TECTONICS, VOL. 29, TC2004, doi:10.1029/2009TC002476, 2010

Lithosphere delamination with foundering of lower crust and mantle caused permanent subsidence of New Caledonia Trough and transient uplift of Lord Howe Rise during Eocene and Oligocene initiation of Tonga-Kermadec subduction, western Pacific

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Received 19 February 2009; revised 30 July 2009; accepted 18 November 2009; published 20 March 2010.

[1] We use seismic reflection and rock sample data to propose that the first-order physiography of New Caledonia Trough and Norfolk Ridge formed in Eocene and Oligocene time and was associated with the onset of subduction and back-arc spreading at the Australia-Pacific plate boundary. Our tectonic model involves an initial Cretaceous rift that is strongly modified by Cenozoic subduction initiation. Hence, we are able to explain (1) complex sedimentary basins of inferred Mesozoic age; (2) a prominent unconformity and onlap surface of middle Eocene to early Miocene age at the base of flat-lying sediments beneath the axis of New Caledonia Trough; (3) gently dipping, variable thickness, and locally deformed Late Cretaceous strata along the margins of the trough; (4) platform morphology and unconformities on either side of the trough that indicate a phase of late Eocene to early Miocene uplift to near sea level, followed by rapid Oligocene and Miocene subsidence of ~1100–1800 m; and (5) seismic reflection facies tied to boreholes that suggest absolute tectonic subsidence at the southern end of New Caledonia Trough by 1800–2200 m since Eocene time. The Cenozoic part of the model involves delamination and subduction initiation followed by rapid foundering and rollback of the slab. This created a deep (>2 km) enclosed oceanic trough, ~2000 km long and 200–300 km across, in Eocene and Oligocene time as the lower crust detached, with simultaneous uplift and local land development along basin flanks. Disruption of Late Cretaceous and Paleogene strata was minimal during this Cenozoic phase and involved only subtle tilting and local reverse faulting or folding. Basin formation was possible through the action of at least one detachment fault that allowed the lower crust to either be subducted into the mantle or exhumed eastward into Norfolk Basin. We suggest that delamination of the lithosphere, with possible mixing of the lower crust back into the mantle, is more widespread than previously thought and may be commonly associated with subduction initiation, such as Cenozoic events in the Mediterranean and western Pacific.

Citation: Sutherland, R., et al. (2010), Lithosphere delamination with foundering of lower crust and mantle caused permanent subsidence of New Caledonia Trough and transient uplift of Lord Howe Rise during Eocene and Oligocene initiation of Tonga-Kermadec subduction, western Pacific, *Tectonics*, *29*, TC2004, doi:10.1029/2009TC002476.

1. Introduction

[2] The New Caledonia Trough and Lord Howe Rise are located in the southwest Pacific between New Zealand, Australia, and New Caledonia (Figure 1). The New Caledonia Trough is a NNW trending physiographic depression lying in 2000–3500 m water depth between the Lord Howe Rise (500–2000 m water depth) and Norfolk Ridge (0–2000 m water depth). We draw a clear distinction in this paper between the modern well-defined physiography of the New Caledonia Trough (Figure 1) and the sedimentary basins beneath its axis and flanks, which have previously been referred to as the, e.g., New Caledonia Basin, Deepwater Taranaki Basin, Aotea Basin, Fairway Basin [*Collot et al.*, 2008, 2009; *Uruski and Wood*, 1991; *Uruski*, 2008]. The details of the sedimentary record are deduced from core and dredge samples and geophysical data, many of which have not previously been published.

[3] The 2000 km long region is of interest because its strata and structures represent one of the largest unexplored systems of sedimentary basins left on Earth, and they collectively contain a unique record of Mesozoic and Cenozoic past environments and tectonic movements in the southwest Pacific. The history of topographic evolution that we aim to recover has implications for the tectonics and geodynamics of Tonga-Kermadec subduction initiation, past oceanographic connections in the western Pacific, and the dispersal of fauna and flora.

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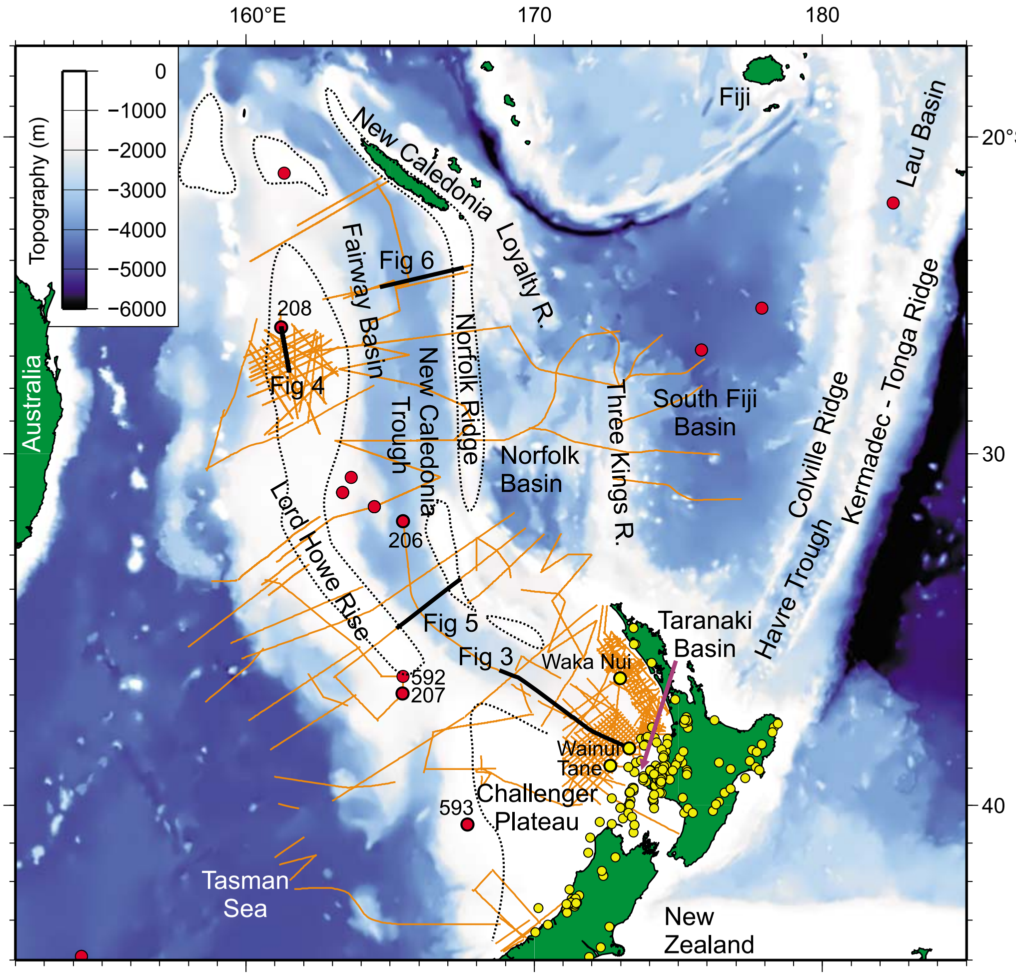

Figure 1. Location of the New Caledonia Trough. Seismic lines used in this study are shown as orange lines, with sections presented in Figures 3–6 indicated. Red circles show DSDP wells, and yellow circles show petroleum wells around New Zealand. Dotted lines show approximate area with evidence for local Eocene-Oligocene coastal erosion.

[4] It is a long-held view that the modern physiography of the New Caledonia Trough formed during Cretaceous rifting associated with the final stages of Gondwana breakup and formation of the Tasman Sea and that the region west of Norfolk Ridge has not undergone any significant tectonic event during Cenozoic time and indeed may preserve Mesozoic sedimentary basins that are similar to those found in eastern Australia [*Burns and Andrews*, 1973; *Crook and Belbin*, 1978; *Eade*, 1988; *King and Thrasher*, 1996; *Lafoy et al.*, 2005; *Uruski and Wood*, 1991; *Uruski et al.*, 2003; *Uruski*, 2008; *Wood*, 1993]. We review data and reasoning that this conclusion is based on, and we use seismic reflection and rock sample data to propose a new hypothesis in which the basin has a two-phase history. Although it is incontrovertible that significant physiographic depressions filled to create sedimentary basins in the region during Cretaceous time (Figure 2), we propose that the first-order physiography of the New Caledonia Trough was substantially modified in Eocene and Oligocene time; widespread subsidence followed uplift and crustal deformation in response to subduction initiation at the plate boundary that has since evolved into the Tonga-Kermadec and Lau-Havre system (Figure 1).

2. Geology of New Zealand

[5] Evidence for Cretaceous normal faulting and crustal thinning is widespread onshore New Zealand and within nearby sedimentary basins [*Cook et al.*, 1999; *King and Thrasher*, 1996; *Laird*, 1993; *Nathan et al.*, 1986]. In most places, rifting was complete by 80 Ma [*Cook et al.*, 1999; *Laird*, 1993; *Nathan et al.*, 1986], but minor fault activity may have locally continued until circa 60 Ma in Taranaki Basin and northwestern South Island [*King and Thrasher*, 1996; *Laird*, 1993; *Nathan et al.*, 1986].

[6] Renewed rifting south of New Zealand was associated with Eocene tectonic reconfiguration of the Australia-Pacific plate boundary [*Sutherland*, 1995; *Turnbull and Uruski*, 1993], concurrent with reverse faulting along the northeastern margin of Taranaki Basin [*Stagpoole and

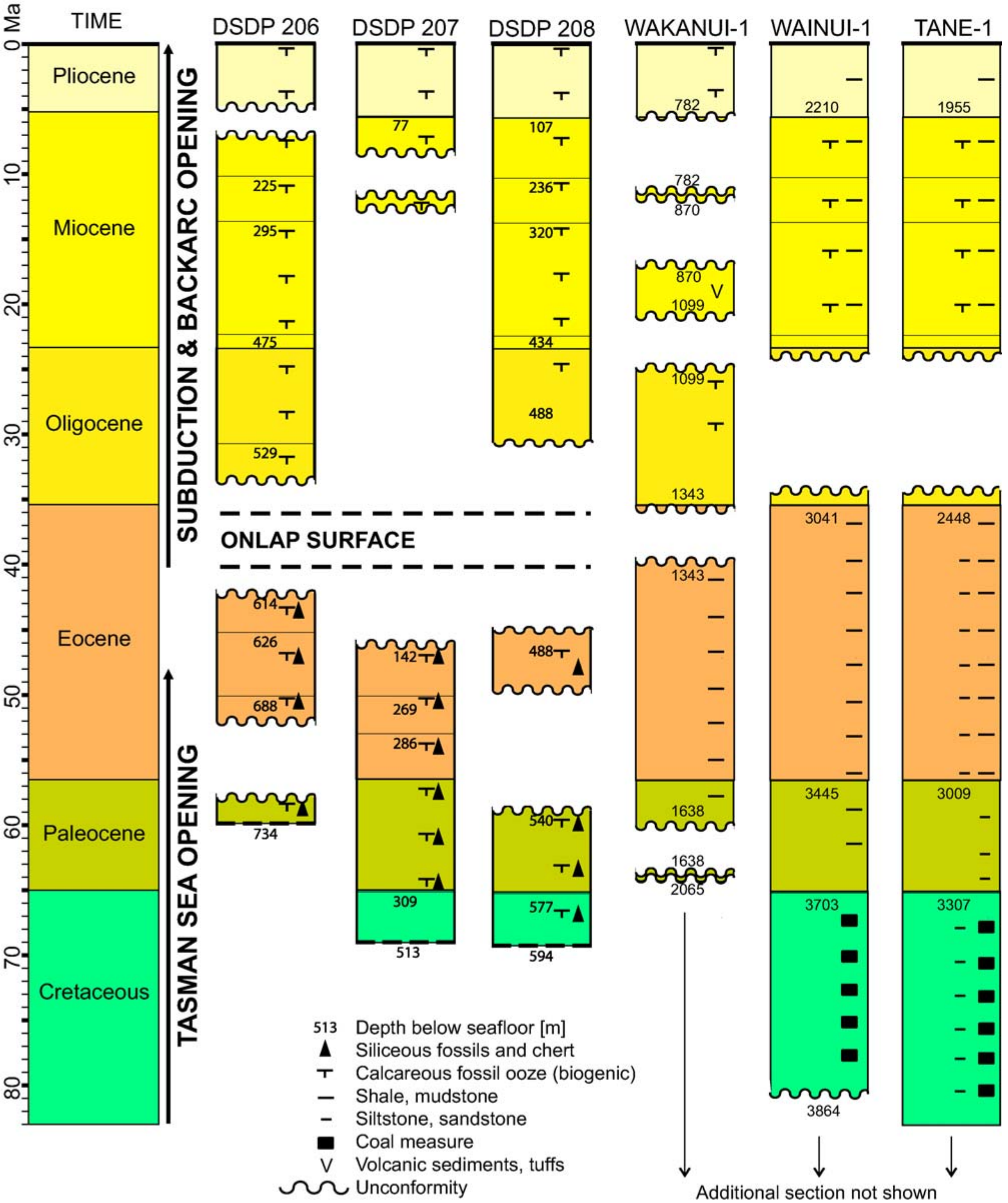

Figure 2. Chronostratigraphic summary of significant wells (Figure 1) and tectonic phases.

