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Hikurangi Plateau: Crustal structure, rifted formation, and Gondwana subduction history

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[1] Seismic reflection profiles across the Hikurangi Plateau Large Igneous Province and adjacent margins reveal the faulted volcanic basement and overlying Mesozoic-Cenozoic sedimentary units as well as the structure of the paleoconvergent Gondwana margin at the southern plateau limit. The Hikurangi Plateau crust can be traced 50–100 km southward beneath the Chatham Rise where subduction cessation timing and geometry are interpreted to be variable along the margin. A model fit of the Hikurangi Plateau back against the Manihiki Plateau aligns the Manihiki Scarp with the eastern margin of the Rekohu Embayment. Extensional and rotated block faults which formed during the breakup of the combined Manihiki-Hikurangi plateau are interpreted in seismic sections of the Hikurangi Plateau basement. Guyots and ridge-like seamounts which are widely scattered across the Hikurangi Plateau are interpreted to have formed at 99–89 Ma immediately following Hikurangi Plateau jamming of the Gondwana convergent margin at ∼100 Ma. Volcanism from this period cannot be separately resolved in the seismic reflection data from basement volcanism; hence seamount formation during Manihiki-Hikurangi Plateau emplacement and breakup (125–120 Ma) cannot be ruled out. Seismic reflection data and gravity modeling suggest the 20-Ma-old Hikurangi Plateau choked the Cretaceous Gondwana convergent margin within 5 Ma of entry. Subsequent uplift of the Chatham Rise and slab detachment has led to the deposition of a Mesozoic sedimentary unit that thins from ∼1 km thickness northward across the plateau. The contrast with the present Hikurangi Plateau subduction beneath North Island, New Zealand, suggests a possible buoyancy cutoff range for LIP subduction consistent with earlier modeling.

1. Introduction

[2] The Hikurangi Plateau is a Large Igneous Province (LIP) [*Mortimer and Parkinson*, 1996; *Mahoney and Coffin*, 1997; *Hoernle et al.*, 2003], one of the many Early Cretaceous LIPs that are found in the Pacific [*Coffin and Eldholm*, 1993] (Figure 1). The Hikurangi Plateau (Figure 1) has been recently interpreted as part of a much larger (~3.6 × 10⁶ km²), ~120 Ma, combined Ontong-Java/Manihiki/Hikurangi Plateau (OJMHP) [*Taylor*, 2006; K. Hoernle et al., The Hikurangi Large Igneous Province: Two major Cretaceous volcanic events, manuscript in preparation, 2008]. The resultant superplateau would have formed the largest known oceanic plateau on Earth; however the origin of such a voluminous eruption of magma over an apparently short interval (a few million years) remains enigmatic [*Taylor*, 2006; *Ingle and Coffin*, 2004; *Fitton et al.*, 2004; *Tarduno et al.*, 1991].

[3] The Hikurangi Plateau includes numerous large seamounts and other volcanic features, one of the longest submarine channels in the world, the Hikurangi Channel [*Lewis*, 1994], thick sediments along the southern, eastern and western margins, and a steep, faulted boundary in the north (Figure 2). Average depth of the plateau is 2,500–4,000 m, similar to the depth range of the Manihiki Plateau. In contrast the Ontong Java plateau lies at depths of 1500–4000 m. The Hikurangi Plateau is interpreted to have rifted apart from the OJMHP at ~120 Ma and drifted south from the Manihiki Plateau as a result of seafloor spreading at the Osbourn Trough [*Lonsdale*, 1997; *Billen and Stock*, 2000; *Downey et al.*, 2007] (Figure 1). Basement rocks of the Hikurangi Plateau dredged from the base of the Rapuhia Scarp, the northern margin of the plateau, have geochemical compositions similar to the Ontong Java Plateau [*Mortimer and Parkinson*, 1996; *Hoernle et al.*, 2005]. Late Cretaceous sediments have been dredged from seamounts on the plateau [*Strong*, 1994] and were also drilled near the northeast margin of the Hikurangi Plateau at ODP site 1124 (Figures 2 and 3). Rocks dredged by R/V *Sonne* survey SO168 [*Hoernle et al.*, 2003] from the northern plateau basement are tholeiitic basaltic to gabbroic samples 94–118 Ma in age (Hoernle et al., manuscript in preparation, 2008). The geochemistry, flat chondrite-normalized REE patterns and enriched mantle (EM)-type isotopic compositions, and the age distribution of these rocks are very similar to that of the basement rocks from Manihiki and Ontong-Java plateaus [*Hoernle et al., 2005*]. In the Late Cretaceous the Hikurangi plateau has entered, and been partially subducted beneath, the interpreted Gondwana subduction margin along the north Chatham Rise. The buoyant 20 Ma old plateau choked the subduction system leading to the cessation of subduction along this section of the margin [*Bradshaw*, 1989; *Davy*, 1992; *Wood and Davy*, 1994].

[4] Early reconnaissance seismic reflection surveying of the Hikurangi Plateau generally involved low-fold seismic reflection surveying with subseafloor penetration of 1–2 s Two-Way Travel time (TWT) [*Wood and Davy*, 1994]. Deep crustal seismic reflection surveying undertaken in 2001 as part of New Zealand's UNCLOS (United Nations Commission for the Law of the Sea) continental shelf delineation program, combined with dredge and swath bathymetry data collected by the 2002 SO168 survey by R/V *Sonne* [*Hoernle et al.*, 2003, 2004], has revealed much of the crustal structure detailing the formation, rift, drift and subduction history of the plateau. This paper focuses on the interpretation of these data, and in particular deep crustal seismic line HKDC1, which traverses 1000 km from the ocean crust north of the plateau to the crest of the Chatham Rise to the south of the plateau (Figure 1).

[5] Examination of line HKDC1, as well as sections of lines less than 300 km further east, enables the following:

[6] 1. Rift structures associated with Manihiki-Hikurangi Plateau breakup ~120 Ma are recognized.

[7] 2. The fossil Gondwana convergent margin including the fossil accretionary prism and subducted Hikurangi Plateau basement is identified (interpreted from seismic data) beneath the central-northern Chatham Rise. Along HKDC1 this basement extends ~100 km south beneath the Chatham Rise from the toe of the accretionary prism to an interpreted leading southern edge of the plateau.

