctic spreading based on the oldest marine magnetic anomaly constraint (i.e., C34y; 83.7 Ma; Wobbe et al., 2012), similar to other reconstructions (e.g., Seton et al., 2012; Wright et al., 2016; Müller et al., 2019). Any extension in the region (i.e., between Chatham Rise and Campbell Plateau and West Antarctica) before that time is considered to not have involved the Pacific Plate (Fig. 6D and Fig. 7C). We reconstruct the motion between the Pacific Plate and West Antarctica using finite rotation poles of Croon et al. (2008) (present-C20; 43.5 Ma) and Wright et al. (2016) (C21-C34y; 47.8–83.7 Ma). This is similar to the reconstruction of Müller et al. (2019), although we incorporate all published rotation poles whereas Müller et al. (2019) only used rotation poles for selected polarity chrons.

Shortly before the start of chron C33 (79.9 Ma) a piece of lithosphere broke off of West Antarctica to form the Bellingshausen Plate (Stock and Molnar, 1987). The Bellingshausen Plate started to rotate clockwise relative to West Antarctica and acted as an independent plate until chron C27 (62.5 Ma; Stock and Molnar, 1987; Cande et al., 1995; Eagles et al., 2004b; Wobbe et al., 2012; Wright et al., 2016). During this time window, the northern margin of the Bellingshausen Plate was formed by a spreading ridge with the Pacific Plate and its western margin was defined by a short transform margin with the Marie Byrd Land sector of West Antarctica, close to the Euler pole of Bellingshausen-West Antarctica motion (Wright et al., 2016). To the east, the Bellingshausen Plate was bounded by a right-lateral transform fault from the Aluk Plate (Larter et al., 2002; Eagles et al., 2004b). To the south, the Bellingshausen Plate was converging with the Thurston Island sector of West Antarctica, although the maximum total amount of convergence was <250 km and no mature subduction zone developed (Wright et al., 2016). Like the reconstruction of Müller et al. (2019), we reconstruct the 79.9–62.5 Ma motion of the Bellingshausen Plate relative to the Pacific Plate using the finite rotation poles of Wright et al. (2016), which are based on marine magnetic anomalies of chrons C33-C27.

We tentatively suggest that friction at the transform fault that formed the eastern plate boundary of the West Antarctic Plate (Heezen Fracture Zone) led to the partial coupling of West Antarctica with the Aluk Plate. After the formation of the Pacific-Antarctic ridge, the Pacific-West Antarctica spreading and Pacific-Aluk spreading ridges were parallel, but Pacific-Aluk spreading occurred at a higher rate (~3.5 and ~ 7 cm/yr half-spreading rate, respectively). This resulted in lengthening of the transform fault that formed the plate boundary between the West Antarctic and Aluk plates. The partial coupling of part of West Antarctica with Aluk caused the formation of the Bellingshausen Plate, which was being dragged along by the Aluk Plate. This dragging resulted in clockwise rotation of Bellingshausen relative to West Antarctica. This suggestion is similar to what was proposed by Eagles et al. (2004b), who also suggested that the independent motion of Bellingshausen was related to the lengthening of the West Antarctica-Pacific transform plate boundary. A similar process was responsible for the formation of e.g. the Bauer microplate, which moved independently between c. 18–6 Ma, partitioning strain between the Pacific and Nazca plates (Eakins and Lonsdale, 2003).

Before C34y (83.7 Ma), pre-drift extension had already started to separate the Campbell Plateau from Marie Byrd Land and Chatham Rise from the Campbell Plateau (Molnar et al., 1975; Stock and Cande, 2002; Riefstahl et al., 2020) (Fig. 6D and 7C). We reconstruct c. 180 km of pre-drift extension between the Campbell Plateau and Marie Byrd Land between 95 and 83.7 Ma, based on the estimated 90 km of extension in both margins based on crustal thickness calculations (Wobbe et al., 2012). We reconstruct c. 200 km of extension in the Bounty Trough (between Chatham Rise and the Campbell Plateau) between 92 and 84 Ma, based on the reconstruction derived from crustal thickness calculations of Grobys et al. (2008). This reconstruction also leads to extension between Chatham Rise and the Thurston Island sector of West Antarctica, where plate boundary activity may have started around 89 Ma (Wobbe et al., 2012).

A long-lived volcanic arc and accretionary prism on the Antarctic Peninsula shows that subduction continued throughout the Mesozoic and Cenozoic until the present-day (e.g., Burton-Johnson and Riley, 2015; Jordan et al., 2020). The plate that is subducting below the Antarctic Peninsula, the Aluk Plate, is therefore thought to be a descendent of the Phoenix Plate (Fig. 1; e.g., Barker, 1982; Eagles, 2004). Interestingly, however, for much of the Cenozoic, and until the cessation of spreading around 3.3 Ma, the Aluk Plate has not been spreading relative to the Pacific Plate, but relative to oceanic lithosphere of West Antarctica (Eagles, 2004). Marine magnetic anomalies that formed along the Aluk-West Antarctica Ridge are preserved on the Aluk Plate back to C6A (21.32 Ma; Larter and Barker, 1991; Eagles, 2004), and on conjugate West Antarctica oceanic lithosphere back to C27 (62.52 Ma) (Cande et al., 1982). To the northwest, West Antarctica also contains a set of magnetic anomalies from C21 (47.8 Ma) and younger that record spreading between West Antarctica and the Pacific Plate (Cande et al., 1982; Cande et al., 1995; Croon et al., 2008). This spreading was near-parallel to West Antarctica-Aluk spreading, showing simultaneous and near-parallel spreading of West Antarctica with both the Pacific and Aluk plates (Wright et al., 2016).

Around the time of chron C21 (c. 47 Ma), part of the Pacific Plate that formed through Pacific-Aluk spreading was captured by the West Antarctic Plate (Cande et al., 1982; McCarron and Larter, 1998; Eagles et al., 2004a). Shortly before capture, the transform plate boundary between the West Antarctic and Pacific plates was lengthening due to the higher Pacific-Aluk compared to Pacific-West Antarctic spreading rates, similar to the situation that resulted in the formation of the Bellingshausen Plate. During capture, the Pacific-Antarctic ridge propagated into oceanic crust of the Pacific Plate that formed around C27 (c. 62.5 Ma) (Cande et al., 1982). At the southern end of the captured crust, the Pacific-Aluk ridge was replaced by the West Antarctic-Aluk ridge.

To the northeast, West Antarctica shared a spreading ridge with the Farallon Plate and its daughter Nazca Plate (Fig. 6H-J) (Wright et al., 2016). However, before C21 (47.3 Ma), there was no Antarctic oceanic crust that separated the Aluk Plate from the Pacific Plate (Fig. 6G): instead, the Aluk Plate was spreading directly with the Pacific Plate, recorded by marine magnetic anomalies back to C34y (83.7 Ma) on the Pacific Plate (Cande et al., 1982; Cande et al., 1995; Larter et al., 2002;

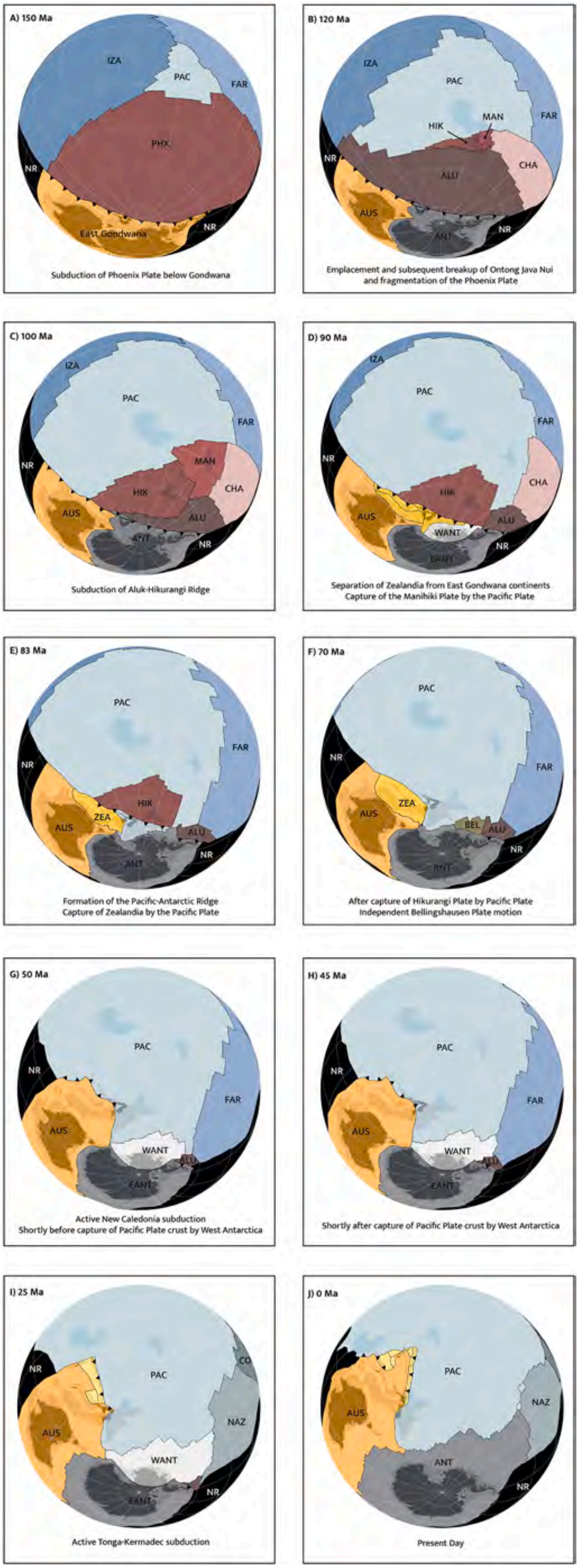


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Fig. 6. Snapshots of our kinematic reconstruction in the Van Der Meer et al. (2010) reference frame, highlighting key events in the evolution of the Phoenix Plate and East Gondwana subduction zone. These events are discussed in the main text. Present-day coastlines and outline of continental crust are shaded behind the plate colors and shown for reference. The Ontong Java Nui Large Igneous Provinces are also outlined, where the outline of the Hikurangi Plateau does not include the parts that were subducted below Chatham Rise and the North Island. Plate names: ALU: Aluk Plate; ANT: Antarctic Plate (EANT: East Antarctic Plate; WANT: West Antarctic Plate); AUS: Australian Plate; CHA: Chasca Plate; FAR: Farallon Plate; HIK: Hikurangi Plate; IZA: Izanagi Plate; MAN: Manihiki Plate; NAZ: Nazca Plate; PAC: Pacific Plate; SAM: South American Plate; ZEA: Zealandia. NR = Not Reconstructed.