Nicol, 2008]. It was not until after late Oligocene time that reverse faulting or dextral strike-slip faulting affected most of the region that is now the land area of New Zealand [*King*, 2000].

[7] The general lack of faulting of Cenozoic strata of Taranaki Basin as they pass into the New Caledonia Trough (Figure 3) and the similarity of seismic reflection character demonstrate that the Cretaceous and Paleocene rift history of

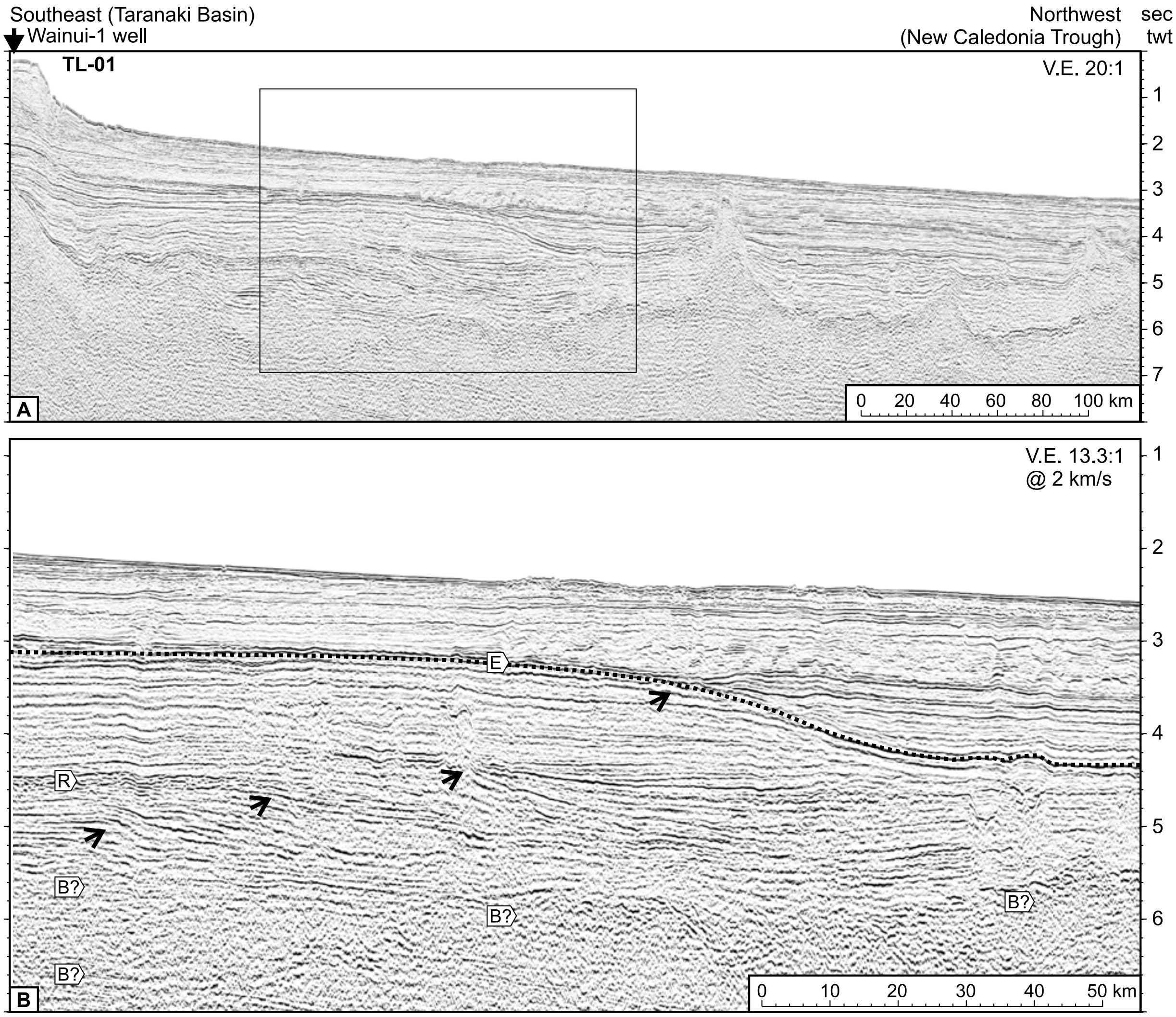

Figure 3. (a) Seismic reflection section showing part of line TL-01, which ties between the Wainui-1 well in Taranaki Basin and DSDP Site 206 in the southern New Caledonia Trough (Figure 1). (b) Enlargement of area outlined by box in Figure 3a. Picks of Basement (B), Rakopi coal measures (R), and the Eocene-Oligocene unconformity (E) used for tectonic subsidence calculations are labeled. Arrows show the progressive migration of a break in slope at the outer edge of a sedimentary fan that built out from the ancient New Zealand landmass during Late Cretaceous to Eocene time.

Taranaki Basin was similar to that at the southern end of the New Caledonia Trough [*King and Thrasher*, 1996; *Uruski and Wood*, 1991; *Uruski et al.*, 2003]. The initial Oligocene and Miocene subsidence of Taranaki Basin has previously been interpreted as platform subsidence caused by mantle flow related to the onset of subduction beneath New Zealand [*Stern and Holt*, 1994].

3. Geology of New Caledonia

[8] There are many similarities between the Late Paleozoic and Mesozoic "basement" geology of New Caledonia and New Zealand. The Cretaceous and Cenozoic sedimentary record of New Caledonia is also similar and reveals a general subsidence trend with time from Late Cretaceous shallow marine sandstone facies to Paleocene siliceous bathyal mudstones and then to Eocene deepwater carbonates [*Paris*, 1981]. Of key significance and difference in New Caledonia is the widespread southwestward emplacement of ophiolitic nappes in middle Eocene to early Oligocene time along low-angle faults [*Aitchison et al.*, 1995; *Auzende et al.*, 2000; *Cluzel et al.*, 2001; *Klingelhoefer et al.*, 2007].

[9] Thermochronology results show that high-pressure metamorphic rocks, inferred to be associated with nappe emplacement, were formed at 44 Ma in northern New Caledonia, and these rocks were then inferred to be exhumed rapidly along extensional detachment faults between 40 and 34 Ma [*Baldwin et al.*, 2007]. This interpretation is con-

sistent with field and seismic reflection observations from southern New Caledonia [*Lagabrielle et al.*, 2005] and other geological observations [*Aitchison et al.*, 1995]. The crustal structure and basin stratigraphy west of New Caledonia are consistent with foreland loading and tilting associated with Eocene and younger southwest verging thrusts along the eastern margin of the sedimentary basin [*Collot et al.*, 2008; *Klingelhoefer et al.*, 2007].

4. Offshore Crust Type and Thickness

[10] On the basis of bathymetry and shipboard geophysical measurements, it was recognized in the 1960s and 1970s that the Tasman Sea abyssal plain and basins east of Norfolk Ridge were likely to have oceanic crustal character and, on the basis of geographic position and onshore geology, were likely to have Cretaceous to Paleogene and Cenozoic ages, respectively [*Karig*, 1971; *Packham and Falvey*, 1971; *Weissel and Hayes*, 1977]. This was confirmed by the Deep Sea Drilling Program (DSDP) [*Burns and Andrews*, 1973].

[11] There is agreement between many authors that the Tasman Sea abyssal plain is composed of oceanic crust that formed between chrons 34y and 24y (circa 83–52 Ma) [*Gaina et al.*, 1998; *Weissel and Hayes*, 1977]. By association, the western physiographic margin of the Lord Howe Rise is inferred to have formed circa 90–80 Ma, though rifting appears to have been asymmetric and may have involved low-angle detachment faults [*Lister et al.*, 1991].

[12] There are many different hypotheses concerning the tectonic development of the North Loyalty, Norfolk, South Fiji, Lau, and Havre basins and the Loyalty, Three Kings, Colville, and Tonga-Kermadec ridges. It is sufficient for this analysis to note a general consensus that these ridges and basins are primarily related to arc and back-arc processes associated with development of the Australia-Pacific plate boundary since Eocene time [*Ballance*, 1999; *Crawford et al.*, 2003; *Davey*, 1982; *DiCaprio et al.*, 2009; *Herzer et al.*, 1997, 2009; *Karig*, 1971; *Malahoff et al.*, 1982; *Mortimer et al.*, 1998, 2007; *Packham and Falvey*, 1971; *Schellart et al.*, 2006; *Sdrolias et al.*, 2004; *Sutherland*, 1999].

[13] It is pertinent to this analysis that the Norfolk Basin does not have typical ocean crust attributes; it has enigmatic physiography and seismic reflection character [*Bernardel et al.*, 2003], a subdued low-amplitude magnetic signature [*Malahoff et al.*, 1982; *Sdrolias et al.*, 2004; *Sutherland*, 1999], and dredge rock samples with compositions that include a wide range of rock types, including ultramafic, volcanic, metamorphic, and sedimentary rocks with old Gondwana-derived zircons and fossil leaves [*Crawford et al.*, 2004; *Meffre et al.*, 2006; *Mortimer et al.*, 1998, 2007].

[14] Crustal models based on bathymetry, gravity, and sediment thickness data predict that the Lord Howe Rise has a crustal thickness of 15–30 km and is inferred to be a continental prolongation of New Zealand and that the New Caledonia Trough has a crustal thickness of 5–15 km [*Klingelhoefer et al.*, 2007; *Uruski and Wood*, 1991; *Wood and Woodward*, 2002; *Woodward and Hunt*, 1971]. In the region that is southwest of New Caledonia, seismic reflection and refraction data confirm that the Lord Howe Rise has a crustal thickness of 23 km and the New Caledonia Trough has a crustal thickness of 6–8 km [*Klingelhoefer et al.*, 2007].

5. Stratigraphy of Lord Howe Rise

[15] Seismic stratigraphy of the Lord Howe Rise can be broadly divided into three units: (1) a basal faulted sequence with normal-faulted tilted geometry, internal unconformities, and variable reflection character (Figure 4a); (2) an overlying blanket sequence of continuous moderate-amplitude reflections (labeled Sag1 in Figure 4); and (3) an upper unit characterized by low-amplitude continuous internal reflections (labeled Sag2 in Figure 4), with either a moderate-amplitude reverse-polarity reflection or a flat high-amplitude normal-polarity reflection at its base. DSDP legs 21, 29, and 90 have investigated the Lord Howe Rise and Challenger Plateau. The discovery of rhyolites dated at 97 ± 4 Ma at the base of a generally undisturbed sequence of Cretaceous-Cenozoic marine sediments in DSDP Borehole 207 and a similar sedimentary sequence in DSDP Borehole 208 is consistent with the hypothesis that the Lord Howe Rise rifted from Australia during Cretaceous inception of Tasman Sea spreading and has been in a passive setting since then [*Burns and Andrews*, 1973; *Tulloch et al.*, 2009; *van der Lingen*, 1973]. However, a significant unconformity of Eocene to Miocene age was recognized in DSDP boreholes on the Lord Howe Rise and in the New Caledonia Trough (Figure 2), and unconformity development was inferred to have been associated with and caused by significant regional oceanographic changes at that time [*Burns and Andrews*, 1973; *Kennett et al.*, 1975, 1986]. This unconformity approximately corresponds to the base of the upper seismic stratigraphic unit, and the local reverse-polarity reflection at the base of the unit is explained by a downward increase in biogenic silica and corresponding decrease in density [*Burns and Andrews*, 1973; *Shipboard Scientific Party*, 1973c; *Collot et al.*, 2008].

[16] At DSDP Site 208, on the northern Lord Howe Rise (Figures 1, 2, and 4), late Oligocene (29–24 Ma) and younger strata (correlated with seismic unit Sag2) overlie early middle Eocene (49–41 Ma) and older strata (Sag1 and deeper) [*Shipboard Scientific Party*, 1973c; *Cooper*, 2004]. Northwest of DSDP Site 208, a regional Eocene to Oligocene unconformity is associated with minor reverse faulting and, based on dredged rock samples, subsidence from late Eocene shallow marine calcarenite depositional environments to Oligocene deepwater biogenic sediment deposition [*Exon et al.*, 2006].