[8] 3. The widespread volcanism within the plateau including an interpreted ~0.5–1.5 km thick volcaniclastic/limestone/chert layer overlying basaltic basement is interpreted. The favored model for plateau volcanism involves guyot formation ~99–89 Ma, a period which immediately follows the cessation of Mesozoic subduction beneath the Chatham Rise. An alternative possibility, that the guyots formed soon after plateau formation and breakup at ~120 Ma with later 89–99 Ma volca-
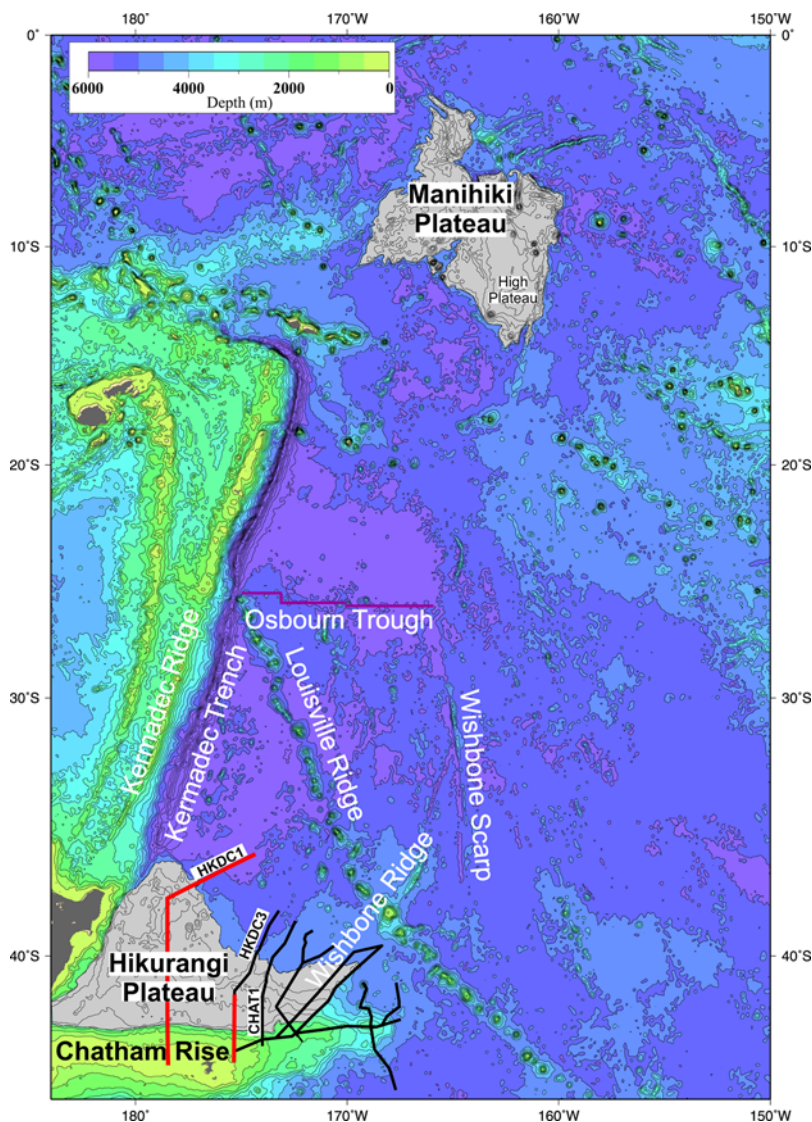
Figure 1. Bathymetry of the Hikurangi Plateau–Manihiki Plateau region of the SW Pacific [from *Smith and Sandwell*, 1997]. Bathymetry contour interval is 500 m. Water depths shallower than 4500 m on the Hikurangi and Manihiki plateaus have been shaded gray. Deep crustal seismic reflection lines collected across the Hikurangi Plateau in 2001 by *Geco Resolution* are indicated by black and red lines. Thickened red sections on HKDC1 and HKDC3 are shown in Figures 4–10.

nism erupting on top of existing peaks, cannot however be discounted.

[9] *Wood and Davy* [1994] noted the difference in crustal structure in the Hikurangi Plateau across the 176°W longitude (e.g., dashed line in Figure 2). East of 176°W crustal and sedimentary structure of the plateau has a pervasive northeast orientation and appears closely linked to the nature of the interpreted Late Cretaceous Wishbone Ridge lineation extending northeast from the Chatham Rise (Figures 1 and 3) [*Davy*, 2004]. The Wishbone Ridge has been interpreted by *Davy* [2004] as being a crustal boundary involved in the breakup

of the New Zealand sector of the Gondwana subduction margin. Using the interpretation of seismic line HKDC1 and others, this paper will also focus on the earlier history of the Hikurangi Plateau formation; the interpreted rifting apart from a larger LIP, the transport of the plateau southward and eventual subduction beneath the Gondwana convergent margin.

2. Hikurangi Plateau Structure

[10] The crust of the Hikurangi Plateau has been estimated, on the basis of gravity models, to be
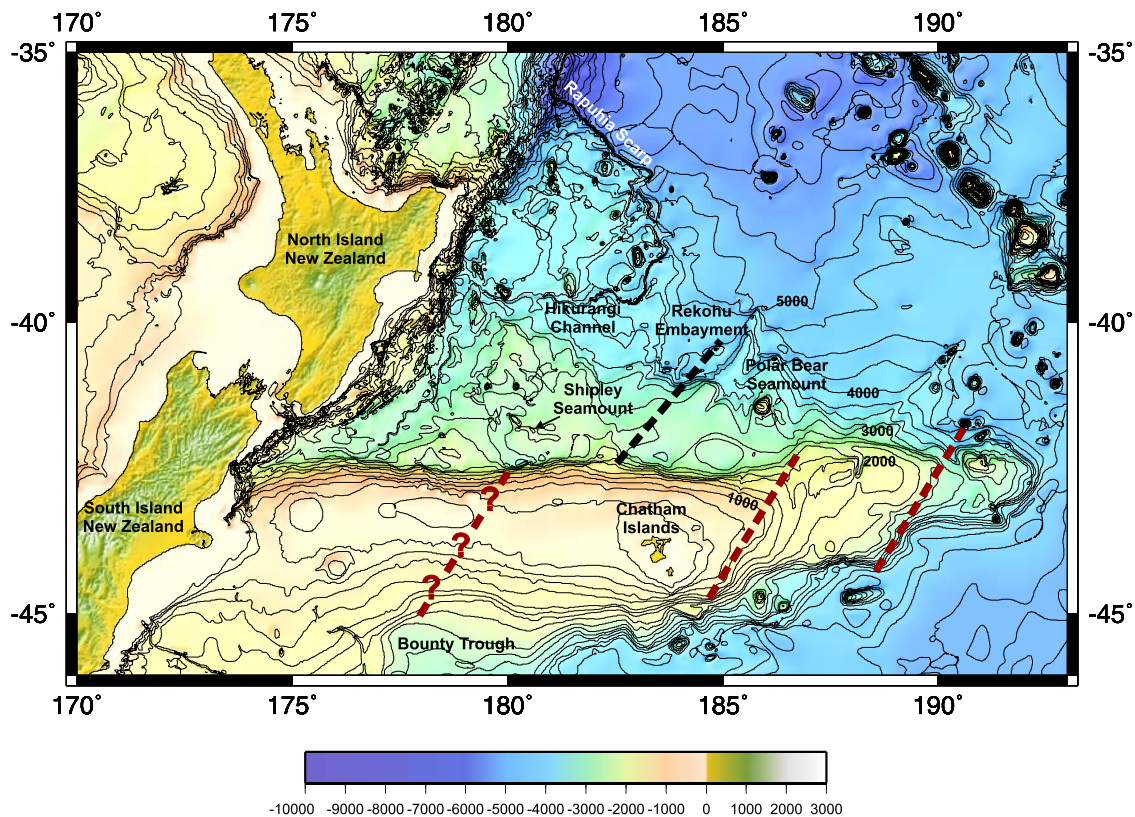
Figure 2. Bathymetry of the Hikurangi Plateau from *Charting Around New Zealand Group* (*CANZ*) [1997]. Bathymetry contour interval is 250 m. Red dashed lines mark postulated segmentation of the Chatham Rise into three (possibly four) sections. Black dashed line marks boundary between southeastern plateau dominated by northeastern volcanic grain and the rest of the Hikurangi Plateau.

12–15 km thick, thicker than the 4–9 km thick oceanic crust to the northeast [*Davy*, 1992; *Davy and Wood*, 1994]. Where it underlies the Chatham Rise in the south the total crustal thickness was modeled to be ~25 km thick. While it was not possible in these 1992–1994 models to subdivide the Chatham Rise into the subducted Hikurangi Plateau and the overlying Gondwana margin crust lines HKDC1 and HKDC3 now make such model refinement possible (see later).