Eagles et al., 2004a; Croon et al., 2008). In our reconstruction we use rotation poles of Aluk-West Antarctica and Aluk-Pacific motion back to C34y (83.7 Ma) of Eagles (2004), Eagles and Scott (2014) and Wright et al. (2016), similar to Müller et al. (2019). The tectonic history of the Aluk Plate prior to C34y (83.7 Ma) cannot be constrained by magnetic anomalies due to the Cretaceous Quiet Zone.

In the East Pacific, the Nazca Plate is spreading along the East Pacific Rise from the Pacific Plate and along the Chile Ridge from West Antarctica, while subducting below South America (Fig. 1). The Nazca Plate formed c. 22 Ma (chron C6B), as the southern remnant of the broken up Farallon Plate (Barckhausen et al., 2001, 2008; Wright et al., 2016). The East Pacific Rise records spreading between the Pacific and Nazca plates (and its predecessor the Farallon Plate) back to chron C23 (51.7 Ma) on the Nazca Plate (older magnetic anomalies have been lost to subduction below South America) and back to chron C34y (83.7 Ma) on the Pacific Plate (Atwater and Severinghaus, 1989; Barckhausen et al., 2008; Wilder, 2003; Handschumacher, 1976). Spreading between the Nazca Plate and West Antarctica is recorded at the Chile Ridge back to chron C24 (53.9 Ma) on West Antarctica and back to chron C5E (18.5 Ma) on the Nazca Plate (Cande et al., 1982; Tebbens et al., 1997). We reconstruct the Nazca Plate relative to the Pacific Plate, using the finite rotation poles based on marine magnetic anomalies back to chron C6B (22.3 Ma) of Tebbens and Cande (1997), as published in Wright et al. (2016), similar to Müller et al. (2019). We include the Bauer Microplate that formed in Miocene times at the Nazca-Pacific ridge using magnetic anomalies C5E-C3A (18.5–6.7 Ma) identified by Eakins and Lonsdale (2003), with rotations computed in GPlates. We do not include the Galapagos, Easter and Juan Fernandez microplates in our reconstruction, which formed about 5 Ma (Tebbens and Cande, 1997; Wright et al., 2016). The Farallon Plate is reconstructed relative to the Pacific Plate between 22.3 (chron C6B) and 83.7 Ma (chron C34y) using the finite rotations poles of Wright et al. (2016), like in Müller et al. (2019). The record of Farallon-Pacific spreading during and before the Cretaceous Quiet Zone will be discussed in section 4.2.

Cenozoic relative motion between East Antarctica and West Antarctica is constrained by marine magnetic anomalies that formed in the Adare and Northern basins between chrons C5 and C27 (11.1–62.5 Ma) (Cande and Stock, 2004b; Granot et al., 2013; Granot and Dyment, 2018). We incorporate the finite rotation poles of Granot and Dyment (2018), Granot et al. (2013), and Cande and Stock, (2004a, 2004b) for chrons C5-C8, C12-C18, and C20-C27, respectively. Mesozoic extension in the West Antarctic Rift System (WARS) between West Antarctica and East Antarctica is poorly constrained, but a main phase of extension was proposed to have occurred in the mid-Late Cretaceous, based on low temperature geochronology studies (Lawver and Gahagan, 1994; Fitzgerald, 2002; Spiegel et al., 2016; Veevers, 2012). Based on crustal thickness estimates (An et al., 2015; Llubes et al., 2018; Shen et al., 2018), we reconstruct c. 100 km of extension in the West Antarctic Rift System between 95 and 84 Ma (Fig. 7).

Australia-East Antarctica motion is based on marine magnetic anomalies back to chron C34y (83.7 Ma), although seafloor spreading was slow before chron C17o (~38 Ma) (Cande and Stock, 2004b; Whittaker et al., 2007, 2013). Pre-drift extension between East

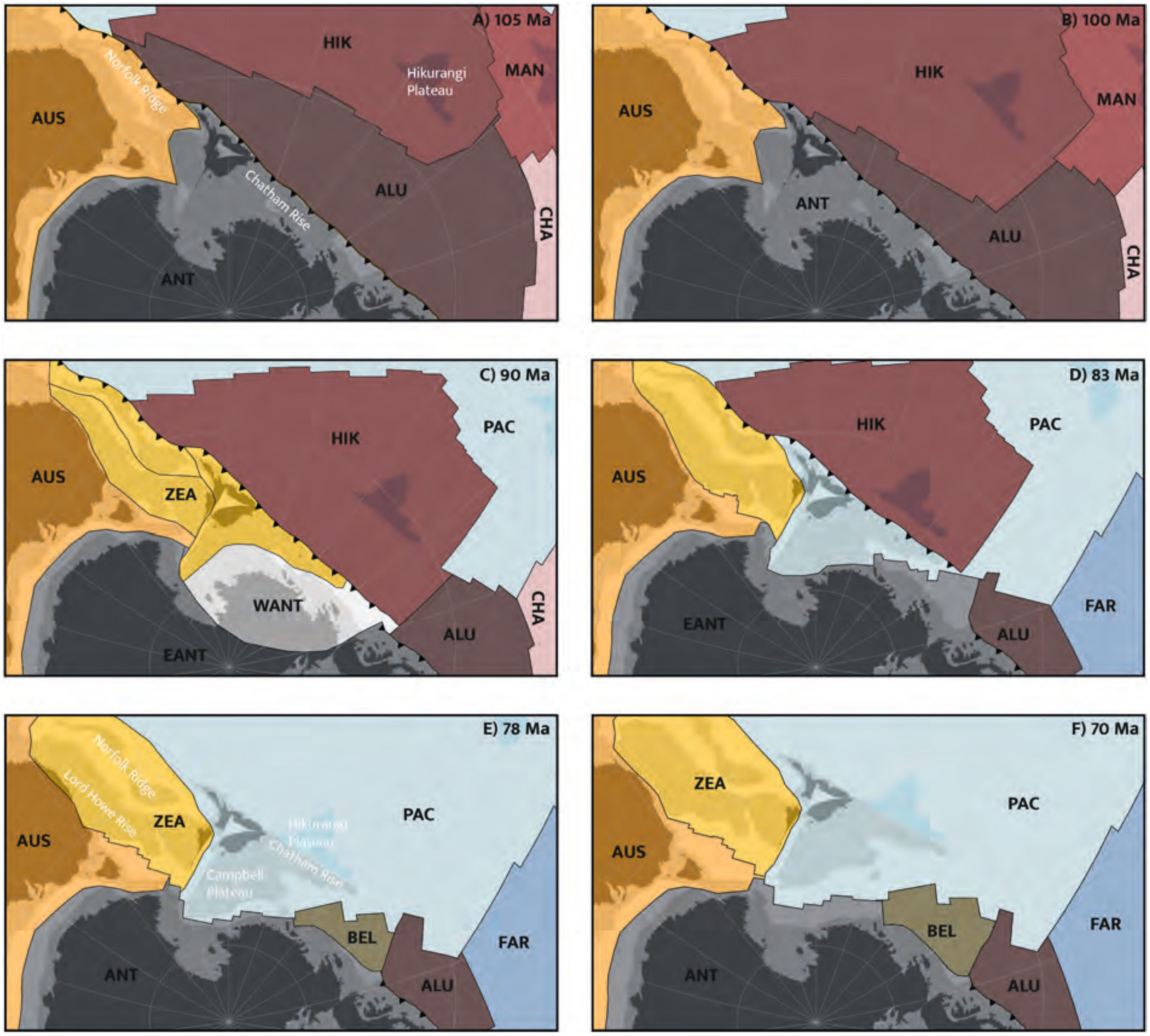

Fig. 7. Detailed snapshots of our kinematic reconstruction in an East Antarctica fixed reference frame, highlighting the transition from the East Gondwana subduction zone margin to a passive margin. Extension within East Gondwana started in multiple locations before subduction of the Hikurangi Plate ended and this plate as well as Zealandia became part of the Pacific plate. Present-day coastlines and outline of continental crust are shaded behind the plate colors and shown for reference. The Ontong Java Nui Large Igneous Provinces are also outlined, where the outline of the Hikurangi Plateau does not include the parts that were subducted below Chatham Rise and the North Island. Plate names: ALU: Aluk Plate; ANT: Antarctic Plate (EANT: East Antarctic Plate; WANT: West Antarctic Plate); AUS: Australian Plate; HIK: Hikurangi Plate; MAN: Manihiki Plate; NAZ: Nazca Plate; PAC: Pacific Plate; ZEA: Zealandia.

Antarctica and Australia started at 136 Ma (Whittaker et al., 2013), and we base the East Gondwana fit of East Antarctica and Australia on the reconstruction of the extended conjugate continental margins of Williams et al. (2011) and Gibbons et al. (2012). Our Australia-East Antarctica reconstruction is similar to that of Müller et al. (2019).