[17] At DSDP Site 207, on the crest of the southern Lord Howe Rise, late Eocene, Oligocene, and early Miocene strata are absent. Sediment above this unconformity has an age of 15–13 Ma and contains reworked or slumped late Eocene (37–34 Ma) microfossils, and the youngest strata beneath the unconformity have an age of 43–37 Ma [*Shipboard Scientific Party*, 1973a; *Cooper*, 2004]. Nearby at DSDP Site 592, late Eocene and early Oligocene nannofossil oozes, near the base of the upper seismic unit, contain reworked middle and late Eocene microfossils and abundant *Zygrhablithus bijugatus* and *Braarudosphaera*

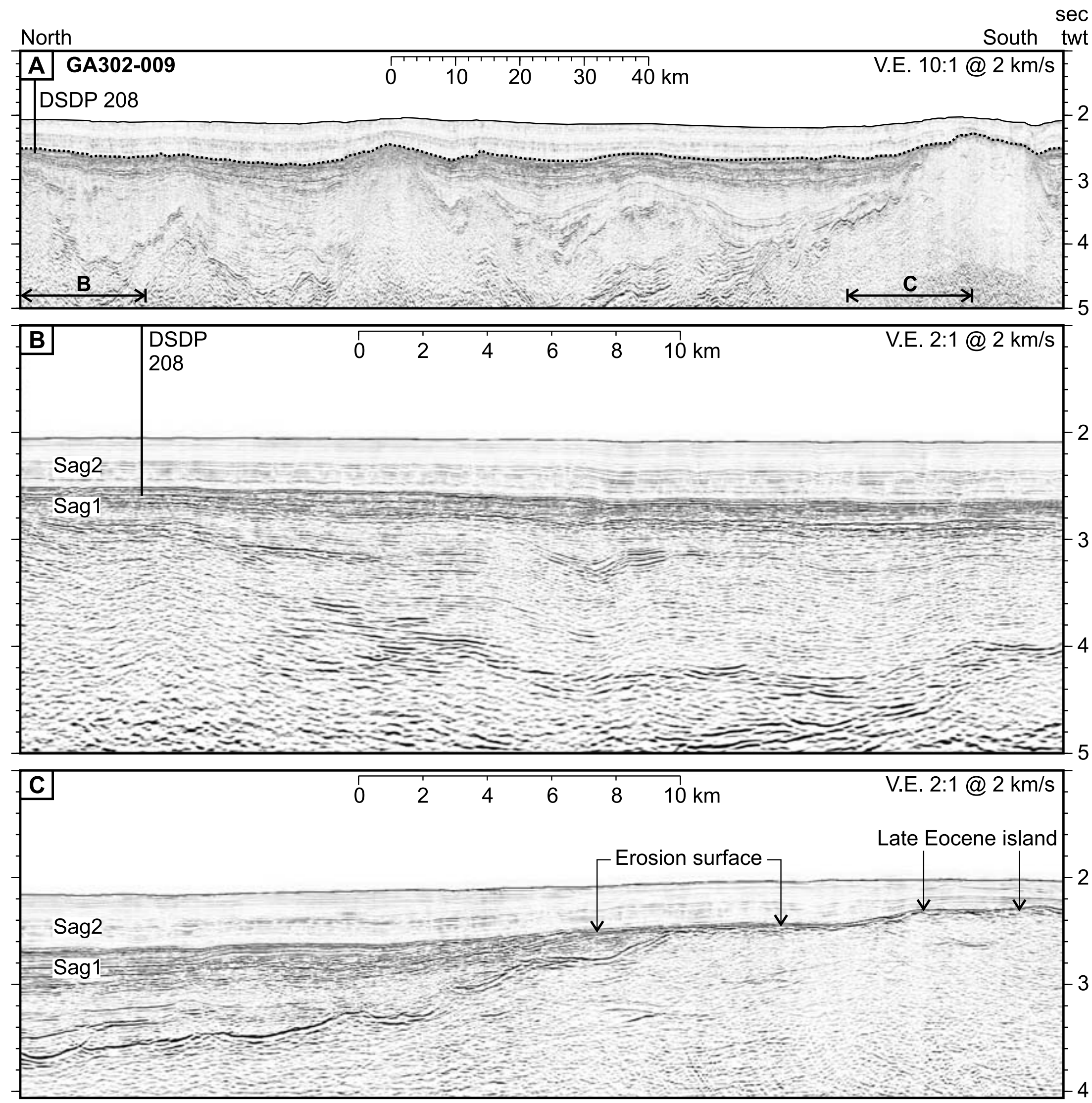

Figure 4. (a) Seismic reflection section GA302-009 on the northern Lord Howe Rise showing seismic stratigraphy near DSDP Site 208. (b) In most places, the Eocene-Oligocene unconformity is marked by a reverse-polarity reflection and a change in reflection character below and above: we identify the seismic stratigraphic units defined by the surface as Sag1 and Sag2, respectively (see main text). (c) We interpret flat surfaces that truncate underlying reflections and have high reflection amplitude, as sea level modulated erosion surfaces and possibly, in some cases, fossil biogenic reefs. Areas of relict late Eocene land are identified in a number of places (e.g., see Figures 1 and 4c).

bigelowi, which indicate a shallow water paleoenvironment [*Kennett et al.*, 1986].

[18] Farther south at DSDP Site 593, on the western Challenger Plateau, volcanogenic sediments are associated with the Eocene-Oligocene boundary (34 Ma), which approximately corresponds to the depth of an unconformity identified on seismic data, but no deeper strata were sampled [*Nelson et al.*, 1986]. Paleoenvironmental analyses were not able to detect significant changes in paleowater depth above or below the Eocene to Oligocene unconformity at DSDP sites that sampled it, but the sensitivity of fossil indicators is low within the inferred bathyal paleoenvironments, and benthic foraminifera are generally poorly preserved over this interval [*Shipboard Scientific Party*, 1973a, 1973c].

6. Stratigraphy and Structure of New Caledonia Trough

[19] The seismic stratigraphy of southern New Caledonia Trough can be divided into three units: (1) an upper unit of moderate-amplitude horizontal continuous reflectors that onlap the basin margin; (2) an underlying unit of high-amplitude or moderate-amplitude continuous or semi-continuous reflectors that are weakly deformed, slightly

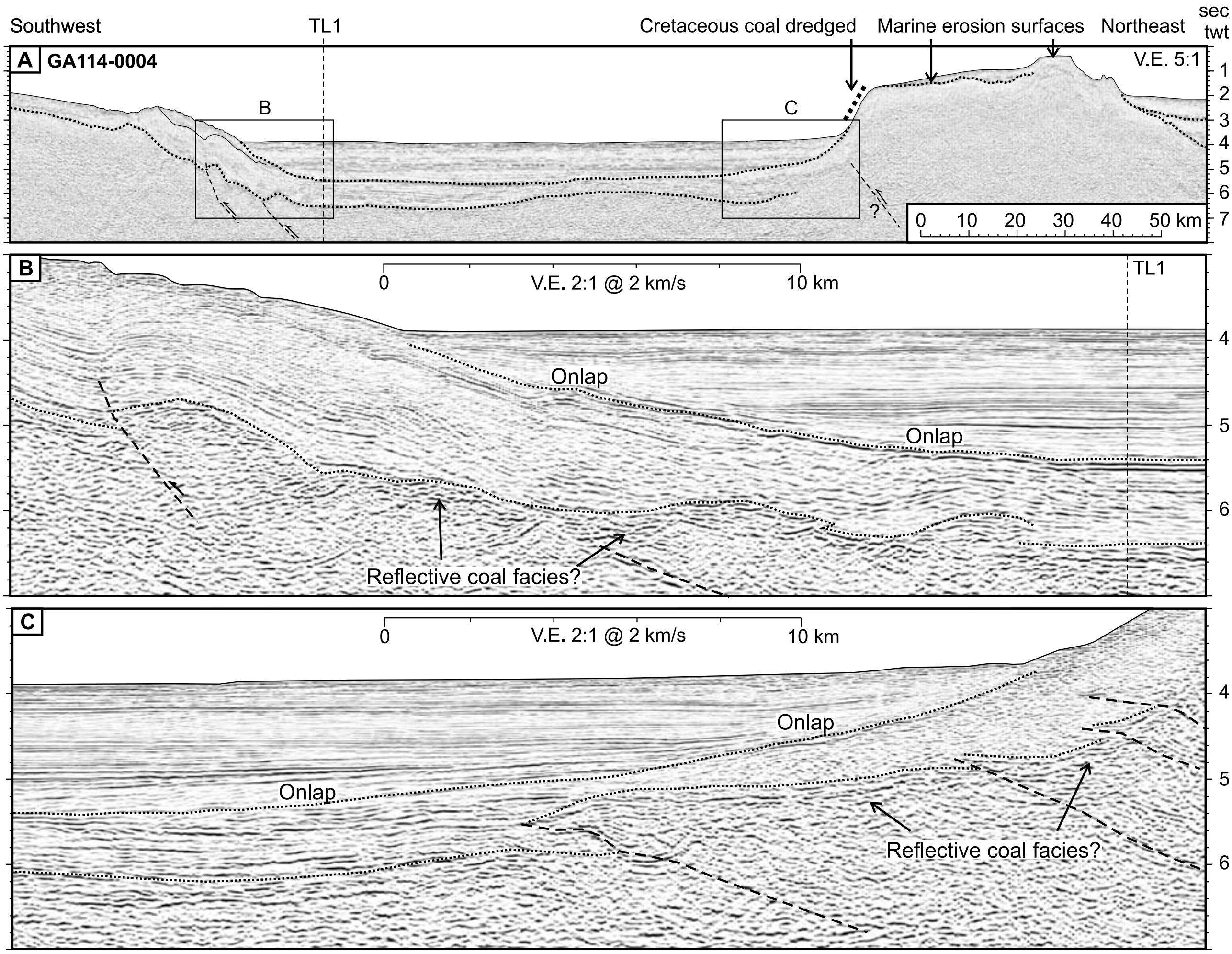

Figure 5. (a) Seismic reflection section GA114-0004 across southern New Caledonia Trough, between the Lord Howe Rise and the West Norfolk and Reinga ridges (for location see Figure 1), (b and c) with enlargement of stratigraphic relationships at the physiographic basin margins. Strata beneath the Eocene-Oligocene onlap surface are slightly folded and faulted, and the lower unit has a similar thickness beneath the margin and axis of the New Caledonia Trough. Marine erosion surfaces of inferred Eocene and Miocene age are indicated on the northeast basin flank (Figure 5a).

tilted, folded, and reverse faulted and locally show evidence for slumping that was contemporaneous with development of the onlap surface above; and (3) a basal sequence of variable thickness that appears from the asymmetric half graben geometries to be normal faulted in many places, though there is local reverse-fault deformation too. We have amalgamated the lower two units in our diagrams (see Figures 5 and 6).

[20] The geometry of the reflectors suggests that the onlap surface at the base of the upper seismic stratigraphic unit corresponds to the time of formation of the present New Caledonia Trough physiography. This conclusion is based on the lack of substantial change in average thickness of the lower two units as basin flanks are traversed; the tilting of reflectors to roughly parallel with the seabed on the basin flanks, with associated slump deposits inferred; and an onlap surface within the basin.

[21] DSDP 206 (3196 m water depth) is the only well that has directly sampled the seismic stratigraphic unit boundary within the New Caledonia Trough (Figures 1 and 2). Nannofossil and foraminifera oozes overlie an Eocene to Oligocene unconformity, below which there are better lithified and locally deformed calcareous oozes of Eocene and Paleocene age. Sediment immediately above the unconformity has an age of 27–25 Ma and immediately beneath the unconformity has an age of 43–37 Ma [*Shipboard Scientific Party*, 1973b; *Cooper*, 2004]. Benthic foraminifera are rare in DSDP 206 cores, except near the base of the Eocene and Paleocene section, where a paleowater depth environment of middle or lower bathyal is inferred (~600–1500 m, possibly 600–3500 m). The maximum estimate of water depth is weakly constrained by fossil evidence because of a lack of appropriate modern or ancient analogs. The depth of the Eocene-Oligocene unconformity approximately corresponds