[11] Seamounts scattered across the Hikurangi Plateau are associated with 15–20 large (~20 km diameter, 40–80 mgal) discrete gravity anomaly highs (Figure 3). The northern two thirds of the Northern Volcanic Region are dominated by three northeast alignments of gravity highs, with each alignment composed of three gravity highs (seamounts). A fourth triplet of gravity highs lies within the eastern Rekohu Embayment. *Hoernle et al.* [2003] categorized the seamounts of the Hikurangi Plateau into (1) ridge-like seamounts and (2) massive (diameter up to 25 km) guyot seamounts. The ridge-like seamounts west of

~176°W are generally within 100 km of the northern plateau margin and have long axes subparallel to the margin (Figure 3). East of 176°W the ridge-like volcanics are all northeast oriented subparallel to the trend of the Wishbone Ridge [*Davy*, 2004].The massive guyots dredged from the southern, southeastern, northern and central Hikurangi Plateau have flat tops interpreted as wave-cut surfaces formed before subsidence of ~1500–3000 m. The numerous volcanic guyots within the Hikurangi Plateau contrast with the Ontong-Java Plateau which is largely free of high-relief volcanism and the Manihiki Plateau, which, although similarly rich in seamounts [*Winterer et al.*, 1974] has no guyots. Dating of rocks dredged from the Hikurangi Plateau guyots (Hoernle et al., manuscript in preparation, 2008) reveals rocks erupted between 99 and 89 Ma. In contrast to the Hikurangi Plateau basement rocks, the above rocks dredged from Hikurangi Plateau seamounts display light REE enrichment and have nearly identical HIMU (high time-integrated U/Pb)-like incompatible element and isotopic compositions to alkalic volca-
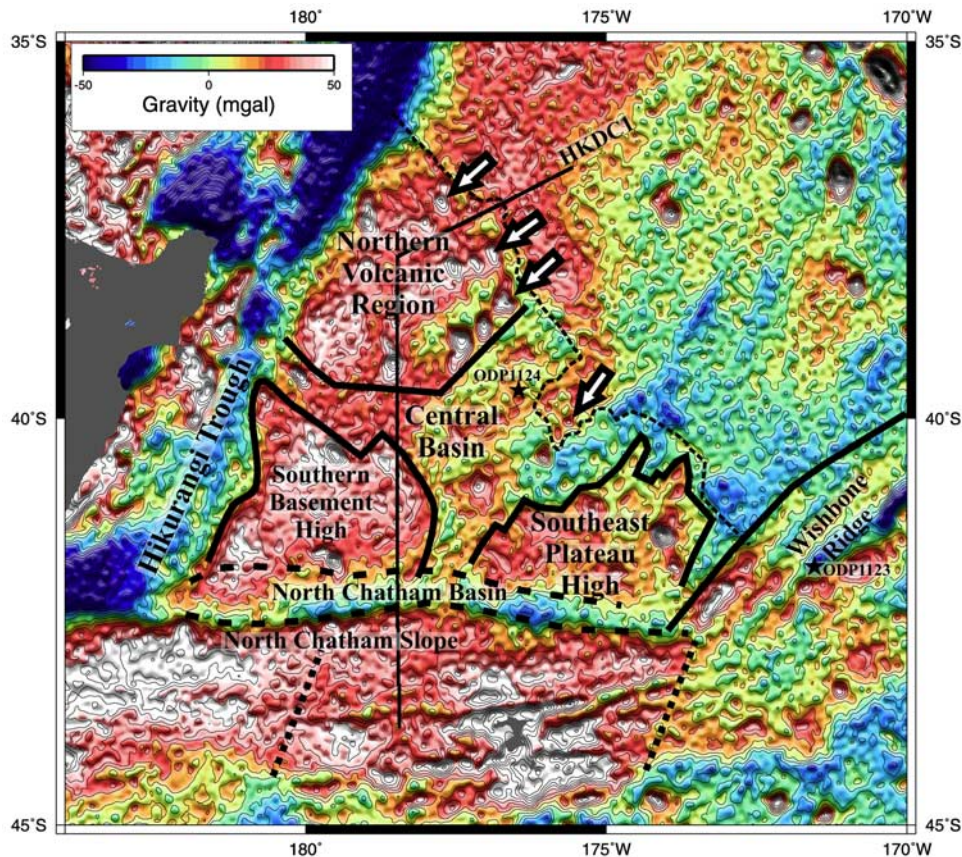
Figure 3. Hikurangi Plateau provinces [*Wood and Davy*, 1994] outlined and labeled on top of satellite gravity data [from *Sandwell and Smith*, 1997]. Contour interval is 10 mgal. ODP holes 1123 and 1124 locations are marked by stars. The northeastern boundary of the Hikurangi Plateau is plotted as a fine dashed line. Northeast alignments of volcanic peaks are highlighted by arrows. Short dashed lines highlight major segmentation in gravity lineation along the Chatham Rise.

nism of similar age on the South Island of New Zealand (Hoernle et al., manuscript in preparation, 2008).

[12] *Wood and Davy* [1994] subdivided the Hikurangi Plateau into five major regions: Northern Volcanic Region, Central Basin, Southern Basement High, North Chatham Basin, and the Hikurangi Trough. Figure 3 includes a slightly modified definition of these areas, based upon more recent seismic and gravity data. The structure of the plateau revealed by line HKDC1 (Figures 4–9) is discussed within the framework of these areas.

[13] The Northern Volcanic region is bound to the northeast by the Rapuhia Scarp (Figures 2–4) which is ~1 km high in the west and almost buried by sediment east of 176°W. This southeast trending scarp, which marks the northern boundary between the plateau and oceanic crust further north, is almost linear along its westernmost 150 km bordering the Kermadec Trench, but irregular and

convoluted further east. This more complex boundary may be related to the prolongation of the northeast aligned volcanic ridges mentioned above to the northeastern plateau margin in this region.

[14] The Southeast Plateau High and Wishbone Ridge areas are dominated by northeast trending ridge structures which are generally volcanic in appearance (Figures 2–3) [*Hoernle et al.*, 2003; *Davy*, 2004]. The ~25 km wide, 1300 m high Polar Bear Seamount (Figure 2) [*Hoernle et al.*, 2003], with a ~70 mgal relative gravity anomaly high, is the highest gravity anomaly on the plateau. It is a mixture of a large ~15 km diameter circular guyot and an abutting northeast trending ridge. Later northeast trending lineations of ~200 m high volcanic cones lie atop Polar Bear Seamount.

[15] The western arm of the Central Basin contains several seamounts with both a northwest and a north-northwest oriented structural grain. A number of similar seamounts are interpreted to lie
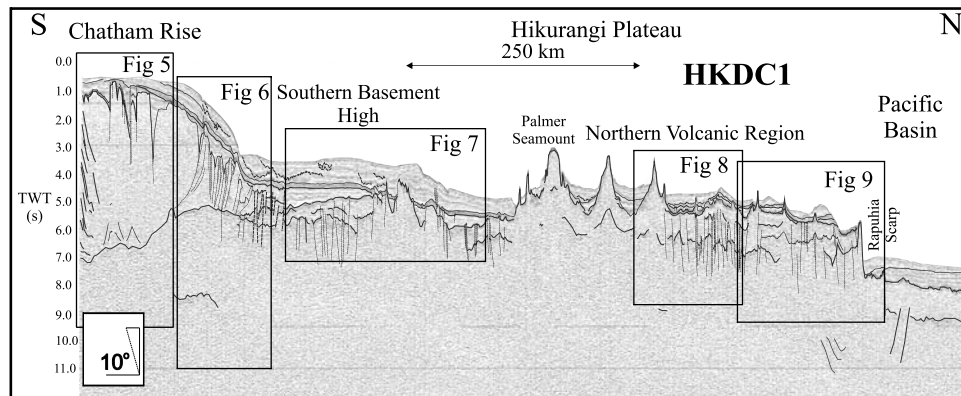
Figure 4. Seismic reflection line profile HKDC1. Location as in Figure 1. Seismic sections shown in Figures 5–9 are outlined. An expanded interpreted version of line HKDC1 is available as Figure S1. An uninterpreted version of this line is available as Figure S2. An expanded uninterpreted version of line HKDC1 is available as Figure S10. HKDC1 location and associated provinces are plotted in Figure 3. Inset shows the true slope, from horizontal, of sloping line calculated assuming sound speed of 3 km/s, i.e., lower sedimentary section/upper basement.