The Late Cretaceous to early Eocene separation of Lord Howe Rise (North Zealandia) from Australia is recorded by marine magnetic anomalies C24-C34y (53.9–83.7 Ma) in the Tasman Sea (Gaina et al., 1998). We use the finite rotation poles of Gaina et al. (1998) in our reconstruction, like Seton et al. (2012) and Müller et al. (2019). Pre-drift extension is thought to have started c. 95 Ma, concurrently with extension in the New Caledonia Basin, between the Norfolk Ridge and Lord Howe Rise (Fig. 6D-E and 7C) (Grobys et al., 2008). The back-arc basins between Lord How Rise and the Tonga-Kermadec-Hikurangi subduction zone are reconstructed as in Van de Lagemaat et al. (2018a), using marine magnetic anomaly constraints from Yan and Kroenke (1993), Sdrolias et al. (2003), and Herzer et al. (2011).

We connect the plate circuit of our reconstruction to Africa by reconstructing East Antarctica-Africa motion through the South Atlantic Ocean. This is based on finite rotation poles based on marine magnetic anomalies back to chron M38 (c. 164 Ma) of DeMets et al. (2021) (C1-C23; 0–51.7 Ma), Cande et al. (2010) (C23-C29; 51.7–64.9 Ma), (Bernard et al., 2005) (C29-C33; 64.9–79.9 Ma), and Mueller and Jokat (2019) (C34y-M38; 84.7–162.9 Ma).

4.2. Pre-C34y plate reconstruction of the Paleo-Pacific realm

4.2.1. Evolution of the Phoenix Plate

Direct kinematic constraints on the evolution of the Phoenix Plate come from marine magnetic anomalies preserved on the Pacific Plate (Nakanishi et al., 1992). The oldest of these anomalies, preserved in the west Pacific Ocean, formed at the Pacific-Phoenix Ridge (Larson and Chase, 1972), and were identified as M29n.2n – M1n (Nakanishi et al., 1992), indicating that Pacific-Phoenix spreading was active from at least 155.9 to 123.8 Ma. We reconstruct the motion of Phoenix for this time interval using GPlates, by mirroring the marine magnetic anomalies that are preserved on the Pacific Plate, assuming symmetric spreading (Fig. 6A-B). We reconstruct Pacific-Phoenix spreading until 120 Ma, the timing of Ontong Java Nui break-up (see section 4.2.2) (Taylor, 2006; Chandler et al., 2012).

While Pacific-Phoenix spreading was active, the Pacific Plate was also spreading with the Farallon and Izanagi plates (or Izanami Plate; see Boschman et al., 2021b). Marine magnetic anomalies that formed during chrons M29 – M0 (156.9–121.4 Ma) were identified on the eastern side of the Pacific triangle (Nakanishi et al., 1992), which constrain spreading between the Pacific and Farallon plates. Pacific-Izanagi spreading is constrained by marine magnetic anomalies that formed during chrons M35 – M5 (160.9–127.5 Ma) (Nakanishi et al., 1992). We reconstruct Farallon-Pacific and Izanagi-Pacific spreading in this time interval based on the marine magnetic anomalies (Nakanishi et al., 1992), using the reconstruction poles of Boschman et al. (2021a).

Marine magnetic anomalies that formed in the southeast corner of the Pacific triangle suggest the formation of two microplates (the Trinidad and Magellan microplates) around the Pacific-Farallon-Phoenix triple junction (Nakanishi and Winterer, 1998). The Trinidad microplate formed around chron M21 (146.6 Ma) and stopped acting as a separate plate around chron M14 (136.9 Ma) (Nakanishi and Winterer, 1998). The Magellan microplate formed around chron M15 (138.5 Ma) and remained active until chron M9 (129.9 Ma), when it merged with the Pacific Plate (Nakanishi and Winterer, 1998). We incorporate the independent motion of the Magellan microplate between chrons M15 and M9 (138.5–129.9 Ma) in the reconstruction. We computed finite rotation poles for this reconstruction in GPlates, based on the magnetic anomaly picks of Nakanishi and Winterer (1998). We do not reconstruct the Trinidad microplate, because there are not enough marine magnetic anomaly identifications for a reliable reconstruction of this microplate.

From reconstruction of Pacific-Farallon and Pacific-Izanagi spreading, it follows that the Phoenix Plate also formed mid-oceanic ridges with the Farallon and Izanagi plates (Fig. 6A). The location of these spreading ridges relative to the Pacific triangle is unknown, but undated marine magnetic anomalies in the Caribbean plate have orientations that are consistent in direction with those that would have formed at the Farallon-Phoenix ridge, and ages of ocean floor exposed in western Costa Rica are consistent with a Jurassic age of spreading of this lithosphere (Boschman et al., 2019). This suggests that prior to the Cretaceous Quiet Zone, the Farallon-Phoenix ridge was located at the longitude of (and subducting below) northern South America. The Izanagi-Phoenix ridge is generally assumed to have remained north of Australia (e.g., Seton et al., 2012). The Phoenix Plate and its Cretaceous to Cenozoic daughters were therefore lost along a continuous subduction margin that spanned from the Caribbean region, down along the westcoast of South America, continuing along the West Antarctic and Zealandia margins to northeast Australia and possibly into Southeast Asia (Fig. 2 and 6A).

4.2.2. Ontong Java Nui break-up

The Phoenix lineations on the Pacific Plate are overlain in the west by the Ontong Java Plateau (Larson, 1997). South of the Phoenix lineations is the oceanic Ellice Basin, which is devoid of marine magnetic anomalies due to its formation during the Cretaceous Normal Superchron, but has east-west trending fracture zones (Benyshek et al., 2019). According to the 'superplateau' hypothesis, the Ontong Java Plateau, together with the Manihiki and Hikurangi plateaus was emplaced as a single Large Igneous Province, known as Ontong Java Nui, around 125–120 Ma (Fig. 6B; Taylor, 2006; Chandler et al., 2012). Shortly after emplacement, Ontong Java Nui broke up into the three modern plateaus through spreading in the Ellice Basin and Osbourn Trough (Taylor, 2006; Chandler et al., 2012; Hochmuth et al., 2015). The Ontong Java Nui LIP erupted on either side of the already existing Pacific-Phoenix spreading ridge: the Ontong Java Plateau represents the part of the LIP that formed on the Pacific Plate, whereas the Manihiki and Hikurangi plateaus formed on the former Phoenix Plate (Larson, 1997; Seton et al., 2012). After separation, the Manihiki and Hikurangi plateaus became part of independent tectonic plates, which grew larger than the original LIPs through the formation of new oceanic crust at their bounding mid-ocean ridges (Fig. 6B-C) (Seton et al., 2012). We refer to these plates as the Manihiki and Hikurangi plates. When we discuss the actual LIPs, we will refer to them as Manihiki and Hikurangi plateaus. Restoration of spreading in the Ellice Basin and Osbourn Trough reconstructs the Hikurangi Plate via the Manihiki Plate relative to the Pacific. The emplacement and subsequent break-up of Ontong Java Nui also resulted in the fragmentation of the Phoenix Plate (e.g., Seton et al., 2012). The spreading history of the Ellice Basin and Osbourn Trough is thus of key importance in the search of the Phoenix Plate and for reconstructing the convergence history between the Pacific realm plates and the Zealandia margin of East Gondwana.

The Ontong Java Nui fit of the three plateaus is based on the interpretation of conjugate rifted margins (Taylor, 2006; Chandler et al., 2012). The general absence of marine magnetic anomalies in the Cretaceous Quiet Zone makes the opening history of these basins challenging to reconstruct in detail. The start of opening of the basins postdated the main formation phase of Ontong Java Nui, which occurred at 125–120 Ma. This age is based on $^{40}Ar/^{39}Ar$ dating of tholeiitic basalts dredged from the three plateaus (Mahoney et al., 1993; Hoernle et al., 2010; Timm et al., 2011) and on the age of sediments directly overlying pillow basalts (Winterer et al., 1974; Sliter et al., 1992). Spreading at the Osbourn Trough started before 115 Ma, based on 115 ± 1 Ma U-Pb zircon ages from dredged lavas and volcaniclastic sandstones from the West Wishbone Ridge (Mortimer et al., 2006). Dating of rift-related structures revealed a c. 120 Ma age of separation between the Hikurangi and Manihiki plateaus (Davy et al., 2008). The onset of rifting in the Ellice Basin between the Manihiki and Ontong Java plateaus is thought to have occurred concurrently with the onset of spreading at the Osbourn Trough, although this is not confirmed by radiometrically dated dredge samples (e.g., Chandler et al., 2012; Hochmuth et al., 2015).

A tectonic reconstruction for the final stages of opening of the Ellice Basin was presented by Benyshek et al. (2019), based on detailed bathymetric data from the center of the basin. They tentatively suggested ages for their rotation poles, based on estimated spreading rates, but these await confirmation by radiometric dating of basement samples (Benyshek et al., 2019). The end of spreading in the Ellice Basin most likely occurred before the end of the CNS, i.e., before 83.7 Ma.