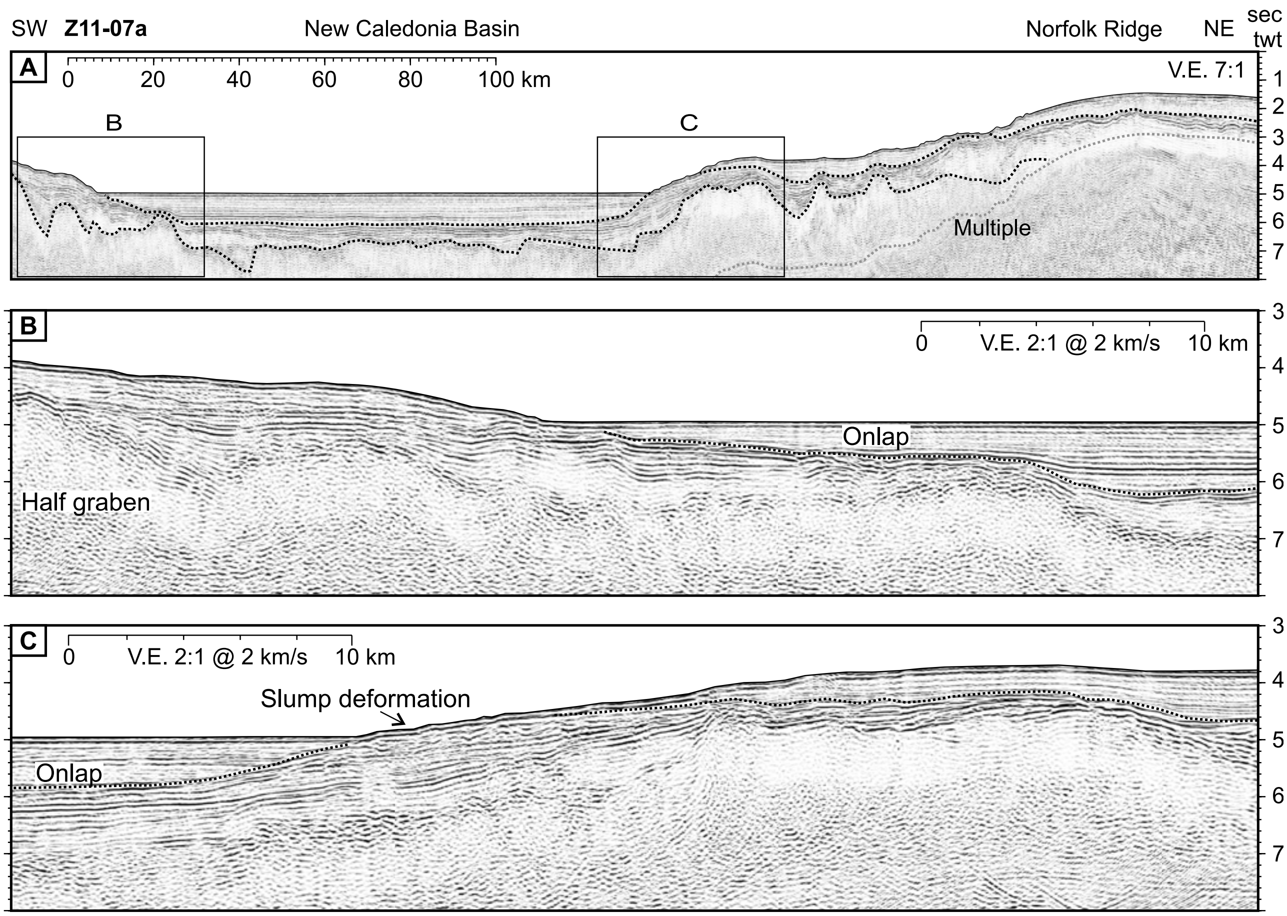

Figure 6. (a) Seismic reflection section Z11-07a across northern New Caledonia Trough, between the Lord Howe Rise and Norfolk Ridge (for location see Figure 1), (b and c) with enlargement of stratigraphic relationships at the basin margins. The onlap surface (B, C) has an inferred middle Eocene to early Miocene age [*Collot et al.*, 2008], and there is evidence for slumping along the basin margin (C). The unconformity surface is erosional along Norfolk Ridge. Asymmetric normal-faulted half grabens are visible beneath the onlap surface (B); the regional thickness of Cretaceous-Eocene sediments is similar beneath the axis and flanks of New Caledonia Trough and increases farther west into the Fairway Basin [*Collot et al.*, 2008; *Lafoy et al.*, 2005].

to the two-way travel time of the onlap surface identified at the base of the upper unit on seismic sections, using reasonable velocities and well log correlation methods.

[22] Boreholes in Taranaki Basin can be tied to New Caledonia Trough using seismic reflection data (Figures 1 and 2). Postrift Late Cretaceous coal measures (Rakopi Formation) are interpreted as a deltaic topset within a Cretaceous-Eocene delta and marine shelf system [*King and Thrasher*, 1996; *Uruski and Baillie*, 2002; *Uruski*, 2008]. The location of the Late Cretaceous coastal environment is now found at 1200–1700 m water depth and 3000–4000 m (2.2–2.8 s two-way traveltime) below seabed. Near New Zealand, it is known from many wells and outcrops that there was rapid and widespread marine flooding of Taranaki Basin during Oligocene time [*King and Thrasher*, 1996; *Stern and Holt*, 1994].

[23] Carbonaceous sediment containing Late Cretaceous pollen is exposed on and has been dredged from the flank of West Norfolk Ridge (Figure 3) [*Herzer et al.*, 1999]. Recent zircon dating has shown that detrital zircons from this dredged carbonaceous sediment have a statistical age peak at 96 ± 4 Ma, which is indistinguishable from the 97 ± 4 Ma age of zircons from rhyolites sampled at DSDP Site 207 [*Mortimer et al.*, 2010]. Although there may be alternate sources for these detrital zircons, the result provides a tantalizing suggestion that sediments were being transported across the region that is now the New Caledonia Trough.

[24] Near New Caledonia, the Eocene-Oligocene unconformity in DSDP 208 has been tied to a seismic reflector that represents the event horizon that corresponds to Eocene overthrusting of New Caledonia along the northeast margin of the New Caledonia Trough and the emplacement of nappes within New Caledonia [*Aitchison et al.*, 1995; *Collot et al.*, 2008]. Immediately south of New Caledonia, we observe much less faulting and basin asymmetry (Figure 6), and there is a remarkable geometrical similarity to the

southern New Caledonia Trough (Figure 5), which confirms the extent of the Eocene unconformity.

7. Evidence for Large Values of Tectonic Subsidence

[25] Tectonic subsidence is calculated as the subsidence that would have occurred if there was no loading effect by younger sediment and the basin was filled only with water. Tectonic subsidence can only be estimated if the level of an ancient sea level can be identified and mapped over a region, or minimum values can be calculated if sediments are known to be deposited above sea level (e.g., coal facies). Details of tectonic subsidence calculations are given in the auxiliary material.[1]

[26] The southern end of New Caledonia Trough has been mapped in considerable detail using seismic reflection methods and tied back to several Taranaki Basin wells [*Uruski and Baillie*, 2002; *Uruski et al.*, 2003]. Late Cretaceous Rakopi Formation coal measures are flat lying and continuous, with little or no faulting (Figure 3). They are clearly recognizable by their discontinuous high-amplitude seismic reflection facies and can be reliably mapped over a region of ~10,000 km^2. Tectonic subsidence values calculated from this unit, assuming it was deposited near sea level, are in the range 2200–2500 m.

[27] It is also possible to derive tectonic subsidence values from sediment facies indicative of shallow marine or shelf environments. Such facies of Paleocene and Eocene age have been sampled in the Tane-1 and Waka Nui-1 wells and can be tentatively recognized from the geometry of the prograding fan system that built out into Taranaki Basin (Figure 3). On the basis of estimates of paleowater depths <100 m, we determine tectonic subsidence values of 1800–2200 m. It is notable that these values are similar to previous determinations from Taranaki Basin farther southeast [*Stern and Holt*, 1994].

[28] On the northern Lord Howe Rise, we identify a high-amplitude reflection with flat geometry at the base of the upper seismic stratigraphic unit. This boundary locally truncates underlying reflections and is interpreted as evidence for wave erosion: we infer the high reflection amplitude as biogenic reefs or hard rock surfaces that formed at the time of its development. Evidence from DSDP 208 indicates that this unconformity formed during Eocene to Miocene time. An example of this seismic reflection facies is shown in Figure 4. We calculate 1300–1800 m of tectonic subsidence from this surface in the region just south of DSDP 208.

[29] We identify two wave-cut surfaces on West Norfolk Ridge (Figure 5). The deeper surface, which we interpret on the basis of seismic facies similarities to the Lord Howe Rise and the adjacent onlap surface in New Caledonia Trough to be a correlative of the Eocene to Oligocene surface, has undergone 1100–1300 m of tectonic subsidence. The shallower surface is observed to the northeast at a number of locations and can be tied through its accompanying unconformity and onlap surface through a grid of seismic reflection lines to early Miocene strata of eastern New Zealand [*Herzer et al.*, 1997]. This younger surface has undergone tectonic subsidence of ~300 m. We suggest that the transient uplift that we infer either side of New Caledonia Trough was also partly responsible for creating elevated Oligocene environments in the Norfolk Basin, where fossil Oligocene leaves and shallow water limestones have been recovered [*Meffre et al.*, 2006].

8. Requirement for a Two-Phase Tectonic Model

[30] It has been widely assumed that the New Caledonia Trough and eastern margin of Lord Howe Rise were formed as physiographic features at the same time as the Tasman Sea margins and by a similar rifting process that culminated in ocean crust formation circa 85–80 Ma. The following observations cannot be readily explained by this single-tectonic-phase rift-drift model.

[31] Stratal geometries beneath the New Caledonia Trough suggest that the modern physiography formed at the time of the middle Eocene-Oligocene unconformity. Strata younger than the unconformity are close to horizontal and onlap the basin margins, whereas older strata are tilted, follow topography, and are slumped (Figures 5 and 6). Hence, there is clear evidence for tectonic modification and establishment of physiography similar to present in Eocene to Oligocene time.

[32] There is not a perfect correlation between sediment thickness beneath the Eocene unconformity and the location of the modern physiographic trough; in some places, there is a similar sediment thickness beneath the unconformity on the flanks and in the center of the New Caledonia Trough (Figures 5 and 6). However, at a larger scale (2000 by 300 km), there is a general correlation between the location of the Cretaceous fore arc, the locations of significant Cretaceous sedimentary basins, and the modern physiographic trough. Hence, we infer Gondwana margin inheritance as a controlling factor for both Cretaceous and younger physiography but note that in some locations there has been a significant shift in the locations of sediment depocenters and style of sedimentation between Cretaceous and Neogene time.

[33] Platform morphology on either side of New Caledonia Trough indicates a phase of Eocene-Oligocene uplift to near sea level, followed by rapid Oligocene-Miocene subsidence of ~1100–1800 m (Figures 3 and 4). The observations on Lord Howe Rise are 400–700 km distant from the nearest significant shallow crustal Eocene-Miocene faults, which are along the Norfolk Ridge (Figure 1).

[34] What mechanisms could explain the large spatial scale (>700 km half wavelength), the amplitude of 1–2 km of transient uplift-subsidence, and the 200–300 km width of ~2 km of permanent subsidence in the New Caledonia Trough? Candidates are (1) local isostatic response to crustal thinning, (2) regional isostatic response to an imposed load, (3) cooling of the lithosphere, and (4) changes in dynamic support from mantle circulation beneath the lithosphere. Tectonic subsidence predicted between 70–40 Ma and 0 Ma

[1]Auxiliary materials are available in the HTML. doi:10.1029/2009TC002476.

based upon passive cooling of a rift basin with a stretching factor in the range 2.0–3.0 that ended rifting at 80 Ma (mechanism 3, our null hypothesis) would be 900–1600 m and would follow an exponential decay after rifting [*McKenzie*, 1978; *Sclater and Christie*, 1980], but the observed values are larger than expected and have most subsidence during the interval circa 40–30 Ma, so post-Cretaceous rift thermal subsidence is not a good fit. We have clear evidence that there was very little or no Eocene or Oligocene upper crustal extension in the New Caledonia Trough or on Lord Howe Rise, but we suggest that crustal thinning by delamination occurred (mechanism 1). We see evidence for significant regional loading (mechanism 2) only in the most northern part of New Caledonia Trough adjacent to New Caledonia, where southwest directed thrusting is apparent and there are paired topographic and free-air gravity anomalies across an asymmetric basin, but it cannot fully explain the amplitude of subsidence. If delamination removed the entire mantle lithosphere beneath Lord Howe Rise, then we would expect thermal uplift followed by ~1400 m of tectonic subsidence since 40 Ma (mechanism 3), so this is one explanation for the regional transient signal. It is outside the scope of this paper to model dynamic topography from mantle flow (mechanism 4), but we speculate that this did play a role in providing part of the broad (>700 km) transient signal.

9. A New Hypothesis: Delamination of the Lithosphere

[35] We suggest that there was a reconfiguration of basin and ridge topography in Eocene to Oligocene time during initiation of the Australia-Pacific convergent plate boundary. Hence, what is now referred to as the New Caledonia Trough subsided to become a major physiographic feature. We believe that a model of formation involving subduction initiation can explain (1) the variation in crustal thickness and hence present physiography, (2) the geometry of basin strata, (3) the long-wavelength large-amplitude uplift and subsequent subsidence of basin flanks, and (4) the large values of Cenozoic tectonic subsidence that are not accompanied by upper crustal deformation in the New Caledonia Trough. In addition, we show that such a hypothesis is generally consistent with the known onshore geology of New Caledonia and New Zealand, and we suggest a reason for the basin geometry and subsequent pattern of arc volcanism.

[36] The New Caledonia Trough is now close to isostatic equilibrium, and a shallow (~15–17 km) Moho depth is confirmed [*Klingelhoefer et al.*, 2007], so we must explain how the crust beneath the basin became thin. The occurrence of Late Cretaceous and Paleocene nonmarine strata at the southern end of the basin suggests either that the crust was thicker when the strata were deposited or that the basin was extremely far from isostatic equilibrium at that time. New Zealand is widely regarded as being in a passive tectonic environment during that time interval [*King and Thrasher*, 1996; *King*, 2000; *Laird*, 1993], so it seems most unlikely that very large dynamic topography signals can be invoked to explain a highly anomalous elevation of thin crust. Therefore, we suggest that thin crust is causally related to Eocene-Oligocene basin subsidence.

[37] Pervasive crustal strain cannot be invoked as a mechanism to produce the relatively thin crust beneath New Caledonia Trough during Eocene-Oligocene time. This is because the observed stratal geometry includes only minor faulting of Paleocene and Eocene strata and very little total fault displacement and the limited fault movements have reverse sense. Therefore, we suggest that a detachment fault must have been involved in removal of the lower crust.