further west beneath the Hikurangi accretionary margin where they have been responsible for subduction erosion of the margin [*Collot et al.*, 2001]. The eastern Central Basin is both a bathymetric and gravity low region. The central deepest 100×75 km segment of this basin, the Rekohu Embayment (Figure 2), is however a ~20 mgal relative gravity high and is interpreted as an embayment of oceanic crust, with the gravity high attributable to thin crust and shallow mantle. On the few low-fold lines that cross the embayment (see section 4.2.1) the center of the embayment is cored by interpreted oceanic crust at ~7.2 s (TWT) (~5.6 km) bsl, often flanked by 1–2 km high relief on volcanic basement masking any boundary faults. Basement outside this area, within the plateau, is interpreted at ~5.2 km bsl, whereas oceanic crust basement northeast of the plateau lies at ~7.8 s (TWT) (~6.2 km) bsl.
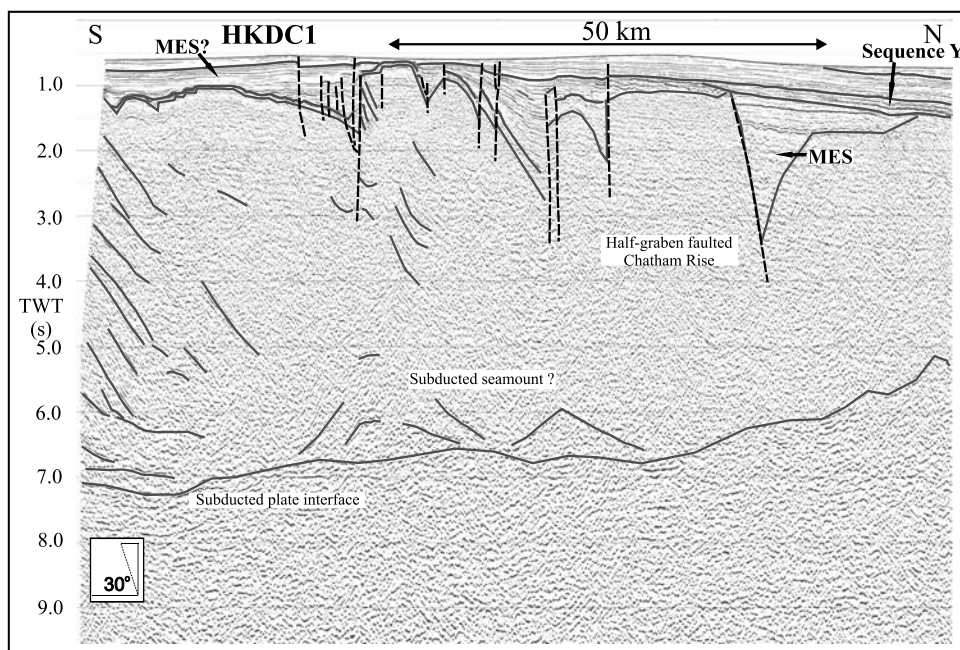

Figure 5. Seismic section on profile HKDC1. Location as in Figures 1 and 4. MES, Mesozoic sediment (100−70 Ma), Sequence Y (70−32 Ma). An uninterpreted version of this line is available as Figure S3. Inset shows the true slope, from horizontal, of sloping line calculated assuming sound speed of 3 km/s, i.e., lower sedimentary section/upper basement.
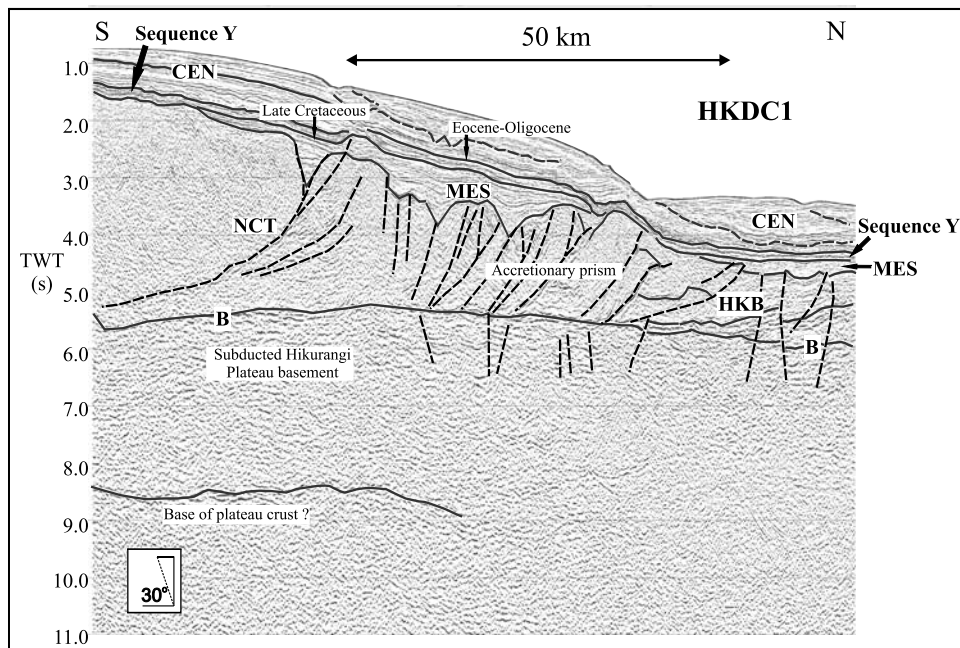
Figure 6. Seismic section on profile HKDC1. Location as in Figures 1 and 4. MES, Mesozoic sediment (100−70 Ma), Sequence Y (70–32 Ma); CEN, Cenozoic sediment (<32 Ma); NCT, North Chatham Thrust; HKB, Hikurangi Basement (volcaniclastic/limestone/chert); B, Hikurangi Basal Interface (lavas?). An uninterpreted version of this line is available as Figure S4. Inset shows the true slope as per Figure 5.

[16] The Southern Basement High is mantled by >1 km thickness of sediment, although some large seamounts protrude above the regional seafloor. Along the northern edge of this basement high area there is a strong northwest grain to the seamount structure. The few seamounts that have been mapped in detail in the southern and central part of the Southern Basement High area are more evenly circular or polygonal in aerial view (e.g.,

Shipley Seamount, Figure 2) than those mapped further north on the plateau. The North Chatham Basin is formed by sediment infill on south dipping Hikurangi Plateau basement. The plateau basement has been imaged, on several seismic reflection lines, dipping south beneath the Chatham Rise. The North Chatham Basin is interpreted as the fossil trench-slope basin associated with Cretaceous subduction of the Hikurangi Plateau at the

Figure 7. Seismic section on profile HKDC1. Location as in Figures 1 and 4. MES, Mesozoic sediment (100−70 Ma), Sequence Y (70–32 Ma); CEN, Cenozoic sediment (<32 Ma); VB, 90–100 Ma volcanics; HKB, Hikurangi Basement (volcaniclastic/limestone/chert); B, Hikurangi Basal Interface (lavas?). An uninterpreted version of this line is available as Figure S5. Inset shows the true slope as per Figure 5.