Because no marine magnetic anomalies have been confidently identified, spreading at the Osbourn Trough is also widely interpreted to have occurred entirely during the Cretaceous Normal Superchron (e.g., Chandler et al., 2012). The age of arrest of the Osbourn Trough opening is important for the age of cessation of subduction at the Gondwana margin of New Zealand. The age of 86 Ma incorporated in the widely used global plate models (Seton et al., 2012; Matthews et al., 2016; Müller et al., 2019) came from Worthington et al. (2006), who interpreted the age of arrest of spreading from an age for arrest of subduction based on geological interpretations from New Zealand: occurrence of calc-alkaline volcanism until 89 Ma (Smith and Cole, 1997), the interpreted ongoing outbuilding of an accretionary wedge (Mazengarb and Harris, 1994; Kamp, 1999, 2000) and an 86 Ma episode of metamorphism (Vry et al., 2004), all recognized in New Zealand. But because this young age of subduction arrest is widely disputed by the geological community of New Zealand who prefer a 105–100 Ma (e.g., Bradshaw, 1989; Luyendyk, 1995; Crampton et al., 2019; Mortimer et al., 2019; Gardiner et al., 2021), and it is this debate that we aim to reconcile, our reconstruction of Osbourn Trough should remain independent from the interpretations of the geology of New Zealand. Billen and Stock (2000) tentatively identified anomalies C33 and C32 (79.9 and 73.6 Ma) in the Osbourn Trough. Because the magnetic anomalies are not obvious lineations, they called for more magnetic data and dredge samples. The magnetic anomalies have thus far not been independently confirmed, but Mortimer et al. (2019) reported an 84.4 ± 3.5 Ma $^{40}Ar/^{39}Ar$ age of plagioclase in a basalt flow recovered from bore hole DSDP595, which is located c. 200 km north of the former Osbourn Trough spreading center

(Fig. 4). Through extrapolation of spreading rates, they proposed that Osbourn Trough spreading may have continued until ~79 Ma (Mortimer et al., 2019), implying that spreading may indeed have continued after the Cretaceous Quiet Zone as suggested by Billen and Stock (2000). However, the 84.4 Ma age is a tentative age, as the effects of seawater alteration could not be entirely ruled out (Mortimer et al., 2019). On the other hand, Zhang and Li (2016) suggested that spreading at the Osbourn Trough ceased around 101 Ma, which would require ultrafast spreading rates of 19 cm/yr. This is based on a 103.7 ± 2.3 Ma *Re*—Os isochron age of basalts recovered from bore hole U1365 (Fig. 4), adjacent to bore hole DSDP595, which contradicts the 84.4 Ar/Ar age of Mortimer et al. (2019).

We reconstruct the start of spreading in both basins at 120 Ma (Fig. 6B), following Chandler et al. (2012), similar to Seton et al. (2012) and Müller et al. (2019). For the Osbourn Trough, we use rotation poles of Chandler et al. (2012) to reconstruct the spreading history, but we incorporate the new constraints from Mortimer et al. (2019) of spreading until 79 Ma rather than the contested, New Zealand geology-based 86 Ma estimate of Worthington et al. (2006) that is used in Seton et al. (2012) and Müller et al. (2019). We note that the age for the end of

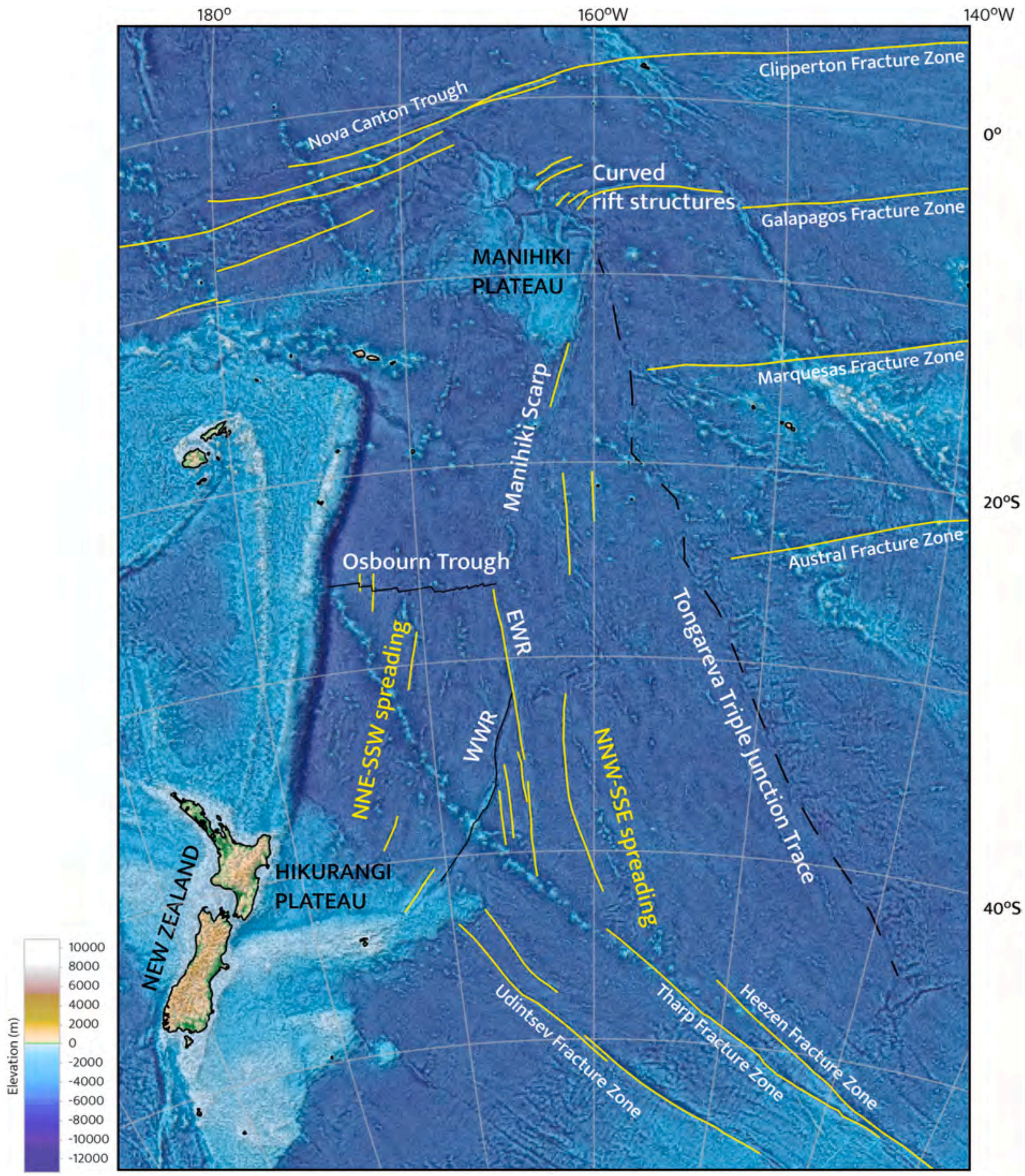

Fig. 8. Bathymetry of the region to the northeast of New Zealand, highlighting features of the seafloor fabric. Digitalized fracture zone data (in yellow) were obtained from the GSFML database (Matthews et al., 2011; Seton et al., 2014; Wessel et al., 2015). Background image is ETOPO1 1 Arc-Minute Global Relief Model (Amante and Eakins, 2009; NOAA National Geophysical Data Center, 2009). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Osbourn Trough spreading may change in the future if more reliable radiometric dating of the Osbourn Basin becomes available, and we will discuss below what difference a different age would make for the estimate for subduction arrest at the New Zealand margin. For the Ellice Basin, we use the Chandler et al. (2012) rotation pole for the Ontong Java-Manihiki fit at 120 Ma and the rotation poles of Benyshek et al. (2019) for subsequent opening, with spreading ending at 90 Ma.

The contemporaneous opening of the Ellice Basin and Osbourn Trough requires that a mid-ocean ridge existed between the Hikurangi and Pacific plates (Fig. 6B-E). The rate and direction of spreading along this ridge follows from the Pacific-Manihiki and Manihiki-Hikurangi reconstructions. This spreading ridge, as well as the Pacific-Manihiki-Hikurangi triple junction was lost to subduction at the Tonga-Kermadec-Hikurangi subduction zone during the Cenozoic (Fig. 6E-J).

4.2.3. Seafloor fabric

To the east of the Manihiki and Hikurangi plates, Seton et al. (2012) identified two more daughter plates of the Phoenix Plate: Chasca and Catequil. We continue using the name Chasca Plate, but the Catequil Plate of Seton et al. (2012) is the same as the Aluk Plate in our reconstruction. We prefer the name Aluk Plate, because it is the established name for the remnant of this plate whose lithosphere remains in the southeast Pacific today. As with Seton et al. (2012), we derive the former existence of the Chasca and Aluk plates from seafloor fabric and marine magnetic anomaly identifications.

The pre-83.7 Ma existence of the Aluk Plate follows from trends in the seafloor fabric east of the Osbourn Trough. The extinct Osbourn Trough spreading center can be followed eastwards until longitude 165°W, where it suddenly stops (Fig. 8). North and south of the Osbourn Trough abyssal hill trends are WNW-ESE for the older part of the basin, and E-W for the youngest part (Downey et al., 2007, see also their Fig. 6). These abyssal hill trends, together with NNE-SSW trending fracture zones constrains the NNE-SSW to N-S spreading direction of the Hikurangi Plate relative to the Manihiki Plate. This trend in seafloor fabric that formed at the Osbourn Trough is delineated by the NNE-SSW trending Manihiki Scarp and the West Wishbone Ridge, clear traces in the ocean floor (Fig. 8). East of the Manihiki Scarp and West Wishbone Ridge, abyssal hills are trending ENE-WSW (Downey et al., 2007, their Fig. 6) and fracture zones are trending NNW-SSE (Fig. 8). This suggests that the oceanic crust here formed at a different spreading center, between different plates (Downey et al., 2007). We suggest here that this part of oceanic crust formed through spreading between the Manihiki and Aluk plates, both daughters of the Phoenix Plate. There is no remnant of an extinct spreading ridge preserved in this part of the Pacific Plate, which suggests that all oceanic crust preserved here formed as part of the Manihiki Plate (e.g., Seton et al., 2012). As Manihiki was incorporated into the Pacific Plate at c. 90 Ma (Benyshek et al., 2019), the Manihiki-Aluk ridge became the Pacific-Aluk ridge at this time. The location of the Pacific-Aluk Ridge is constrained after 83.7 Ma by marine magnetic anomalies preserved on the Pacific Plate (Fig. 4 and 6E; see also section 4.1) (Cande et al., 1995; Larter et al., 2002; Eagles, 2004; Wobbe et al., 2012). The direction of spreading between the Manihiki/Pacific and Aluk plates follows from the NNW-SSE directed fracture zones that are preserved on the Pacific Plate (Fig. 8). The average rate of Manihiki-Aluk spreading follows from the 120 Ma break-up configuration of the Phoenix Plate into these plates and the chron C34y (83.7 Ma) location of the Aluk-Pacific ridge, which is constrained by marine magnetic anomalies on the Pacific Plate (Larter et al., 2002; Eagles, 2004). We reconstruct a constant spreading rate in this 120–83.7 Ma period.