[38] Our model involves four phases (Figure 7): (1) lithospheric thickening that was detached from the upper crust by low-angle faults that surfaced near Norfolk Ridge; (2) removal of the lower crust as the lithospheric root became gravitationally unstable, delaminated, and sank to form a slab; (3) rapid foundering and rollback of the slab, trench migration, and back-arc extension that may have reversed the sense of lower crustal detachment; and (4) finally, a thermal relaxation phase. During the initial phase (Figure 7d), reverse faulting and folding had large geographic extent but were generally only associated with very small amounts of upper crustal convergence. We infer that most of the surface convergence occurred east of Norfolk Ridge, which may have grown through crustal thickening at that time. When sufficient lithospheric root had grown, which numerical models suggest is after 100–150 km of convergence for normal lithosphere [*Gurnis et al.*, 2004], then the root became unstable, detached, and formed a discreet slab. At the plate motion rates that are predicted for that time, this was accomplished in just 2–4 Myr [*Sutherland*, 1995], though less convergence may have been required if the region was preconditioned by Cretaceous Gondwana subduction. As the slab delaminated, we suggest that it removed the lower crust beneath New Caledonia Trough, causing local surface subsidence in isostatic response, but reverse faulting and crustal thickening may have continued along the east margin of Norfolk Ridge (Figure 7c). However, as mantle flowed upward and sideways beneath the Lord Howe Rise and New Caledonia Trough, to preserve continuity of volume, a broad dynamic and thermal uplift of the Lord Howe Rise region occurred. The region of uplift broadened toward Norfolk Ridge as the slab rolled back and the thrust front migrated east (Figure 7b). The rapid flow of mantle material is inferred to have driven back-arc extension and detachment faulting, which surfaced along the Norfolk Ridge System as observed in New Caledonia [*Baldwin et al.*, 2007]. Finally, as the subduction system stabilized to the east, mantle flow rates beneath New Caledonia Trough decreased, dynamic topography died away, and then thermal diffusion led to further subsidence (Figure 7a).

[39] We suggest that many enigmatic aspects of the Norfolk Basin can be explained if it is partly composed of the exhumed footwall of an extensional detachment system that previously lay beneath Norfolk Ridge. Indeed, a variant on the model presented in Figure 7 has the lower crust of New Caledonia Trough exhumed along a low-angle detachment toward Norfolk Basin during the slab foundering phase. An attraction of this variant is that the relatively low density of the crust does not need to be overcome. However, by analogy with other extensional detachment systems, a

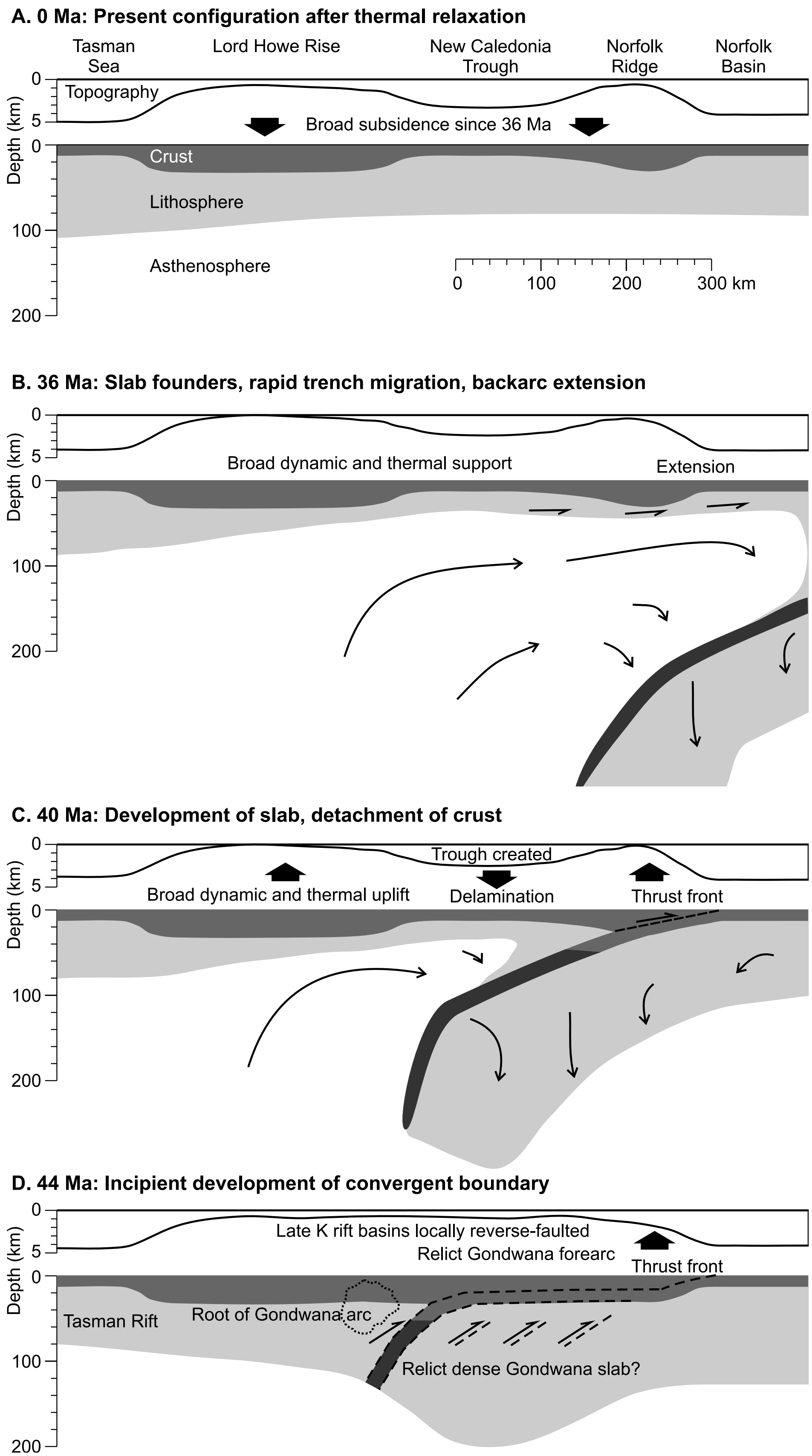
A. 0 Ma: Present configuration after thermal relaxation
Tasman Sea
Lord Howe Rise
New Caledonia Trough
Norfolk Ridge
Norfolk Basin
Topography
Broad subsidence since 36 Ma
Crust
Lithosphere
Asthenosphere
Depth (km)
0
5
100
200
0 100 200 300 km
B. 36 Ma: Slab founders, rapid trench migration, backarc extension
Broad dynamic and thermal support
Extension
C. 40 Ma: Development of slab, detachment of crust
Trough created
Broad dynamic and thermal uplift
Delamination
Thrust front
D. 44 Ma: Incipient development of convergent boundary
Late K rift basins locally reverse-faulted
Relict Gondwana forearc
Thrust front
Tasman Rift
Root of Gondwana arc
Relict dense Gondwana slab?

Figure 7

criticism of this variant is that it would likely produce Eocene-Oligocene normal faulting within New Caledonia Trough, but this is not observed.

[40] We note that the location of New Caledonia Trough coincides with high-amplitude magnetic anomalies that are interpreted to be caused by Mesozoic Gondwana arc rocks. Hence, the location where Cenozoic subduction is inferred to have started is approximately the same as where Gondwana subduction is inferred to have ceased [*Davy et al.*, 2008; *Mortimer et al.*, 1999; *Sutherland*, 1999], and relict chemical or thermal anomalies and inherited weaknesses may have played a role in controlling the location of Cenozoic subduction initiation. If the Cretaceous subduction of very young and relatively thick oceanic crust is what led to the demise of Gondwana subduction [*Davy et al.*, 2008], then it may be that the New Caledonia Trough was previously underlain by a very shallowly dipping oceanic lower crust of this type.

[41] Finally, we identify that west dipping subduction is in conflict with some previous models, most notably for the South Loyalty Basin [*Cluzel et al.*, 2001], even though most authors agree that marginal basins and ridges immediately east of Norfolk Ridge were formed in response to the development of subduction since middle Eocene time [*Ballance*, 1999; *Davey*, 1982; *Herzer et al.*, 1997; *Karig*, 1971; *Malahoff et al.*, 1982; *Mortimer et al.*, 1998, 2007; *Packham and Falvey*, 1971; *Schellart et al.*, 2006; *Sutherland*, 1999]. We suggest that substantial mantle flow, almost certainly associated with subduction initiation, is required to produce the profound Eocene-Miocene physiographic effects observed (this paper) at the New Caledonia Trough and Lord Howe Rise, which is 300–700 km west of Norfolk Ridge. Hence, if this mantle flow was connected to a subduction trench near (or east of) Norfolk Ridge, then a west dipping slab at depth seems to be required.

[42] We do not rule out the possibility of east dipping convergence zones, in addition to the large-scale west dipping system that we infer beneath New Caledonia Trough, particularly during the early stages of subduction zone evolution when reverse faulting was widespread. This is consistent with some previous suggestions [*Cluzel et al.*, 2001; *Crawford et al.*, 2003; *Schellart et al.*, 2006]. Indeed, there is little doubt that the thrust-faulted zone bounding the southwest margin of New Caledonia has northeast dip and southwestward vergence [*Aitchison et al.*, 1995; *Auzende et al.*, 2000; *Cluzel et al.*, 2001; *Collot et al.*, 2008]. Further geological and geophysical data are required to determine the surface manifestations and hence the precise geometrical evolution of the subduction and back-arc spreading systems, but we also require an explanation for the significant topographic effects that were felt so far west of Norfolk Ridge. We assert that positive seismic velocity anomalies in the New Caledonia Trough region at depths of 800–1500 km are consistent, within the uncertainties of tomographic models [*Schellart et al.*, 2009], with foundering of lithosphere beneath New Caledonia Trough and toward the north and east during the interval circa 40–20 Ma.

10. Lithospheric Delamination in a Global Context

[43] We present evidence that the New Caledonia Trough formed in Eocene and Oligocene time and suggest its lower crust was removed by delamination of the lithosphere during initiation of the subduction zone that has since evolved into the Tonga-Kermadec system. We propose a mechanism of gravitational instability in which convergence caused lithospheric thickening along a preexisting weakness associated with Gondwana subduction, and it may also be that the lithosphere was preconditioned chemically through preservation of the Gondwana slab with its metamorphosed oceanic crust composed of denser phases than surrounding mantle [*Hacker et al.*, 2003]. We see local evidence for convergent deformation that immediately predates formation of the New Caledonia Trough (e.g., Figure 4), and regional convergent plate motion is predicted in late Eocene and Oligocene time [*Sutherland*, 1995].

[44] Models of subduction initiation are shown to become self-sustaining within "normal" lithosphere after 100–150 km of convergence produces a gravitational instability of sufficient size to overcome the strength of the lithosphere [*Gurnis et al.*, 2004], but less convergence would be required if the lithosphere was anomalously dense or weak. Lithospheric and crustal delamination through spontaneous gravitational instability has been suggested to explain Miocene-Quaternary foundering of the root of the Sierra Nevada in the western United States [*Zandt et al.*, 2004]. A similar drip-like model has been suggested to explain formation of the Pannonian-Carpathian region in central Europe [*Houseman and Gemmer*, 2007]. An alternative model of the Carpathian region invokes a sheet-like geometry of mantle foundering and includes lateral variation in thermomechanical properties [*Cloetingh et al.*, 2004]. This alternate model is intended to explain subduction death rather than initiation, but it bears striking similarities to what we suggest happened in the New Caledonia Trough and includes delamination of lower crustal material.

[45] We speculate that there may be large-scale geodynamic similarities between the evolution of the Tonga-Kermadec subduction system and the Alpine-Mediterranean region. Several subduction initiation, rollback, and slab detachment events have been identified that have formed the main tectonic features of southern Europe during Cenozoic time [*Faccenna et al.*, 2001; *Hafkenscheid et al.*, 2006; *Jolivet and Faccenna*, 2000; *Wortel and Spakman*, 2000], and widespread extension was associated with slab rollback

Figure 7. Tectonic model to explain the Cenozoic phase of formation of the New Caledonia Trough: (d) 44 Ma, lithospheric thickening beneath New Caledonia Trough was detached from the upper crust by low-angle thrust faults that exploited the Mesozoic Gondwana subduction interface and surfaced near Norfolk Ridge; (c) 40 Ma, removal of the lower crust as it detaches with the newly developed lithospheric slab; (b) 36 Ma, rapid foundering of the slab, trench migration, and back-arc extension; and (a) thermal relaxation to produce the present configuration.

[*Gueguen et al.*, 1998; *Jolivet et al.*, 2009; *Rollet et al.*, 2002]. Positive dynamic topography associated with mantle flow during rollback has been suggested because extension and thermal processes alone cannot otherwise explain the high values of Neogene subsidence observed around the Mediterranean [*Morley*, 1993]. In the western Mediterranean, lithospheric delamination has previously been proposed to explain the mantle structure [*Calvert et al.*, 2000; *Seber et al.*, 1996], kinematics [*Fadil et al.*, 2006], metamorphic and igneous history [*Platt et al.*, 1998; *Turner et al.*, 1999], and subsidence history [*Docherty and Banda*, 1995].