Figure 8. Seismic section on profile HKDC1. Location as in Figures 1 and 4. MES, Mesozoic sediment (100−70 Ma), Sequence Y (70−32 Ma); CEN, Cenozoic sediment (<32 Ma); VB, volcanic unit; HKB, Hikurangi Basement (volcaniclastic/limestone/chert); B, Hikurangi Basal Interface (lavas?). An uninterpreted version of this line is available as Figure S6. Inset shows the true slope as per Figure 5.

Gondwana convergent margin. On line HKDC1, and HKDC3 further east (Figure 10), the accretionary prism structure associated with this subduction system is recognizable as a series of thrust faults (Figures 4, 6, and 10).

[17] The North Chatham Slope boundary of the Chatham Rise forms a ∼1,000 km long, east-west linear feature extending from the South Island of New Zealand to the intersection of the Chatham Rise with the Wishbone Ridge (Figure 2). This
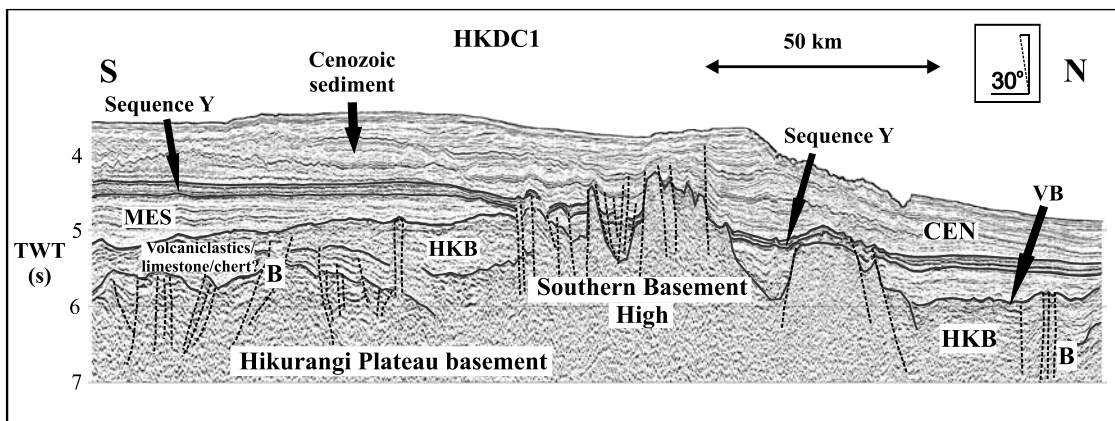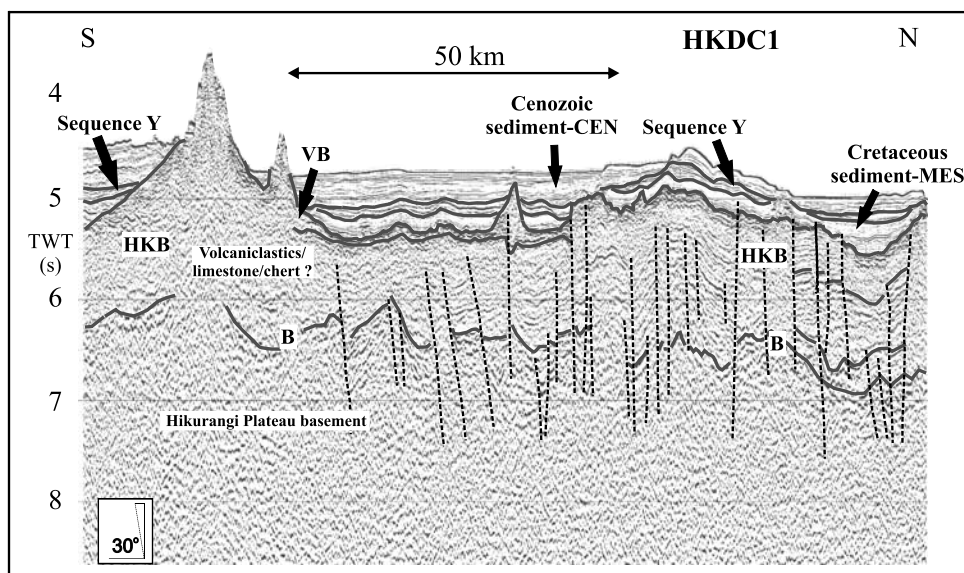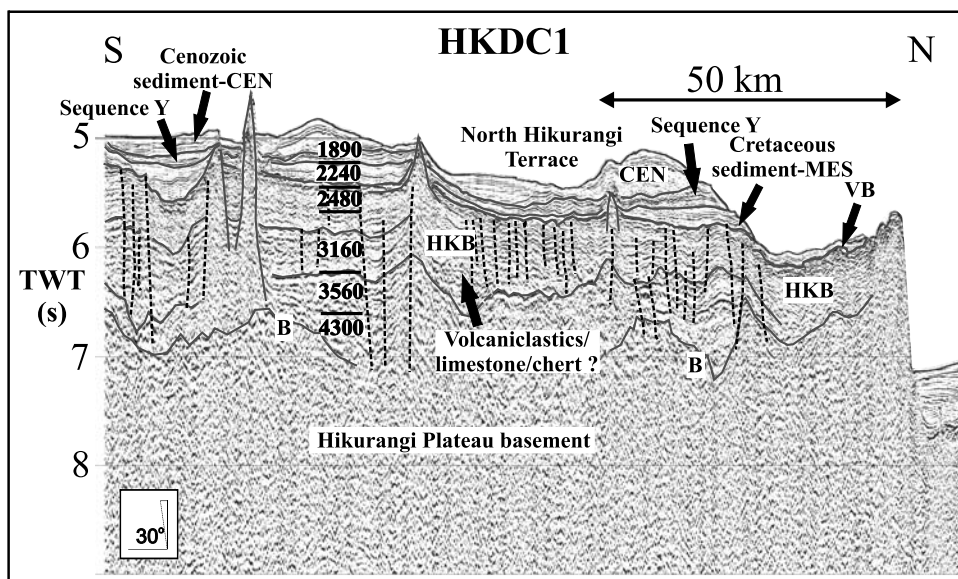

Figure 9. Seismic section on profile HKDC1. Location as in Figures 1 and 4. MES, Mesozoic sediment (100−70 Ma), Sequence Y (70−32 Ma); CEN, Cenozoic sediment (<32 Ma); VB, 90−100 Ma volcanics; HKB, Hikurangi Basement (volcaniclastic/limestone/chert); B, Hikurangi Basal Interface (lavas?). Stacking-derived interval velocities are plotted (m s^{-1}). An uninterpreted version of this line is available as Figure S7. Inset shows the true slope as per Figure 5.