The nature of the plate boundary between the Aluk and Hikurangi plates follows from the reconstruction of the Hikurangi and Aluk plates relative to the Manihiki Plate. In our reconstruction, Aluk-Manihiki spreading occurred at a higher rate than Hikurangi-Manihiki spreading (~8.5 cm/yr and ~ 4.5 cm/yr half-spreading rate, respectively) (Fig. 6B-D). As a result, between 120 and 110 Ma, the plate boundary between the Aluk and Hikurangi plates was a right-lateral transform fault northeast of the Hikurangi Plateau, forming the West Wishbone Ridge. After 110 Ma, some extension occurred between the Aluk and Hikurangi plates east of the Hikurangi Plateau, accommodated by a mid-ocean ridge. South of the Hikurangi Plateau, the plate boundary between the Hikurangi and Aluk plates was a mid-ocean ridge from 120 Ma until its subduction below the Zealandia margin around 100–90 Ma (Fig. 6B-C and 7).

From the northeast corner of the Manihiki Plateau towards the south, there is a clear trace in the seafloor fabric (Fig. 8). This feature has been identified as a trace of a former triple junction (Larson and Chase, 1972), and was named the Tongareva triple junction (Larson et al., 2002). It was previously suggested that the Tongareva triple junction formed the junction between the Pacific, Farallon, and Phoenix plates (e.g., Larson et al., 2002; Viso et al., 2005; Hochmuth and Gohl, 2017). We instead infer that the Tongareva triple junction formed the junction between Manihiki, Chasca and Aluk plates, until c. 90 Ma, when Manihiki merged with the Pacific Plate. Between 90 and 83.7 Ma, the Tongareva triple junction was the junction of the Pacific, Chasca and Aluk plates, after which it became the triple junction between Pacific, Farallon and Aluk plates when Chasca was captured by Farallon (Fig. 6B-F).

The existence of the Chasca Plate follows from rift structures on the northeast margin of the Manihiki Plateau (Fig. 8) (Larson et al., 2002; Viso et al., 2005). It was previously suggested that this fragment was incorporated into the Farallon Plate at 110 Ma (Hochmuth and Gohl, 2017). The location of the Farallon Plate relative to the Pacific Plate is constrained by marine magnetic anomalies of chrons C34y (83.7) and M0 (121.4). Attaching a fragment of oceanic crust that formed east of the Manihiki Plateau to the Farallon Plate at 110 Ma, however, leads to convergence along the southeast margin of the Manihiki Plateau. This convergence is contradicted by the existence of the Tongareva triple junction trace, as described above. Instead, we reconstruct independent motion of the Chasca Plate until 83.7 Ma. The capture of the Chasca Plate by the Farallon Plate, which resulted from the inactivation of the transform fault (Clipperton Fracture Zone) that separated the Chasca and Farallon plates, may have occurred a few millions of years earlier. This would require higher Chasca – Manihiki/Pacific spreading rates, but these are unknown. We therefore choose to reconstruct the capture at the time of C34y (83.7 Ma), as this marine magnetic anomaly provides the first positive evidence that the Chasca plate was captured. Seton et al. (2012) and Chandler et al. (2012) incorporate the Chasca Plate into the Farallon Plate a few million years earlier at 86 Ma, contemporaneous with the cessation of Osbourn Trough spreading in their model (see section 4.2.2).

We reconstruct the start of Chasca – Manihiki motion at 120 Ma, the same time as the onset of spreading between the other daughters of the Phoenix Plate (Chandler et al., 2012). Rotation poles of the Chasca Plate relative to the Manihiki Plate are calculated in GPlates. In our reconstruction, we ensure that early motion of the Chasca Plate follows the trend of the curved rift structures at the NE Manihiki margin (Fig. 8). In addition, we assume that the Pacific-Farallon ridge at 83.7 Ma formed at the location of the Pacific-Chasca ridge, after Manihiki was captured by the Pacific Plate at 90 Ma (Benyshek et al., 2019) and Chasca was captured by Farallon.

The rotation poles for the reconstruction of the Ellice Basin include a rotation of the Manihiki Plate relative to the Pacific Plate between 102 and 98 Ma, based on a change in fracture zone orientation in the Ellice Basin from ~*E*-W to WNW-ESE (Taylor, 2006; Chandler et al., 2012; Benyshek et al., 2019). As the Chasca Plate is reconstructed relative to the Manihiki Plate in our plate circuit, the rotation modeled by Benyshek et al. (2019) also results in a rotation of the Chasca Plate. This in turn leads to convergence at the Chasca-Farallon plate boundary that at that time was still located at the latitude of northern South America. This rotation of the Chasca Plate coincides with the estimated timing of subduction initiation at the western Caribbean plate boundary of modern Central America (Whattam and Stern, 2015; Boschman et al., 2019), which at 100 Ma was still located far west within the eastern Panthalassa realm (e.g., Pindell and Kennan, 2009; Boschman et al., 2014). Rotation of the Manihiki Plate may thus have resulted in subduction initiation at the future western Caribbean plate boundary.

5. Discussion

The kinematic constraints reviewed in section 4 lead to a plate kinematic evolution from 150 Ma onward as portrayed in Fig. 6, and in snaphots highlighting the final stages of subduction in Fig. 7. We provide GPlates reconstruction files and an animation of the reconstruction in the supplementary information. Below, we discuss uncertainties in our reconstruction, offer interpretations of possible dynamic drivers of plate reorganizations, and evaluate until when convergence along the Gondwana margin must have continued. Finally, we discuss how differences between plate kinematics and geology-based interpretations may be reconciled, and what opportunities our reconstruction provides for future geological research.

5.1. Dating the end of convergence across the Gondwana margins

During the Paleozoic and Mesozoic, the vast Phoenix Plate occupied large parts of the south Panthalassa Ocean. After the birth of the Pacific Plate around 190 Ma (Seton et al., 2012; Boschman and Van Hinsbergen, 2016), the Phoenix Plate formed spreading ridges with the Pacific, Izanagi/Izanami and Farallon plates. Subduction at the Gondwana margin of South America, Antarctica, and Australia/Zealandia is not controversial, although more plates may have been involved between the Phoenix Plate and the Gondwana margin (e.g., Boschman et al., 2021a). Our reconstruction from 150 Ma until the 125–120 Ma emplacement of the Ontong Java Nui LIP, placed in the Pacific hotspot reference frame of Torsvik et al. (2019) and the Indo-Atlantic slab-fitted frame of Van Der Meer et al. (2010) straightforwardly shows convergence of Phoenix in the west, south, and east, consistent with geological records from South America, Antarctica, and Zealandia (Mortimer et al., 2014; Burton-Johnson and Riley, 2015; Pepper et al., 2016; Jordan et al., 2020; Maurizot et al., 2020b). Along the eastern margin of the southern Pacific, convergence and subduction continue today (Fig. 1). Conversely, convergence ceased along the southern and western margins in the Late Cretaceous, which was followed by re-initiation of subduction in the west during the Cenozoic (e.g., Seton et al., 2012; Van de Lagemaat et al., 2018a). In this section, we establish until when, according to plate kinematic constraints, subduction continued. In addition, we examine whether the choice of mantle reference frame is of influence on this age estimation.

It is well agreed upon that a subduction zone was present along the entire East Gondwana margin, from the Antarctic Peninsula to New Caledonia, until 105 Ma (Bradshaw, 1989; Luyendyk, 1995; Maurizot