[46] Slab rollback and extension followed delamination during Tonga-Kermadec subduction initiation and led to the formation of basins between the modern Tonga-Kermadec and Norfolk ridges [*Crawford et al.*, 2003; *Davey*, 1982; *Herzer et al.*, 2009; *Malahoff et al.*, 1982; *Mortimer et al.*, 2007; *Schellart et al.*, 2006]. There is evidence for Cenozoic metamorphic core complexes immediately north of New Zealand [*Mortimer et al.*, 2003, 2008] and within New Caledonia [*Baldwin et al.*, 2007; *Schellart et al.*, 2006], as there is in Europe [*Jolivet et al.*, 2009]. We also see evidence for broad transient dynamic and/or thermal topography with an amplitude of 1100–1800 km across a region of 700 km width between Norfolk Ridge and western Lord Howe Rise. In the Mediterranean, anomalous topography has an amplitude of 500–1500 m and is shown to be similar to dynamic topography computed from a mantle flow model based on the conversion of seismic velocities to buoyancy anomalies [*Shaw and Pysklywec*, 2007].

[47] It is significant that we document evidence for permanent tectonic subsidence of >2 km over a width of ~300 km in the New Caledonia Trough in the clear absence of any Cenozoic extensional faulting. We invoke crustal delamination to achieve this subsidence and suggest that such a process represents one end-member of a continuum between crustal delamination and boudinage to achieve permanent subsidence in a slab rollback environment. It may be that lower crust founders into the mantle during delamination events in more cases than was previously thought and could partially explain discrepancies between high values of observed subsidence and those predicted from reasonable extensional models. In particular, we suggest that such a process may have occurred in the Mediterranean and circum-Pacific regions, where subduction initiation is known to have occurred during Cenozoic time and there are notable similarities in many aspects of the geological histories.

11. Conclusion

[48] We compare two alternate hypotheses. Our null hypothesis is the long-held view that the New Caledonia Trough formed during Late Cretaceous rifting and Gondwana breakup [*Burns and Andrews*, 1973; *Crook and Belbin*, 1978; *Eade*, 1988; *King and Thrasher*, 1996; *Lafoy et al.*, 2005; *Uruski and Wood*, 1991; *Wood*, 1993]. In our alternate hypothesis, we propose a two-stage tectonic process: Cretaceous rift basins formed localized tectonic depressions, and then the major physiographic feature that is now identified as the New Caledonia Trough was formed in Eocene-Oligocene time during a second phase of subsidence associated with lithospheric delamination and initiation of the Australia-Pacific convergent plate boundary.

[49] We agree with previous analyses that sedimentary basins formed during Cretaceous rifting and were controlled approximately by the location of structures associated with the Gondwana fore arc, which is approximately along the line of the New Caledonia Trough [*Collot et al.*, 2009]. However, we find that the first hypothesis alone fails to explain a number of significant observations: (1) stratal geometries suggest that the New Caledonia Trough formed as a major physiographic feature, 200–300 km wide and 2000 km long, at the time of a prominent middle Eocene-Oligocene unconformity and onlap surface; (2) stratal geometries do not indicate that the axis of the New Caledonia Trough in its current configuration precisely aligns with significant Late Cretaceous depocenters; (3) platform morphology on either side of New Caledonia Trough indicates a phase of Eocene-Oligocene uplift to near sea level, followed by rapid Oligocene-Miocene subsidence of ~1100–1800 m; and (4) seismic reflection facies tied to boreholes suggest tectonic subsidence of 1800–2200 m at the southern end of New Caledonia Trough and possibly >3000 m in central parts of New Caledonia Trough since Eocene time.

[50] Our alternate hypothesis is able to explain our observations, and the Eocene timing coincides with the known onset of the modern Australia-Pacific boundary and other major plate motion changes within the Pacific [*Steinberger et al.*, 2004; *Sutherland*, 1995]. It is known that subduction initiated in the western Pacific at about that time, and a similar model of catastrophic subduction initiation has been used to explain the history of the Izu-Bonin-Mariana system [*Gurnis et al.*, 2004; *Hall et al.*, 2003]. Our model is subtly different from previous models of Australia-Pacific subduction development but is generally consistent in the sense that we agree that Eocene-Oligocene subduction processes are involved and that rapid late Eocene-Oligocene trench rollback occurred [*Cluzel et al.*, 2001; *Collot et al.*, 2008; *Crawford et al.*, 2003; *Schellart et al.*, 2006; *Stagpoole and Nicol*, 2008].

[51] In conclusion, we propose a new model for the formation of the New Caledonia Trough that involves subduction initiation by delamination and then rollback of the slab. Removal of lower crust caused a permanent subsidence in Eocene-Oligocene time of a deep (>2000 m) enclosed oceanic basin, ~2000 km long and 200–300 km across, which is now called the New Caledonia Trough. Simultaneously, uplift and localized land developed along the basin flanks, which are now called the Lord Howe Rise and Norfolk Ridge System (Figure 1). Disruption of Late Cretaceous and Paleogene strata was minimal during formation of the New Caledonia Trough and involved only subtle tilting and localized reverse faulting or folding, allowing us to be sure that pervasive crustal strain and boudinage did not create the New Caledonia Trough. Basin formation can only have been possible through the action of at least one detachment fault that allowed the lower crust to either be subducted into the mantle or exhumed to become basement rock of Norfolk Basin. Thermal and dynamic processes associated with lithosphere delamination and mantle flow

produced a broad (>700 km) transient uplift with an amplitude of 1–2 km.

[52] Most other regions in the world that have undergone subduction initiation have a complex history of deformation before or after the initiation event or have an incomplete and disjointed sedimentary record. In the case of Tonga-Kermadec subduction initiation, the region above the newly developed slab escaped deformation and remained below or near sea level, so an almost continuous sediment record of events is preserved, and we can rule out pervasive crustal strain as a subsidence mechanism. Our results support suggestions that a delamination process may also have occurred during Cenozoic time in the Mediterranean and circum-Pacific regions, where subduction initiation is known to have occurred and there are notable similarities in many aspects of their geological histories. We suggest that lithosphere delamination and possible mixing of lower crustal material back into the mantle is more widespread than previously thought and large-scale delamination events are associated with subduction initiation.

[53] **Acknowledgments.** We thank Onno Onken, the Associate Editor, Tim Stern, one anonymous reviewer, Phil Symonds, and Andrew Jones for their constructive reviews. Funding for this work was from the governments of New Zealand, Australia, New Caledonia, and France. We acknowledge permission to publish from the Chief Executive of Geoscience Australia.

References

Aitchison, J., G. L. Clarke, S. Meffre, and D. Cluzel (1995), Eocene arc-continent collision in New Caledonia and implications for regional southwest Pacific tectonic evolution, *Geology*, *23*(2), 161–164, doi:10.1130/0091-7613(1995)023<0161:EACCIN>2.3.CO;2.

Auzende, J. M., S. Van de Beuque, M. Regnier, Y. Lafoy, and P. Symonds (2000), Origin of the New Caledonian ophiolites based on a French-Australian seismic transect, *Mar. Geol.*, *162*(2–4), 225–236, doi:10.1016/S0025-3227(99)00082-1.

Baldwin, S. L., T. Rawling, and P. G. Fitzgerald (2007), Thermochronology of the New Caledonian high-pressure terrane: Implications for middle Tertiary plate boundary processes in the southwest Pacific, in *Convergent Margin Terranes and Associated Regions: A Tribute to W.G. Ernst*, edited by M. Cloos et al., *Spec. Pap. Geol. Soc. Am.*, *419*, 117–134.

Ballance, P. F. (1999), Simplification of the southwest Pacific Neogene arcs: Inherited complexity and control by a retreating pole of rotation, in *Continental Tectonics*, edited by C. MacNiocaill and P. D. Ryan, *Geol. Soc. Spec. Publ.*, *164*, 7–19, doi:10.1144/GSL.SP.1999.164.01.03.

Bernardel, G., L. Carson, S. Meffre, P. Symonds, and A. Mauffret (2003), Geological and morphological framework of the Norfolk Ridge to Three Kings Ridge region: The FAUST-2 survey area, *Geosci. Aust. Rec.*, *2002*(8), 1–75.

Burns, R. E., and J. E. Andrews (1973), Regional aspects of deep sea drilling in the southwest Pacific, *Initial Rep. Deep Sea Drill.Proj.*, *21*, 897–906.

Calvert, A., E. Sandvol, D. Seber, M. Barazangi, S. Roecker, T. Mourabit, F. Vidal, G. Alguacil, and N. Jabour (2000), Geodynamic evolution of the lithosphere and upper mantle beneath the Alboran region of the western Mediterranean: Constraints from travel time tomography, *J. Geophys. Res.*, *105*(B5), 10,871–10,898, doi:10.1029/2000JB900024.

Cloetingh, S., E. Burov, L. Matenco, G. Toussaint, G. Bertotti, P. A. M. Andriessen, M. J. R. Wortel, and W. Spakman (2004), Thermo-mechanical controls on the mode of continental collision in the SE Carpathians (Romania), *Earth Planet. Sci. Lett.*, *218*(1–2), 57–76, doi:10.1016/S0012-821X(03)00645-9.

Cluzel, D., J. C. Aitchison, and C. Picard (2001), Tectonic accretion and underplating of mafic terranes in the late Eocene intraoceanic fore-arc of New Caledonia (southwest Pacific): Geodynamic implications, *Tectonophysics*, *340*(1–2), 23–59, doi:10.1016/S0040-1951(01)00148-2.

Collot, J., L. Géli, Y. Lafoy, R. Vially, D. Cluzel, F. Klingelhöefer, and H. Nouzé (2008), Tectonic history of northern New Caledonia Basin from deep offshore seismic reflection: Relation to late Eocene obduction in New Caledonia, southwest Pacific, *Tectonics*, *27*, TC6006, doi:10.1029/2008TC002263.

Collot, J., R. Herzer, Y. Lafoy, and L. Géli (2009), Mesozoic history of the Fairway-Aotea Basin: Implications for the early stages of Gondwana fragmentation, *Geochem. Geophys. Geosyst.*, *10*, Q12019, doi:10.1029/2009GC002612.

Cook, R. A., R. Sutherland, and H. Zhu (1999), *Cretaceous-Cenozoic Geology and Petroleum Systems of the Great South Basin, New Zealand*, 188 pp., Inst. of Geol. and Nucl. Sci., Lower Hutt, New Zealand.

Cooper, R. A. (2004), *The New Zealand Geological Timescale*, 284 pp., Inst. of Geol. and Nucl. Sci., Lower Hutt, New Zealand.

Crawford, A. J., S. Meffre, and P. Symonds (2003), 120 to 0 Ma tectonic evolution of the southwest Pacific and analogous geological evolution of the 600 to 220 Ma Tasman Fold Belt System, *Spec. Publ. Geol. Soc. Aust.*, *22*, 377–397.

Crawford, A. J., S. Meffre, M. J. Baker, P. G. Quilty, P. E. O'Brien, N. F. Exon, G. Bernardel, and R. H. Herzer (2004), Tectonic development of the SW Pacific 120–0 Ma: Implications from the "Norfolk'n Around" cruise of the Southern Surveyor to the Norfolk Basin-New Caledonia Ridge region, March 2003, in *Dynamic Earth: Past, Present and Future. 17th Australian Geological Convention, Hobart, February 2004*, edited by J. McPhie and P. McGoldrick, *Geol. Soc. Aust. Abstr.*, *73*, 201.

Crook, K. A. W., and L. Belbin (1978), The southwest Pacific area during the last 90 million years, *J. Geol. Soc. Aust.*, *25*, 23–40.

Davey, F. J. (1982), The structure of the South Fiji Basin, *Tectonophysics*, *87*(1–4), 185–241, doi:10.1016/0040-1951(82)90227-X.

Davy, B., K. Hoernle, and R. Werner (2008), Hikurangi Plateau: Crustal structure, rifted formation, and Gondwana subduction history, *Geochem. Geophys. Geosyst.*, *9*, Q07004, doi:10.1029/2007GC001855.

DiCaprio, L., R. D. Müller, M. Gurnis, and A. Goncharov (2009), Linking active margin dynamics to overriding plate deformation: Synthesizing geophysical images with geological data from the Norfolk Basin, *Geochem. Geophys. Geosyst.*, *10*, Q01004, doi:10.1029/2008GC002222.

Docherty, C., and E. Banda (1995), Evidence for the eastward migration of the Alboran Sea based on regional subsidence analysis: A case for basin formation by delamination of the subcrustal lithosphere, *Tectonics*, *14*(4), 804–818, doi:10.1029/95TC00501.

Eade, J. V. (1988), The Norfolk Ridge system and its margins, in *The Ocean Basins and Margins*, vol. 7, *The Pacific Ocean*, edited by A. E. M. Nairn et al., pp. 803–824, Plenum, New York.

Exon, N. F., P. J. Hill, Y. Lafoy, C. Heine, and G. Bernardel (2006), Kenn Plateau off northeast Australia: A continental fragment in the southwest Pacific jigsaw, *Aust. J. Earth Sci.*, *53*(4), 541–564, doi:10.1080/08120090600632300.