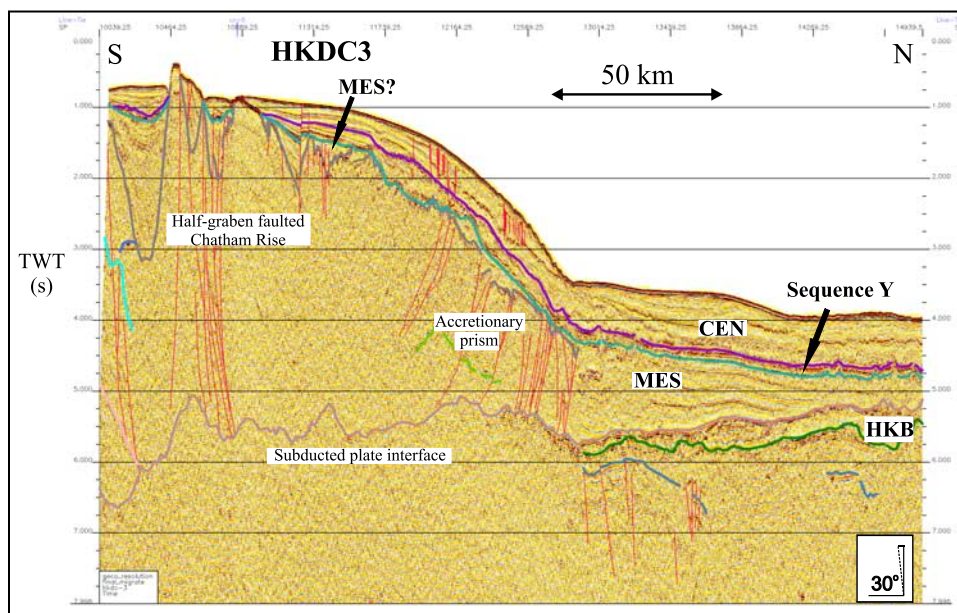**Figure 10.** Seismic section on profile HKDC3 across the Mesozoic Gondwana accretionary margin. Location as in Figure 1. MES, Mesozoic sediment (100 − 70 Ma), Sequence Y (70 − 32 Ma); CEN, Cenozoic sediment (<32 Ma); HKB, Hikurangi Basement (volcaniclastic/limestone/chert). An uninterpreted version of this line is available as Figure S8. Inset shows the true slope as per Figure 5.

margin is offset to the northeast in two steps (up to 50 km) at 174°W and 169°W. There is a similar, but less pronounced (∼10 km) northeast step at ∼180°. Within the Chatham Rise the satellite gravity data (Figure 3) reveal several ∼E−W trending gravity lineations extending up to 500 km and inferred to be caused by basement horst and graben structures. These grabens are interpreted to have been formed during regional Cretaceous extension prior to New Zealand−Antarctic rifting and seafloor spreading [*Wood and Herzer*, 1993]. The orientation of these lineations will relate to the relative plate motions at the time of formation or to inherited basement grain formed by accretion associated with subduction at the Gondwana margin. The intersection of some of the gravity lineations with the Wishbone Ridge raises the possibility that motion on the interpreted transform Wishbone Ridge plate boundary may have translated into extension within the Chatham Rise [*Davy*, 2006a].

3. Mesozoic Chatham Rise Subduction

[18] The southern boundary of the Hikurangi Plateau with the Chatham Rise is interpreted as a fossil subduction zone. The historical Cretaceous subduction of the plateau southward beneath the Chatham Rise provides a potential window as to the fate of present-day subduction beneath the North Island and an informative contrast with the

modern-day subduction of the Ontong-Java Plateau in the North Solomon Trench [*Phinney et al.*, 2004]. *Bradshaw* [1989], *Wood and Davy* [1994] and *Davy* [2001] suggested that the cessation of Cretaceous subduction at the Gondwana margin, along the north Chatham Rise, may have been associated with the subduction of the Hikurangi Plateau LIP in Cretaceous times. The present subduction of the Hikurangi Plateau beneath the North Island is at the lower end of the normal plate convergence velocity and the upper end of the crustal buoyancy observed in subduction systems worldwide [*Smith et al.*, 1989; *Davy*, 1992]. It is anticipated that a modest decrease in "plate convergence" velocity normal to the subduction margin, or a modest increase in plateau crustal thickness, might make subduction of such a plateau dynamically unfavorable. Indeed, it may be that Hikurangi Plateau subduction beneath the North Island will stall prior to the plateau reaching a depth where significant conversion of the basaltic/ gabbroic crust to eclogite provides a density contrast to reinforce the subduction driving forces.

4. Seismic Data

4.1. Line HKDC1

[19] Line HKDC1 was collected in 2001 by the *Geco Resolution* as part of the New Zealand
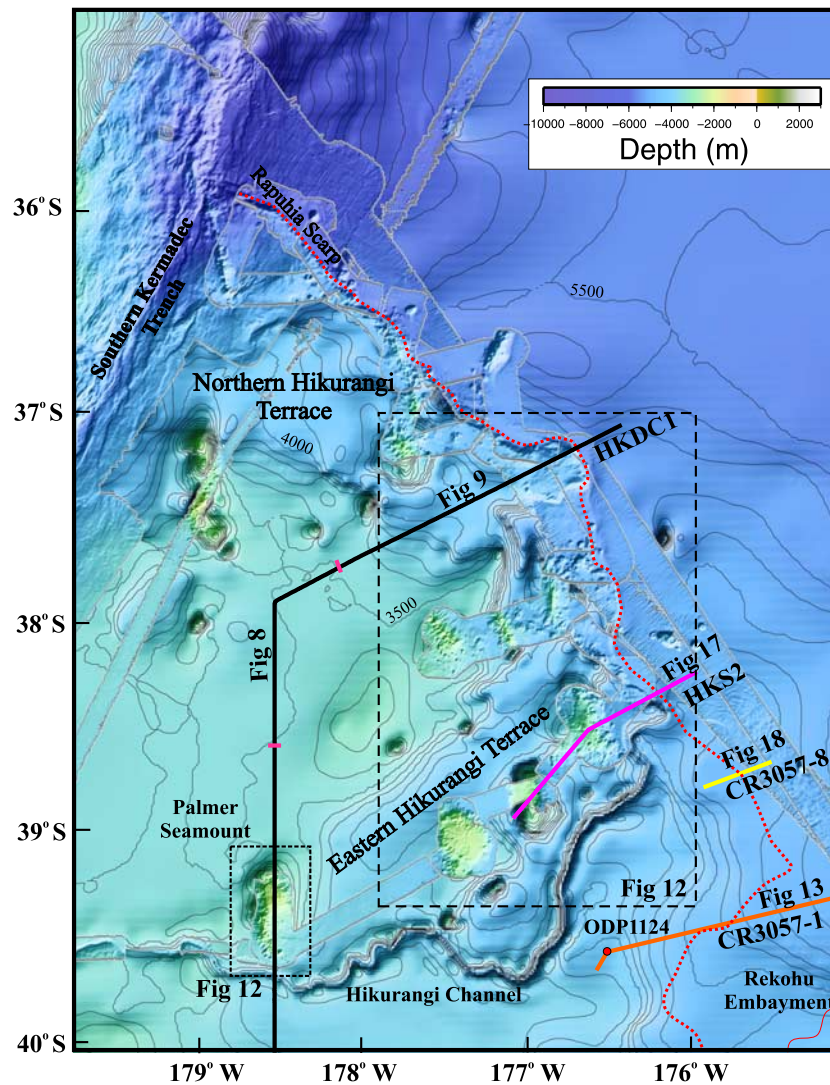
Figure 11. Bathymetry of the northern Hikurangi Plateau. Regional bathymetry is from *CANZ* [1997] with contour interval of 100 m. Overlain swath bathymetry data is from surveys by R/V *L'Atalante*, R/V *N. B. Palmer*, and R/V *Sonne*. Featured seismic section tracks are highlighted. The interpreted northeastern plateau boundary marked by red dotted line. Dashed boxes show the extent of areas plotted in Figure 12.

UNCLOS survey program. Seismic acquisition involved a 480 channel, 6 km streamer and a 134 l air gun array sound source. Shot interval was 50 m for a 16 s record sampled at a 2 ms sampling interval. Air gun source depth was 7 m and the seismic streamer was towed at 8–16 m depth depending upon sea swell.

[20] Line HKDC1 (Figures 1, 4–9, 11, and 12 and Figures S1–S10 in the auxiliary material[1]) starts midway across the crest of the Chatham Rise and extends ~1000 km northward across the North Chatham Slope, the Southern Basement High, Cen-

tral Basin and Northern Volcanic Region before descending the ~1 km high Rapuhia Scarp and traversing oceanic seafloor north of the plateau.