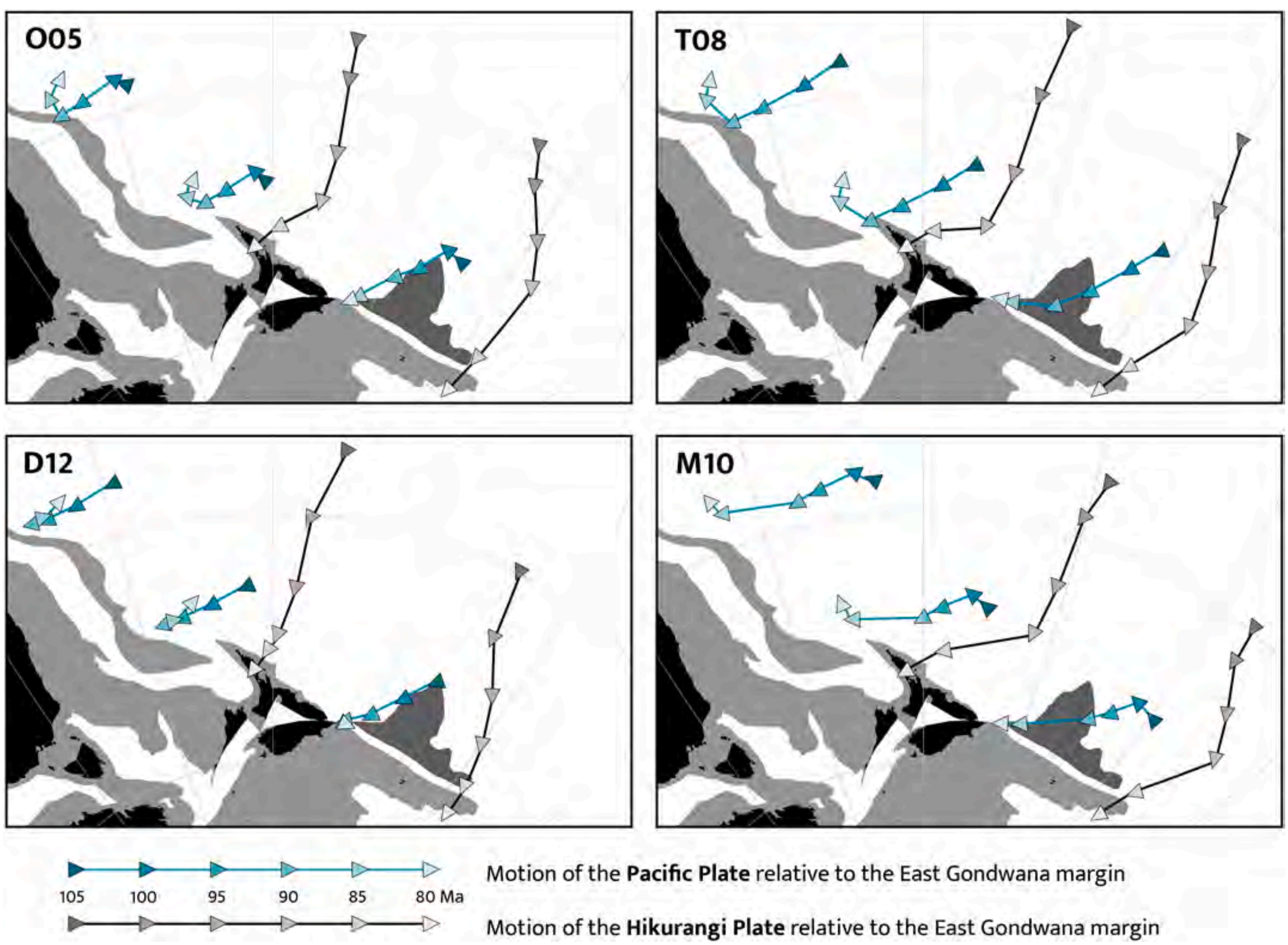

Fig. 9. 80 Ma reconstruction in an East Antarctica fixed reference frame showing the motion of the Pacific Plate and Hikurangi Plate relative to the East Gondwana margin in different mantle reference frames (O05: O'Neill et al., 2005; T08: Torsvik et al., 2008; D12: Doubrovine et al., 2012; M10: Van Der Meer et al., 2010). This figure shows that convergence between the Pacific oceanic plates and the East Gondwana margin continued until at least 90 Ma in all reference frames, and until 79 Ma if Osbourn Trough spreading was still active.

et al., 2020b; Gardiner et al., 2021). In addition, our plate reconstruction shows that convergence at the Zealandia and Antarctic margins continued, until at least 90 Ma and possibly until 85 Ma (Fig. 9). This is well beyond 105–100 Ma, when some models that are based on onshore geology (e.g., the onset of continental extension and the change in geochemistry of magmatism) argue for the cessation of subduction along the Zealandia sector of East Gondwana (Bradshaw, 1989; Davy et al., 2008; Crampton et al., 2019; Mortimer et al., 2019). In our updated plate kinematic model, the timing of the end of convergence is dependent on two variables: the reconstruction of Osbourn Trough spreading and the choice of Indo-Atlantic mantle reference frame for East Gondwana. We only use the Pacific mantle reference frame of Torsvik et al. (2019), because they showed that previous implementations of Pacific reference frames are flawed. Furthermore, the correctly implemented Pacific hotspot reference frame of Wessel and Kroenke (2008) only leads to more convergence across the Gondwana margin than the model of Torsvik et al. (2019). In all Indo-Atlantic mantle reference frames, convergence continues until the cessation of Osbourn Trough spreading; that is, until 79 Ma in our reconstruction following Mortimer et al. (2019). This reconstruction of the Osbourn Trough leads to c. 1500–2000 km of convergence between the Hikurangi Plate and the Gondwana margin between 100 and 79 Ma, depending on the reference frame (Fig. 9). If future radiometric dating of dredge samples would suggest an older age for the end of Osbourn Trough spreading, the convergence between the Hikurangi Plate and the East Gondwana margin would simply be accommodated by higher rates of spreading and subduction between 120 Ma and any new and reliable date suggested. Even if Osbourn Trough spreading had already ceased by 101 Ma, as interpreted by Zhang and Li (2016), there would still have been 800–1100 km of post-100 Ma convergence between the Pacific Plate and the Zealandia margin (Fig. 9). In this scenario, convergence at the Zealandia margin continued until c. 90 Ma, applying the reference frames of Torsvik et al. (2008) or Doubrovine et al. (2012), or continued until c. 84 Ma (when the Campbell plateau became incorporated in the Pacific Plate) in applying the slab frame of Van Der Meer et al. (2010) or the hotspot reference frame of O'Neill et al. (2005).

The Hikurangi-Pacific ridge formed a triple junction with the subduction zone located along the margin of East Gondwana, in the vicinity of the Norfolk Ridge (Figs. 6 and 7). North of this Hikurangi-Pacific-Gondwana triple junction, the rate and amount of convergence at the East Gondwana margin were not influenced by spreading at the Osbourn Trough, as the Pacific Plate directly subducted below the Norfolk Ridge. The precise location of this triple junction, where the Pacific-Hikurangi ridge subducted below eastern Gondwana, is uncertain, as it has subsequently been consumed at the Cenozoic Tonga-Kermadec and New Caledonia trenches. In our reconstruction, after 95 Ma we place this triple junction just south of New Caledonia (Figs. 6 and 7), which results from the assumption of symmetric spreading between the Pacific and Hikurangi plates between 120 and 79 Ma (see section 4.2.2). The relative motion between the Pacific Plate and the Norfolk Ridge north of the Pacific-Hikurangi-Gondwana triple junction is convergent in all reference frames until at least 90 Ma. In the hotspot reference frames of O'Neill et al. (2005), Torsvik et al. (2008), and Doubrovine et al. (2012), convergence north of the Hikurangi-Pacific-Gondwana triple junction ends at 90 Ma (Fig. 9). In the slab-fitted mantle reference frame of Van Der Meer et al. (2010), convergence between the Pacific Plate and the Norfolk Ridge continues until 85 Ma. Subduction south of the New Caledonia sector of the East Gondwana margin thus continued until 90–85 Ma (Fig. 9).

In an attempt to reconcile geological (on-land) interpretations of cessation of subduction in New Zealand with oceanic plate reconstructions, Mortimer et al. (2019) proposed a solution to avoid convergence beyond 100 Ma at the Zealandia margin. In this scenario, the Hikurangi Plateau arrives in the trench at 100 Ma, and the Manihiki and Ontong Java plateaus move northwards relative to the margin between 100 and 79 Ma. However, this model places the Pacific plate mosaic ~2250 km farther to the South at 100 Ma than suggested by the hotspot frame of Torsvik et al. (2019) (Fig. 10), which is well beyond the 3° uncertainty assigned to the hotspot model. The solution of Mortimer et al. (2019) therefore does not work in our kinematic reconstruction that combines relative and absolute plate motions; it would require an absolute hotspot wander between 100 and 90–85 Ma of 10–20 cm/yr, for all hotspots below the Pacific Plate, for which there is no evidence, and which is two orders of magnitude faster than typical hotspot motions (e.g., Doubrovine et al., 2012). In addition, we tested whether the latest and highest-detail published isochron sets from the South Pacific realm change the age for the end of convergence across the Gondwana margin that followed from widely used global plate models (e.g., Seton et al., 2012; Müller et al., 2019). And while our updated model differs in detail, for instance in the reconstruction of plates and plate motions in lithosphere that was lost to subduction, the conclusions from those global models are robust: Plate kinematic models leave no room for a cessation of subduction along the Zealandia margin at 100 Ma or before; instead, convergence between the Phoenix Plate's daughters and the East Gondwana margin must have continued until at least 90–85 Ma. Below the New Zealand margin, convergence likely continued even longer if spreading in the Osbourn Trough continued beyond 85 Ma (e.g., 79 Ma according to Mortimer et al., 2019).

5.2. Reconciling ongoing convergence after 100 Ma with the geology of Zealandia

Our plate kinematic reconstruction requires that convergence, and by inference subduction, continued until at least 90 Ma along the entire East Gondwana margin, and possibly until 79 Ma below New Zealand and Chatham Rise. While our reconstruction is easily reconciled with the geology of New Caledonia, where subduction-related accretion and magmatism continued until c. 90 Ma (Cluzel et al., 2010; Maurizot et al., 2020b), it conflicts with the common interpretation based on geological observations from New Zealand that subduction there ceased at 105–100 Ma. The observations from New Zealand thus require an alternative explanation.

The first often-cited argument for subduction cessation at 105–100 Ma is the timing of the onset of extension that is recognized in the geology of New Zealand (Bradshaw, 1989; Tulloch and Kimbrough, 1989; Field and Uruski, 1997; Laird and Bradshaw, 2004; Crampton et al., 2019), for example in the Canterbury Basin (Barrier et al., 2020). However, extension in the upper plate above an active subduction zone is common, as evidenced by many intra- and back-arc basins across the world. In fact, extension is presently occurring within the Taupo Volcanic Zone in North Island New Zealand, above the Hikurangi subduction zone (e.g., Villamor and Berryman, 2001). In addition, neither numerical models (e.g., Van Hunen and Allen, 2011; Duretz et al., 2014) nor geological observations (Wortel and Spakman, 2000; Webb et al., 2017; Qayyum et al., 2022) suggest a systematic relationship between slab break-off and upper plate extension. It is even questionable whether the onset of extension in East Gondwana, which is recorded from the West Antarctic Rift System to the Tasman Sea region (Gaina et al., 1998; Behrendt, 1999; Fitzgerald, 2002; Raza et al., 2009; Cluzel et al., 2012; Spiegel et al., 2016; Jordan et al., 2020), is directly governed by subduction termination, or related to the intra-continental forces that governed Gondwana breakup. In any case, extension in the Gondwana margin does not necessitate slab break-off and does not exclude ongoing subduction.