Faccenna, C., T. W. Becker, F. P. Lucente, L. Jolivet, and F. Rossetti (2001), History of subduction and back-arc extension in the central Mediterranean, *Geophys. J. Int.*, *145*(3), 809–820, doi:10.1046/j.0956-540x.2001.01435.x.

Fadil, A., P. Vernant, S. McClusky, R. Reilinger, F. Gomez, D. Ben Sari, T. Mourabit, K. Feigl, and M. Barazangi (2006), Active tectonics of the western Mediterranean: Geodetic evidence for rollback of a delaminated subcontinental lithospheric slab beneath the Rif Mountains, Morocco, *Geology*, *34*(7), 529–532, doi:10.1130/G22291.1.

Gaina, C., D. R. Müller, J.-Y. Royer, J. Stock, J. Hardebeck, and P. Symonds (1998), The tectonic history of the Tasman Sea: A puzzle with 13 pieces, *J. Geophys. Res.*, *103*(B6), 12,413–12,433, doi:10.1029/98JB00386.

Gueguen, E., C. Doglioni, and M. Fernandez (1998), On the post-25 Ma geodynamic evolution of the western Mediterranean, *Tectonophysics*, *298*(1–3), 259–269, doi:10.1016/S0040-1951(98)00189-9.

Gurnis, M., C. Hall, and L. Lavier (2004), Evolving force balance during incipient subduction, *Geochem. Geophys. Geosyst.*, *5*, Q07001, doi:10.1029/2003GC000681.

Hacker, B. R., G. A. Abers, and S. M. Peacock (2003), Subduction factory: 1. Theoretical mineralogy, densities, seismic wave speeds, and H2O contents, *J. Geophys. Res.*, *108*(B1), 2029, doi:10.1029/2001JB001127.

Hafkenscheid, E., M. J. R. Wortel, and W. Spakman (2006), Subduction history of the Tethyan region derived from seismic tomography and tectonic reconstructions, *J. Geophys. Res.*, *111*, B08401, doi:10.1029/2005JB003791.

Hall, C. E., M. Gurnis, M. Sdrolias, L. L. Lavier, and R. D. Mueller (2003), Catastrophic initiation of subduction following forced convergence across fracture zones, *Earth Planet. Sci. Lett.*, *212*(1–2), 15–30, doi:10.1016/S0012-821X(03)00242-5.

Herzer, R. H., et al. (1997), Seismic stratigraphy and structural history of the Reinga Basin and its margins southern Norfolk Ridge system, *N. Z. J. Geol. Geophys.*, *40*, 425–451.

Herzer, R. H., R. Sykes, S. D. Killops, R. H. Funnell, D. R. Burggraf, J. Townend, J. I. Raine, and G. J. Wilson (1999), Cretaceous carbonaceous rocks from the Norfolk Ridge system, southwest Pacific: Implications for regional petroleum potential, *N. Z. J. Geol. Geophys.*, *42*, 57–73.

Herzer, R. H., B. W. Davy, N. Mortimer, P. G. Quilty, G. C. H. Chaproniere, C. M. Jones, A. J. Crawford, and C. J. Hollis (2009), Seismic stratigraphy and structure of the Northland Plateau and the development of the Vening Meinesz transform margin, SW Pacific Ocean, *Mar. Geophys. Res.*, *30*(1), 21–60, doi:10.1007/s11001-009-9065-1.

Houseman, G. A., and L. Gemmer (2007), Intra-orogenic extension driven by gravitational instability:

Carpathian-Pannonian orogeny, *Geology*, *35*(12), 1135–1138, doi:10.1130/G23993A.1.

Jolivet, L., and C. Faccenna (2000), Mediterranean extension and the Africa-Eurasia collision, *Tectonics*, *19*(6), 1095–1106, doi:10.1029/2000TC900018.

Jolivet, L., C. Faccenna, and C. Piromallo (2009), From mantle to crust: Stretching the Mediterranean, *Earth Planet. Sci. Lett.*, *285*(1–2), 198–209, doi:10.1016/j.epsl.2009.06.017.

Karig, D. E. (1971), Origin and development of marginal basins in the western Pacific, *J. Geophys. Res.*, *76*(11), 2542–2561, doi:10.1029/JB076i011p02542.

Kennett, J. P., et al. (1975), Cenozoic paleoceanography in the southwest Pacific Ocean, Antarctic glaciation, and the development of the circumantarctic current, *Initial Rep. Deep Sea Drill. Proj.*, *29*, 1155–1169, doi:10.2973/dsdp.proc.29.144.1975.

King, P. R. (2000), Tectonic reconstructions of New Zealand 40 Ma to the present, *N. Z. J. Geol. Geophys.*, *43*, 611–638.

King, P. R., and G. P. Thrasher (1996), *Cretaceous-Cenozoic Geology and Petroleum Systems of the Taranaki Basin, New Zealand*, 243 pp., Inst. of Geol. and Nucl. Sci., Lower Hutt, New Zealand.

Klingelhoefer, F., Y. Lafoy, J. Collot, E. Cosquer, L. Géli, H. Nouzé, and R. Vially (2007), Crustal structure of the basin and ridge system west of New Caledonia (southwest Pacific) from wide-angle and reflection seismic data, *J. Geophys. Res.*, *112*, B11102, doi:10.1029/2007JB005093.

Lafoy, Y., I. Brodien, R. Vially, and N. F. Exon (2005), Structure of the basin and ridge system west of New Caledonia (southwest Pacific): A synthesis, *Mar. Geophys. Res.*, *26*(1), 37–50, doi:10.1007/s11001-005-5184-5.

Lagabrielle, Y., P. Maurizot, Y. Lafoy, G. Cabioch, B. Pelletier, M. Regnier, I. Wabete, and S. Calmant (2005), Post-Eocene extensional tectonics in southern New Caledonia (SW Pacific): Insights from onshore fault analysis and offshore seismic data, *Tectonophysics*, *403*(1–4), 1–28, doi:10.1016/j.tecto.2005.02.014.

Laird, M. G. (1993), Cretaceous continental rifts: New Zealand region, in *Sedimentary Basins of the World*, vol. 2, *South Pacific Sedimentary Basins*, edited by P. F. Ballance, pp. 37–49, Elsevier, Amsterdam.

Lister, G. S., M. A. Etheridge, and P. A. Symonds (1991), Detachment models for the formation of passive continental margins, *Tectonics*, *10*(5), 1038–1064, doi:10.1029/90TC01007.

Malahoff, A., R. H. Feden, and H. S. Fleming (1982), Magnetic anomalies and tectonic fabric of marginal basins north of New Zealand, *J. Geophys. Res.*, *87*(B5), 4109–4125, doi:10.1029/JB087iB05p04109.

Martini, E., and D. G. Jenkins (1986), Biostratigraphic synthesis, Deep Sea Drilling Project Leg 90 in the southwest Pacific Ocean, *Initial Rep. Deep Sea Drill. Proj.*, *90*, 1459–1470.

McKenzie, D. P. (1978), Some remarks on the development of sedimentary basins, *Earth Planet. Sci. Lett.*, *40*(1), 25–32, doi:10.1016/0012-821X(78)90071-7.

Meffre, S., A. J. Crawford, and P. G. Quilty (2006), Arc-continent collision forming a large island between New Caledonia and New Zealand in the Oligocene, *ASEG Extended Abstr.*, *2006*(1), doi:10.1071/ASEG2006ab111.

Morley, C. K. (1993), Discussion of origins of hinterland basins to the Rif-Betic Cordillera and Carpathians, *Tectonophysics*, *226*(1–4), 359–376, doi:10.1016/0040-1951(93)90127-6.

Mortimer, N., R. H. Herzer, P. B. Gans, D. L. Parkinson, and D. Seward (1998), Basement geology from Three Kings Ridge to West Norfolk Ridge, southwest Pacific Ocean: Evidence from petrology, geochemistry and isotopic dating of dredge samples, *Mar. Geol.*, *148*(3–4), 135–162, doi:10.1016/S0025-3227(98)00007-3.

Mortimer, N., A. J. Tulloch, R. N. Spark, N. W. Walker, E. Ladley, A. Allibone, and D. L. Kimbrough (1999), Overview of the Median Batholith, New Zealand: A new interpretation of the geology of the Median Tectonic Zone and adjacent rocks, *J. Afr. Earth Sci.*, *29*(1), 257–268, doi:10.1016/S0899-5362(99)00095-0.

Mortimer, N., R. H. Herzer, N. W. Walker, A. T. Calvert, D. Seward, and G. C. H. Chaproniere (2003), Cavalli Seamount, Northland Plateau, SW Pacific Ocean: A Miocene metamorphic core complex?, *J. Geol. Soc.*, *160*, 971–983, doi:10.1144/0016-764902-157.

Mortimer, N., R. H. Herzer, P. B. Gans, C. Laporte-Magoni, A. T. Calvert, and D. Bosch (2007), Oligocene–Miocene tectonic evolution of the South Fiji Basin and Northland Plateau, SW Pacific Ocean: Evidence from petrology and dating of dredged rocks, *Mar.Geol.*, *237*(1–2), 1–24, doi:10.1016/j.margeo.2006.10.033.

Mortimer, N., W. J. Dunlap, J. M. Palin, R. H. Herzer, F. Hauff, and M. Clark (2008), Ultra-fast early Miocene exhumation of Cavalli Seamount, Northland Plateau, southwest Pacific Ocean, *N. Z. J. Geol. Geophys.*, *51*, 29–42.

Mortimer, N., P. B. Gans, M. Palin, S. Meffre, R. H. Herzer, and D. N. B. Skinner (2010), Location and migration of Miocene-Quaternary volcanic arcs in the SW Pacific region, *J. Volcanol. Geotherm. Res.*, doi:10.1016/j.jvolgeores.2009.1002.1017, in press.

Nathan, S., H. J. Anderson, R. A. Cook, R. H. Herzer, R. H. Hoskins, J. I. Raine, and D. Smale (1986), *Cretaceous and Cenozoic Sedimentary Basins of the West Coast Region, South Island, New Zealand*, *N. Z. Geol. Surv. Basin Stud.*, vol. 1, Dep. of Sci. and Ind. Res., Wellington.

Nelson, C. S., R. M. Briggs, and P. J. J. Kamp (1986), Nature and significance of volcanogenic deposits at the Eocene-Oligocene boundary, Hole 593, Challenger Plateau, Tasman Sea, *Initial Rep. Deep Sea Drill. Proj.*, *90*, 1175–1187.

Packham, G. H., and D. A. Falvey (1971), An hypothesis for the formation of marginal basins in the western Pacific, *Tectonophysics*, *11*(2), 79–109, doi:10.1016/0040-1951(71)90058-8.

Paris, J.-P. (1981), *Géologie de la Nouvelle-Calédonie: Un Essai de Synthèse, Mem. BRGM*, *113*, 278 pp.

Platt, J. P., J. I. Soto, M. J. Whitehouse, A. J. Hurford, and S. P. Kelley (1998), Thermal evolution, rate of exhumation, and tectonic significance of metamorphic rocks from the floor of the Alboran extensional basin, western Mediterranean, *Tectonics*, *17*(5), 671–689, doi:10.1029/98TC02204.

Rollet, N., J. Déverchère, M.-O. Beslier, P. Guennoc, J.-P. Réhault, M. Sosson, and C. Truffert (2002), Back arc extension, tectonic inheritance, and volcanism in the Ligurian Sea, western Mediterranean, *Tectonics*, *21*(3), 1015, doi:10.1029/2001TC900027.

Schellart, W. P., G. S. Lister, and V. G. Toy (2006), A Late Cretaceous and Cenozoic reconstruction of the southwest Pacific region: Tectonics controlled by subduction and slab rollback processes, *Earth Sci. Rev.*, *76*(3–4), 191–233, doi:10.1016/j.earscirev.2006.01.002.

Schellart, W. P., B. L. N. Kennett, W. Spakman, and M. Amaru (2009), Plate reconstructions and tomography reveal a fossil lower mantle slab below the Tasman Sea, *Earth Planet. Sci. Lett.*, *278*(3–4), 143–151, doi:10.1016/j.epsl.2008.11.004.

Sclater, J. G., and P. A. F. Christie (1980), Continental stretching: An explanation of the post-Mid-Cretaceous subsidence of the central North Sea Basin, *J. Geophys. Res.*, *85*(B7), 3711–3739, doi:10.1029/JB085iB07p03711.

Sdrolias, M., R. D. Müller, A. Mauffret, and G. Bernardel (2004), Enigmatic formation of the Norfolk Basin, SW Pacific: A plume influence on back-arc extension, *Geochem. Geophys. Geosyst.*, *5*, Q06005, doi:10.1029/2003GC000643.

Seber, D., M. Barazangi, A. Ibenbrahim, and A. Demnati (1996), Geophysical evidence for lithospheric delamination beneath the Alboran Sea and Rif-Betic mountains, *Nature*, *379*, 785–790, doi:10.1038/379785a0.

Shaw, M., and R. Pysklywec (2007), Anomalous uplift of the Apennines and subsidence of the Adriatic: The result of active mantle flow?, *Geophys. Res. Lett.*, *34*, L04311, doi:10.1029/2006GL028337.