[21] Line HKDC1 images structure within the upper 3.0 s TWT (~6–7 km) of the crust of both the northern oceanic crust and the Hikurangi Plateau. The Hikurangi Plateau basement can be traced to a depth of 7 s TWT (~18 km) below seafloor at the southern limit of line HKDC1. The Moho at the base of the crust is not imaged unambiguously at any point on line HKDC1 although it has been tentatively interpreted at 8.5–9 s TWT beneath the Northern Chatham Rise slope.

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[1]Auxiliary materials are available in the HTML. doi:10.1029/2007GC001855.
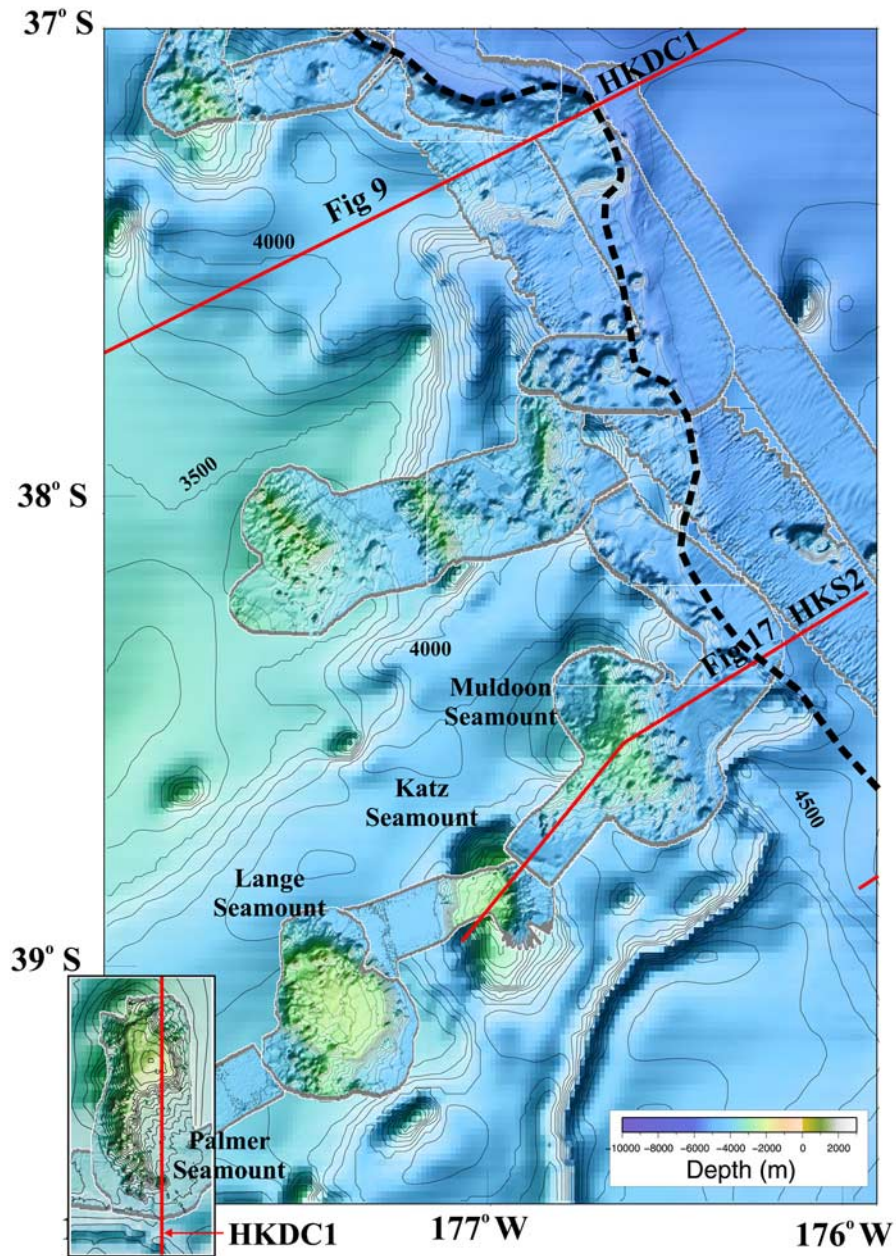
Figure 12. Bathymetry of the northeastern Hikurangi Plateau. Regional bathymetry is from *CANZ* [1997]. Swath bathymetry data is from survey SO168 by R/V *Sonne*. Contour interval is 100 m. Featured seismic section tracks are highlighted. Inset (at same scale) is of Palmer Seamount lying to the SW; its location is shown in Figure 11. The Hikurangi Plateau boundary is marked as black dashed line.

[22] Interpretation of HKDC1 (Figures 4–6), and less clearly lines HKDC3 (Figure 10) and CHAT1 (that used similar acquisition parameters) further east (Figure 1), shows basement of the Hikurangi Plateau dipping southward beneath the Chatham Rise and the development of thrust structures in the overlying sedimentary wedge. Half-grabens are well imaged within the Chatham Rise basement across the crest of the Chatham Rise. The Southern Basement High is distinguished on line HKDC1 by a ~80 km wide zone of faulted, uplifted basement along its northern edge. Within the Northern Volcanic Region, a series of 20- to 30-km-wide volcanoes are apparent on line HKDC1. These volcanoes, imaged on HKDC1 and adjacent seismic lines, rise up to 2.5 km above volcanic basement (unit HKB, Figures 8 and 9), which is ~700 m shallower than the equivalent basement surface within the Central Basin region.

Figure 13. Seismic section on profile CR3057-1. Location as in Figure 11. Sequence Y (70–32 Ma). HKB, Hikurangi Basement (volcaniclastic/limestone/chert); B, Hikurangi Basal Interface (lavas?). The 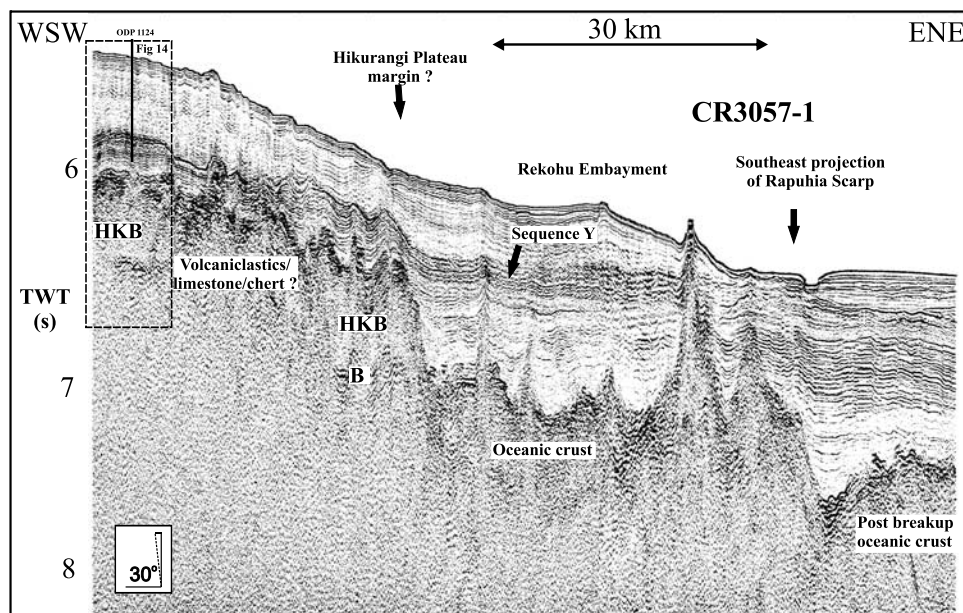location and depth extent of ODP hole 1124 is shown. Seismic data provided by the National Institute of Water and Atmospheric Research (NIWA). Inset shows the true slope as per Figure 5.