A cessation of subduction around 105-100 Ma is also interpreted from the unconformity between deformed accretionary prism rocks and overlying sedimentary sequences that are then interpreted as 'passive margin' sediments (e.g., Field and Uruski, 1997; Adams et al., 2013; Crampton et al., 2019; Gardiner et al., 2021, 2022) However, this transition from accretion to undisturbed sedimentation is only evidence for the cessation of subduction accretion locally, whereby the age of the overlying undeformed rocks dates the accretion of the rocks below the

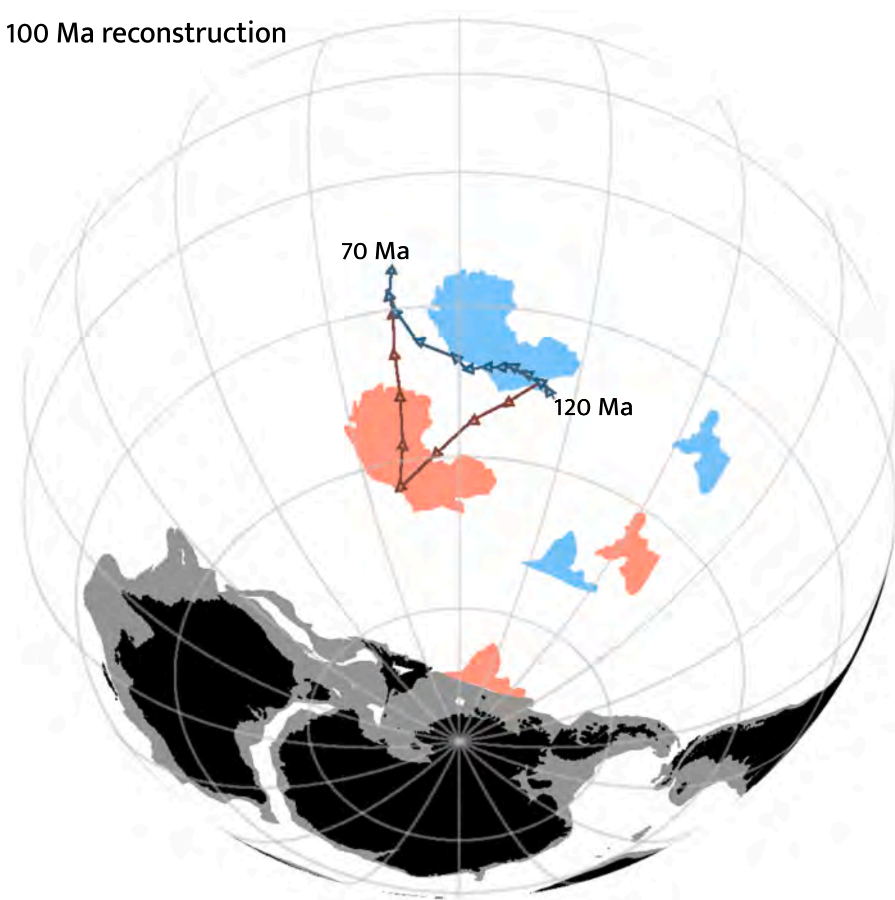

Fig. 10. 100 Ma reconstruction that shows the difference in the location of the Pacific Plate between the hotspot reference frame of Torsvik et al. (2019) and the model of Mortimer et al. (2019) in which the Hikurangi Plateau arrives in the East Gondwana trench at 100 Ma. The difference is shown by the location of the LIPs, of which the Ontong Java Plateau has always been part of the Pacific Plate. In blue is the location of the LIPs as constrained by Torsvik et al. (2019), and in pink is the location of these plateaus in the model of Mortimer et al. (2019). Also shown are the 120–70 Ma motion paths of the Pacific Plate that result from the two models. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

unconformity, but it does not exclude ongoing subduction (Fig. 11; e.g., Mazengarb and Harris, 1994; Kamp 1999, 2000). The accretionary system in New Zealand is younging from west to east; while the youngest part of the accretionary system that is subaerially exposed in New Zealand may be 105-100 Ma, younger parts of the former subduction are located offshore, closer to the paleo-trench. This is also illustrated by the geology of New Caledonia, where the minimum age of the oldest 'cover' sediments is Cenomanian (~100-94 Ma), while subduction is considered to have continued until c. 90 Ma (Maurizot et al., 2020b). Kamp (1999, 2000) also suggested that accretionary wedge accumulation occurred contemporaneously with the deposition of cover sequences in New Zealand, based on interpretations from apatite and zircon fission track dating of Torlesse accretionary wedge rocks. Furthermore, the overlying sequences are clastic sediments derived from the New Zealand orogen, and these sediments may represent forearc basin sediments deposited during ongoing subduction rather than passive margin sequences. For example, Mazengarb and Harris (1994) previously suggested that the formation of the accretionary wedge in New Zealand continued until 84 Ma, based on the interpretation that sedimentation in eastern North Island occurred within and in front of an active thrust system. Similarly, Gardiner and Hall (2021) suggested that sedimentary successions on the northern South Island were deposited in a trench-slope basin during ongoing subduction rather than in a passive margin setting. These interpretations are supported by apatite and zircon fission track thermochronology, which suggest that uplift and erosion migrated eastward until c. 85 Ma, resulting from the outbuilding of the accretionary wedge Kamp (1999, 2000). Nevertheless, even if the oldest accretion in New Zealand is truly 105-100 Ma, the geology of New Zealand itself illustrates the conclusion of (Isozaki et al., 1990, 2010) that accretion of OPS is episodic, intervened by long periods of wholesale subduction or even subduction erosion, which are the default modes of oceanic subduction (e.g., Van Hinsbergen and Schouten, 2021). Hence, the youngest accretion in New Zealand provides a maximum age for the end of subduction but does not preclude a younger age.

The geological interpretation of the cessation of subduction around 105–100 Ma is further inferred from interpretation of the geodynamic setting that caused a change in geochemical signature of magmatism in New Zealand. Although the youngest age of 'normal' subduction-related I-type magmatism in New Zealand was dated as 128 Ma (Tulloch and Kimbrough, 2003), the 131–105 Ma adakitic magmatism is also considered to be related to ongoing subduction (Tulloch et al., 2009). The subsequent onset of A-type magmatism around 100 Ma is widely regarded as signaling the end of subduction (Tulloch et al., 2009). However, the increase in A-type magmatism is interpreted as the result of thinning of the continental crust of Zealandia during extension, which caused less crustal contamination of the igneous rocks (Tulloch et al., 2009), indicating that these interpretations were made under the assumption that subduction ended around 100 Ma, and no alternative causes were explored.

While A-type magmatism is generally interpreted as occurring in the absence of subduction (Loiselle and Wones, 1979), such magmas have also been found in active margin settings, for example related to the arrival of a spreading ridge and the influx of sub-slab mantle to the former wedge (e.g., Zhao et al., 2008; Karsli et al., 2012; Li et al., 2012). We here suggest that the transition to A-type magmatism in New Zealand may also be explained by arrival of a spreading ridge. As explained in section 4.2.3, the plate boundary between the Hikurangi and Aluk plates was likely a spreading ridge south of the Hikurangi Plateau. Our reconstruction predicts that this spreading ridge subducted around 100 Ma below New Zealand (Figs. 7 and 11). The progressive arrival of successively younger oceanic crust before arrival of the spreading ridge may then explain the 128–105 Ma adakitic magmatism, which is often