Shipboard Scientific Party (1973a), Site 207, *Initial Rep. Deep Sea Drill. Proj.*, *21*, 197–214.

Shipboard Scientific Party (1973b), Site 206, *Initial Rep. Deep Sea Drill. Proj.*, *21*, 103–125.

Shipboard Scientific Party (1973c), Site 208, *Initial Rep. Deep Sea Drill. Proj.*, *21*, 271–281.

Stagpoole, V., and A. Nicol (2008), Regional structure and kinematic history of a large subduction back thrust: Taranaki Fault, New Zealand, *J. Geophys. Res.*, *113*, B01403, doi:10.1029/2007JB005170.

Steinberger, B., R. Sutherland, and R. J. O'Connell (2004), Prediction of Emperor-Hawaii seamount locations from a revised model of global plate motion and mantle flow, *Nature*, *430*, 167–173, doi:10.1038/nature02660.

Stern, T. A., and W. E. Holt (1994), Platform subsidence behind an active subduction zone, *Nature*, *368*, 233–236, doi:10.1038/368233a0.

Sutherland, R. (1995), The Australia-Pacific boundary and Cenozoic plate motions in the SW Pacific: Some constraints from Geosat data, *Tectonics*, *14*(4), 819–831, doi:10.1029/95TC00930.

Sutherland, R. (1999), Basement geology and tectonic development of the greater New Zealand region: An interpretation from regional magnetic data, *Tectonophysics*, *308*(3), 341–362, doi:10.1016/S0040-1951(99)00108-0.

Tulloch, A. J., J. Ramezani, N. Mortimer, J. Mortensen, P. van den Bogaard, and R. Maas (2009), Mid-Cretaceous felsic volcanism in New Zealand and Lord Howe Rise (Zealandia) as a precursor to continental breakup, in *Extending a Continent: Architecture, Rheology and Heat Budget*, edited by U. Ring and B. Wernicke, *Geol. Soc. Spec. Publ.*, *321*, 89–118, doi:10.1144/SP321.5.

Turnbull, I. M., and C. I. Uruski (1993), *Cretaceous and Cenozoic Sedimentary Basins of Western Southland, South Island, New Zealand*, Inst. of Geol. and Nucl. Sci., Wellington.

Turner, S. P., J. P. Platt, R. M. M. George, S. P. Kelley, D. G. Pearson, and G. M. Nowell (1999), Magmatism associated with orogenic collapse of the Betic-Alboran Domain, SE Spain, *J. Petrol.*, *40*(6), 1011–1036, doi:10.1093/petrology/40.6.1011.

Uruski, C. I. (2008), Deepwater Taranaki Basin, New Zealand: Structural development and petroleum potential, *Explor. Geophys.*, *39*, 94–107, doi:10.1071/EG08013.

Uruski, C. I., and P. Baillie (2002), Petroleum systems of the Deepwater Taranaki Basin, New Zealand, in *2002 New Zealand Petroleum Conference Proceedings*, pp. 402–407, Minist. of Econ. Dev., Wellington.

Uruski, C., and R. Wood (1991), A new look at the New Caledonia Basin, an extension of the Taranaki Basin, offshore North Island, New Zealand, *Mar. Pet. Geol.*, *8*, 379–391, doi:10.1016/0264-8172(91)90061-5.

Uruski, C. I., P. Baillie, and V. M. Stagpoole (2003), Development of the Taranaki Basin and comparisons with the Gippsland Basin: Implications for deepwater exploration, *APPEA J.*, *43*(1), 185–196.

van der Lingen, G. J. (1973), The Lord Howe Rise rhyolites, *Initial Rep. Deep Sea Drill. Proj.*, *21*, 523–540.

Weissel, J. K., and D. E. Hayes (1977), Evolution of the Tasman Sea reappraised, *Earth Planet. Sci. Lett.*, *36*(1), 77–84, doi:10.1016/0012-821X(77)90189-3.

Wood, R. A. (1993), The Challenger Plateau, in *Sedimentary Basins of the World*, vol. 2, *South Pacific Sedimentary Basins*, edited by P. F. Ballance, pp. 351–364, Elsevier, Amsterdam.

Wood, R., and D. Woodward (2002), Sediment thickness and crustal structure of offshore western New

Zealand from 3D gravity modelling, *N. Z. J. Geol. Geophys.*, *45*, 243–255.

Woodward, D., and T. M. Hunt (1971), Crustal structure across the Tasman Sea, *N. Z. J. Geol. Geophys.*, *14*, 39–45.

Wortel, M. J. R., and W. Spakman (2000), Subduction and slab detachment in the Mediterranean-Carpathian region, *Science*, *290*(5498), 1910–1917, doi:10.1126/science.290.5498.1910.

Zandt, G., H. Gilbert, T. J. Owens, M. Ducea, J. Saleeby, and C. H. Jones (2004), Active foundering of a continental arc root beneath the southern Sierra Nevada in California, *Nature*, *431*, 41–46, doi:10.1038/nature02847.

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Figure S1

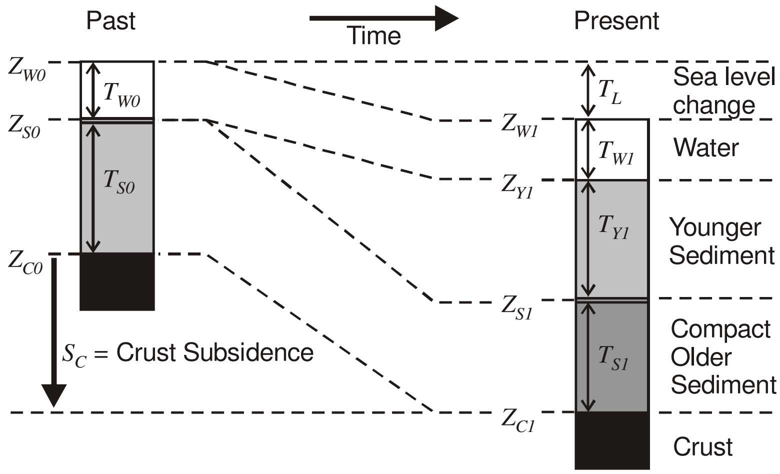

Figure S1. Definition of terms used in subsidence calculations.

$$S_C = T_L + T_{W1} + T_{Y1} + T_{S1} - T_{W0} - T_{S0}$$

See text S1 for description of terms.

Table S1. Calculations for line TL01 (Fig. 3)

Input:	**Rakopi Formation (Late Cretaceous)**				**Top Eocene shelf break**	
Seabed (s twt)	2.0	2.0	2.2	2.2	2.4	2.4
Layer of interest (s twt)	4.5	4.5	4.5	4.5	3.2	3.4
Basement (s twt)	5.6	6.5	5.6	6.5	5.5	6.5
Paleo-water-depth of layer (m)	0.0	0.0	0.0	0.0	100.0	100.0
Sea level at time of deposition, TL	200.0	200.0	200.0	200.0	150.0	150.0
Output:						
Water depth (m)	**1490.0**	**1490.0**	**1639.0**	**1639.0**	**1788.0**	**1788.0**
Base Layer 2 depth beneath seabed (ms one-way-time)	1250.0	1250.0	1150.0	1150.0	400.0	500.0
Base Layer 2 depth beneath seabed, TY1 (m)	**2335.0**	**2335.0**	**1973.2**	**1973.2**	**220.3**	**356.0**
Base Layer 1 below seabed (ms one-way-time)	1800.0	2250.0	1700.0	2150.0	1550.0	2050.0
Base Layer 1 (basement) beneath seabed, TY1+TS1 (m)	**4864.5**	**7612.6**	**4336.7**	**6949.1**	**3601.7**	**6315.8**
Layer 2 (young sediment) is made up of						
Thickness, TY1 (m)	2335.0	2335.0	1973.2	1973.2	220.3	356.0
Integrated pore thickness (m), PY1	785.3	785.3	715.0	715.0	118.9	185.9
Integrated grain thickness (m), GY1	1549.7	1549.7	1258.2	1258.2	101.4	170.1
Mean sediment density, kg/m3	1725.6	1725.6	1657.9	1657.9	1196.7	1242.4
Layer 1 (old sediment) is made up of heights						
Thickness, TS1 (m)	2529.5	5277.5	2363.6	4976.0	3381.4	5959.9
Integrated pore thickness (m), PS1	254.6	329.4	294.7	389.7	832.8	905.7
Integrated grain thickness (m), GS1	2274.9	4948.2	2068.9	4586.2	2548.6	5054.2
Mean sediment density, kg/m3	2338.3	2437.7	2275.9	2396.4	1959.6	2204.9
Remove sediment layer 2						
Decompacted sediment thickness of layer 1, **TS0** (m)	**3182.7**	**6032.3**	**2947.8**	**5658.9**	**3489.4**	**6141.3**
Integrated pore thickness (m), PS0	907.8	1084.2	878.9	1072.7	940.8	1087.1
Integrated grain thickness (m), GS0	2274.9	4948.2	2068.9	4586.2	2548.6	5054.2
Compaction subsidence, Sd (m)	653.3	754.8	584.2	683.0	108.0	181.5
Crustal subsidence, Sc (m)	**3371.8**	**3270.2**	**3227.9**	**3129.2**	**1950.2**	**2012.5**
Isostatic correction C (m)	1071.8	1071.8	870.2	870.2	70.1	117.6
Tectonic subsidence, St (m)	**2299.9**	**2198.4**	**2357.7**	**2259.0**	**1880.1**	**1894.9**

For Paleocene coastal sandstones in the Waka Nui-1 well, we determine a tectonic subsidence value of 2200 m.

Tectonic subsidence values for the Lord Howe Rise and West Norfolk Ridge are computed assuming negligible compaction of material underlying the erosional unconformity.

Text S1

See Fig. S1 for illustration and definition of terms used in the subsidence calculation. The crustal subsidence, S_C , since some earlier time is given by the sum of the sea level change since that time (T_L), present thicknesses of water (T_{W1}), young sediment deposited since the earlier time (T_{Y1}), and the compacted older sediment (T_{S1}); minus the water depth at the previous time (T_{W0}) and the decompacted thickness of older sediment (T_{S0}).

$$S_C = T_L + T_{W1} + T_{Y1} + T_{S1} - T_{W0} - T_{S0} \tag{1}$$

We assume an exponential decrease in porosity with depth D beneath the seabed:

$$\phi(D) = \phi_0 \exp(-cD) \tag{2}$$

Where ϕ_0 is the porosity of sediment at the seabed and c is a depth-decay constant that depends upon sediment type.

We use compaction parameters for muddy sediments (ϕ_0 = 0.57, 1/c = 2000 m), as determined from wells in Taranaki Basin [*Funnell et al.*, 1996]. We convert between two-way time (s twt) on seismic sections and depth (m) using well control from the Great South Basin [*Cook et al.*, 1999], which mainly lies outside the region of Cenozoic tectonics and in similar water depths (200-2000 m).

The 'integrated pore thickness' for a sedimentary unit between depths D_1 and D_2 is

$$P = \int_{D_1}^{D_2} \phi_0 \exp(-cD) dD = \frac{\phi_0}{c}\left(\exp(-cD_1) - \exp(-cD_2)\right) \tag{3}$$

and hence the 'integrated grain thickness' is

$$G = D_2 - D_1 - P \tag{4}$$

The decompacted thickness of a unit is found numerically by assuming that the integrated grain thickness of the unit remains the same, but the top of the unit is now at the seabed, and the decrease in porosity with the revised depth obeys equation 2.

We define tectonic subsidence, S_T , as the subsidence at the top of the crust that would occur if no younger sediment load were present, and subsidence occurred entirely under water.

We define an isostatic correction that is equivalent to the isostatic rebound that would occur if the load of younger sediment was removed and the crust has insignificant strength at the length scale of interest.

$$S_T = S_C - C \tag{5}$$

Any change in eustatic sea level [*Haq et al.*, 1987] is accounted for in the crustal subsidence determination (equation 1), but we do not make an additional isostatic correction, because the isostatic response is experienced across the global ocean and is hence generally already included in global records of sea level change.

By isostatic balance:

$$(G_{Y1} - C)(\rho_m - \rho_w) = G_{Y1}(\rho_m - \rho_c)$$

$$C = G_{Y1} \frac{(\rho_c - \rho_w)}{(\rho_m - \rho_w)} \qquad (6)$$

where ρ_m, ρ_c, ρ_w are the densities of upper mantle, sediment grains, and sea water respectively; and G_{Y1} is the integrated grain thicknesses of the younger sediment load.

References

Cook, R. A., et al. (1999), *Cretaceous-Cenozoic geology and petroleum systems of the Great South Basin, New Zealand*, 188 pp., Institute of Geological and Nuclear Sciences Limited, Lower Hutt, New Zealand.

Funnell, R., et al. (1996), Thermal state of the Taranaki Basin, New Zealand, *Journal Of Geophysical Research-Solid Earth*, *101*(B11), 25197-25215.

Haq, B. U., et al. (1987), Chronology of fluctuating sea levels since the Triassic, *Science*, *235*(4793), 1156-1167.