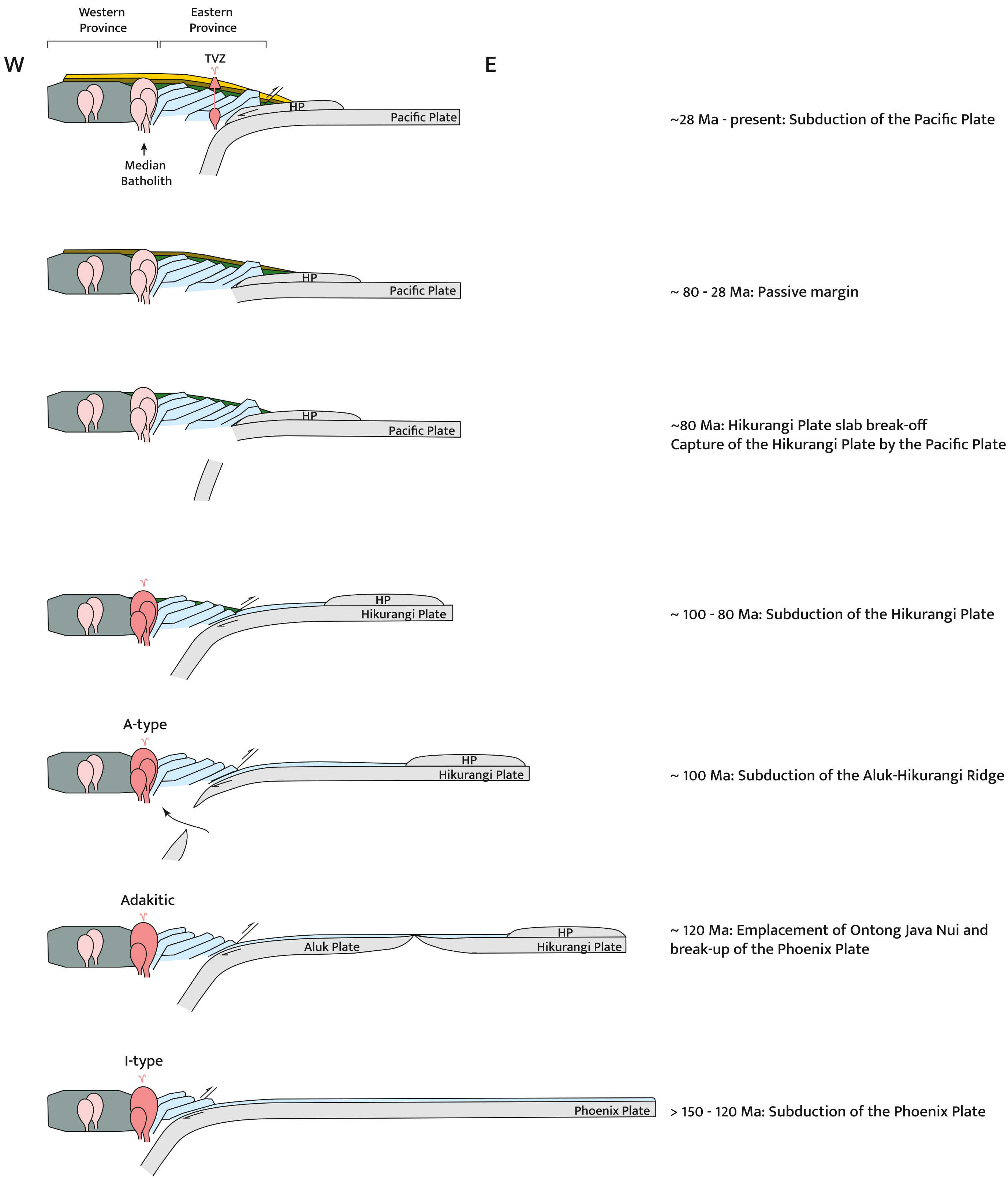

Fig. 11. Schematic cross-sections along the Zealandia margin to highlight the 150 Ma to present tectonic history between the continental margin and the Pacific domain. The 28 Ma-present day cross-section is across North Island, New Zealand, where the Hikurangi Plateau is presently subducting, whereas older cross-sections are across the Chatham Rise where the Hikurangi Plateau entered the trench in the Cretaceous. TVZ refers to the active volcanism of the Taupo Volcanic Zone, which lies between the Waipapa and Torlesse terranes.

related to the subduction of young oceanic crust (Tulloch and Rabone, 1993).

In summary, geological and geochemical interpretations made for New Zealand do not require that subduction ended during c. 105–100 Ma (Figs. 7 and 11). Alternative structural and stratigraphic arguments for the forearc region of New Zealand (Mazengarb and Harris, 1994; Kamp, 1999, 2000; Gardiner and Hall, 2021) are straightforwardly reconciled with ongoing subduction, along with geochemical arguments for the composition of magmatic rocks, which do not exclude ongoing subduction.

5.3. Causes for the end of subduction

Why subduction stopped at the margin of East Gondwana in the Cretaceous is puzzling. Explanations for this cessation have so far mostly focused on regional geological features, such as the arrival of a mid-ocean ridge in the trench (Luyendyk, 1995; Bradshaw, 1989; Matthews et al., 2012). The arrival of a mid-ocean ridge in a subduction zone may indeed change the nature of a plate boundary and trigger slab break-off. The nature of the plate boundary that follows upon ridge arrival commonly depends on the relative motion between the original overriding plate and the plate that was formerly spreading with the original down-going plate. For example, west of the active trench below the Antarctic Peninsula, marine magnetic anomalies young from the ocean towards West Antarctica. This shows that subduction below the Antarctic Peninsula indeed ceased due to the arrival of the Aluk-West Antarctica ridge at the trench below West Antarctica, after which relative motion ceased (Eagles, 2004). However, the Hikurangi-Pacific ridge did not subduct parallel to the trench but subducted at an angle to it (Fig. 6C-E and 7A-D). Moreover, until the ~84 Ma change in absolute plate motion of the Pacific Plate, the whole Panthalassa mosaic was converging with the East Gondwana margin, which means that subduction continued after the arrival of the Hikurangi-Aluk spreading ridge. More importantly, the newly formed Pacific-Antarctic Ridge did not replace the former subduction zone but cut through the suture and formed at a completely different location (Fig. 6D-E and 7C-D). Ridge arrival is thus not a likely candidate to explain the end of subduction.

A second hypothesis for the end of subduction below the Zealandia margin of East Gondwana is the arrival of the Hikurangi Plateau in the trench (e.g., Billen and Stock, 2000; Davy et al., 2008, Davy, 2014; Reyners et al., 2017; Mortimer et al., 2019). In this hypothesis, the plateau chocked the subduction zone after about 150 km of subduction (Riefstahl et al., 2020). However, the Hikurangi Plateau only represents a small portion of the Pacific Plate and only a short length of the trench. If a transform fault could be demonstrated to have bounded the western side of the Hikurangi Plateau, subduction could have continued below the North Island and New Caledonia sections. Moreover, while plateau arrivals at intra-oceanic trenches may cause a polarity reversal (and ongoing subduction), e.g., during the arrival of the Ontong Java Plateau at the Vitiaz trench triggering the formation of the New Hebrides trench and the South Solomon trench (Auzende et al., 1995; Petterson et al., 1997; Quarles van Ufford and Cloos, 2005; Knesel et al., 2008; Lallemand and Arcay, 2021), there is no record of LIP arrival at a trench causing subduction cessation or a plate reorganization on the scale as observed here. Instead, LIP subduction is physically straightforward, even though it may cause shallow dipping slabs (e.g., Yang et al., 2020; Liu et al., 2021). LIP subduction has, for example, been ongoing in the Maracaïbo trench of the southern Caribbean region for more than 50 Ma (White et al., 1999; Boschman et al., 2014), and even the Hikurangi Plateau itself is subducting today at the Hikurangi trench (Collot and Davy, 1998; Reyners et al., 2011, 2017; Timm et al., 2014; Fig. 1), where subduction initiated in the Oligocene (Furlong and Kamp, 2009; Van de Lagemaat et al., 2022). Therefore, while the preservation of the Hikurangi Plateau at the Gondwana margin may suggest that it played a role in determining where the slab broke, it is an unlikely trigger for the cessation of subduction along the entire East Gondwana margin.

Instead, we consider it most likely that the end of subduction in the Zealandia sector of East Gondwana was governed by a change in relative plate motion between the Pacific Plate and East Gondwana (Rey and Müller, 2010). More analysis of the driving forces of the Pacific Plate and the Pacific plate mosaic as a whole, not only of local features on the southernmost Pacific Plate could usefully be undertaken. In the East Asia region, below South China, we note that subduction along the continental margin suddenly stopped at around 90–80 Ma (e.g., Cui et al., 2021). Also, in the North Pacific realm there were prominent changes in plate boundary configuration around 90–85 Ma, including the formation of the Kula Plate (Engebretson et al., 1985; Wright et al., 2016), and initiation of intra-oceanic subduction below the Olyutorsky and Kronotsky arcs (Konstantinovskaia, 2001; Shapiro and Solov'ev, 2009; Domeier et al., 2017; Vaes et al., 2019). An analysis of the causes of plate motion change that formed the prelude to the end of subduction below eastern Gondwana requires a detailed kinematic restoration of the plate boundary reorganization, particularly in the enigmatic transition between the Panthalassa and Tethyan domains of SE Asia, which is beyond the scope of this paper.

6. Conclusions

We have developed a kinematic reconstruction of the South Pacific and East Gondwana realms back to the Late Jurassic (150 Ma). Our aim was to reconstruct the evolution and destruction of the Phoenix Plate, and to reconcile the geological record of New Zealand with the end of Mesozoic subduction along the East Gondwana margin. From our reconstruction we conclude the following:

1) Resulting from the emplacement of Ontong Java Nui around 125–120 Ma, the Phoenix Plate broke into at least four plates: The Manihiki, Hikurangi, Chasca, and Aluk plates. During the Late Cretaceous, the Manihiki and Hikurangi plates were captured by the Pacific Plate, while Chasca was captured by the Farallon Plate. Only the Aluk Plate remained an independent tectonic plate into the Cenozoic.
2) Convergence occurred along the East Gondwana margin until 90 or 85 Ma, depending on choice of mantle reference frame. This convergence occurred independent from spreading at the Osbourn Trough and required the presence of a subduction zone along the entire Zealandia margin until at least 90 Ma and possibly until 85 Ma.
3) Subduction in the New Caledonia region ceased at c. 90 to 85 Ma, but convergence of the Hikurangi Plate with the Chatham Rise must have continued until the cessation of spreading at the Osbourn Trough, recently tentatively estimated at 79 Ma.
4) The cessation of subduction of the Hikurangi Plate along the entire East Gondwana margin was probably a result of a change in Pacific-Gondwana relative plate motion. This was not due to the arrival of the small Hikurangi Plateau compared with the East Gondwana subduction system.
5) The 105–100 Ma structural and magmatic changes within the crust of the overriding New Zealand continental plate may have resulted from subduction of the Aluk-Hikurangi ridge, rather than from the cessation of subduction at the East Gondwana margin.
6) Geological expressions in the overriding plate may be misleading when used to interpret subduction zone dynamics. While a geological record in the overriding plate may provide evidence for the presence of subduction, absence of such evidence should not be interpreted as conclusive evidence for the absence of subduction.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Data availability

No data was used for the research described in the article.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.earscirev.2022.104276.

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