This upper surface of unit HKB is the maximum limit of penetration, or "basement," achieved by the low-fold seismic surveys of *Wood and Davy* [1994]. Line HKDC1 reveals a further ∼1.5 s TWT (∼2 km) unit of coherent, but heavily faulted, reflectors (unit HKB) underlying this surface.

4.2. Sedimentary Stratigraphy

4.2.1. Sequence Y: 70−32 Ma Sedimentary Unit

[23] Seismic reflection horizons on HKDC1 and adjacent *Geco Resolution* lines (Figure 1) can be correlated with Hikurangi Plateau horizons interpreted by *Wood and Davy* [1994] and drilling results from ODP leg 181, holes 1123 and 1124 (Figures 3, 11, 13, and 14) [*Carter et al.*, 1999]. The interpreted units are summarized in Figure 15. The most prominent, widely recognizable, continuous high-amplitude reflection sequence "Sequence Y" [*Wood and Davy*, 1994] is highlighted on Figures 4–10 and 13–15. On the basis of correlation with sediments on the Chatham Rise, *Wood and Davy* [1994] suggested Sequence Y, generally 100–200 ms thick, was a condensed sequence which included most of the Paleogene with Late Cretaceous at its base. Hole 1124, ODP leg 181 (Figures 11, 13, and 14) confirmed the above interpretation.

[24] At site 1124 Sequence Y is early Oligocene to Late Cretaceous nannofossil chalks alternating with mudstones and includes several unconformities [*Carter et al.*, 1999] partly explaining the condensed nature of the sequence. The most wide-
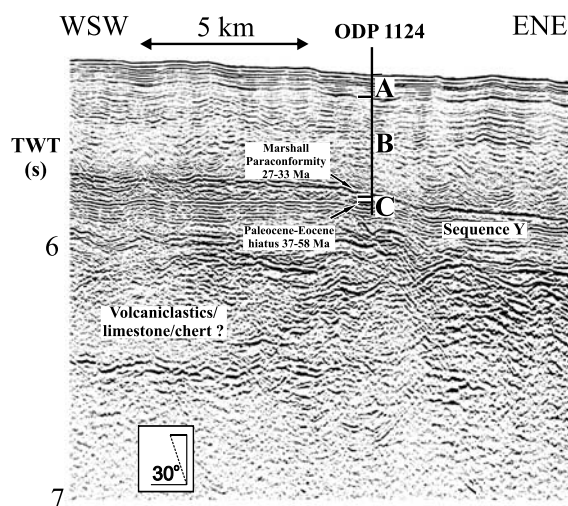

Figure 14. Seismic section on profile CR3057-1. Location as in Figure 11. Sequence Y (70–32 Ma). The location and depth extent of ODP hole 1124 is shown. Seismic units A, B, and C from *Carter et al.* [1999] are plotted. Seismic data provided by the National Institute of Water and Atmospheric Research (NIWA). Inset shows the true slope as per Figure 5.
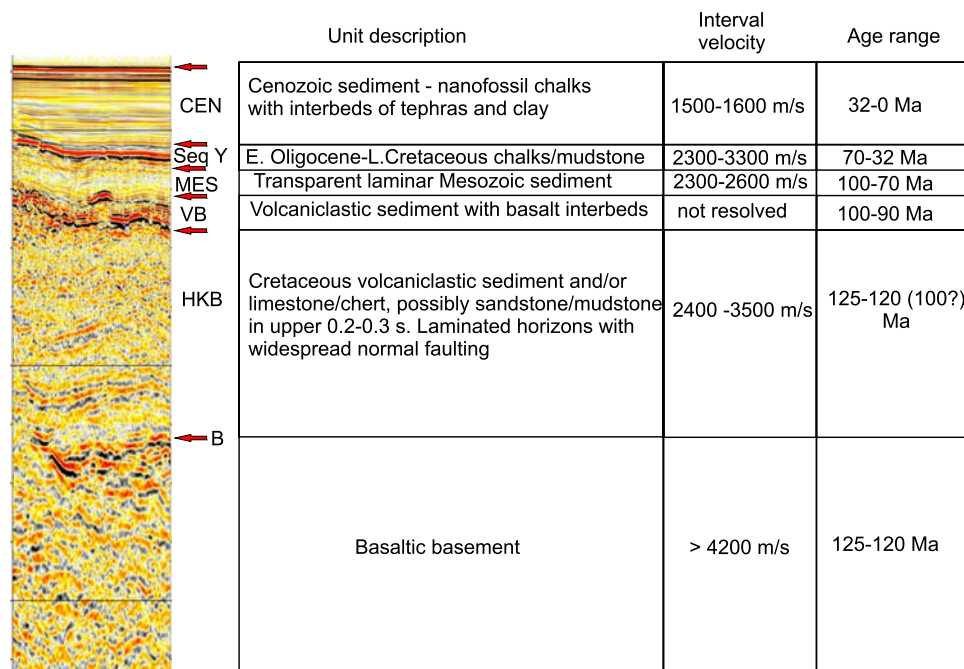
Unit description	Interval velocity	Age range
Cenozoic sediment - nanofossil chalks with interbeds of tephras and clay	1500-1600 m/s	32-0 Ma
E. Oligocene-L.Cretaceous chalks/mudstone	2300-3300 m/s	70-32 Ma
Transparent laminar Mesozoic sediment	2300-2600 m/s	100-70 Ma
Volcaniclastic sediment with basalt interbeds	not resolved	100-90 Ma
Cretaceous volcaniclastic sediment and/or limestone/chert, possibly sandstone/mudstone in upper 0.2-0.3 s. Laminated horizons with widespread normal faulting	2400 -3500 m/s	125-120 (100?) Ma
Basaltic basement	> 4200 m/s	125-120 Ma

Figure 15. Sedimentary units of the Hikurangi Plateau.

spread of these unconformities, the Marshall Paraconformity (27–33 Ma), bounds the upper surface of Sequence Y. The Marshall Paraconformity is a regional New Zealand feature formed by widespread erosion of the ocean floor around 32 Ma [*Kennett et al.*, 1985; *Carter et al.*, 2004; *Fulthorpe et al.*, 1996]. *Carter et al.* [1999] interpreted this unconformity at Hole 1124 as marking the inflow of the Deep Western Boundary Current to the Southwest Pacific Ocean [*Kennett*, 1977; *Carter and McCave*, 1994]. A 3 Ma (34–37 Ma) late Eocene hiatus lay 10 m deeper in hole 1124 and a 21 Ma Paleocene-Eocene (37–58 Ma) hiatus was encountered 9 m deeper in Hole 1124 (Figure 14).

[25] Hole 1124 bottomed at ~65 Ma and did not reach the base of Sequence Y, hence this age is extrapolated to be ~70 Ma. Sequence Y thins on line HKDC1 to ~40 ms TWT thick between the three major seamounts of the southern Northern Volcanic Region. Otherwise it is consistently ~150 ms thick over an area extending from 30 km south of the Rapuhia Scarp southward across the entire plateau, the North Chatham Slope and onlapping onto the crest of the Chatham Rise. The base of Sequence Y is hard to recognize as it traverses the half-grabens near the Chatham Rise crest. *Wood and Herzer* [1993] have interpreted sediments in the half-grabens on the rise crest as being ~100 Ma and younger. Although there is some displacement of Sequence Y as it traverses the North Chatham Slope and the base of the slope (Figure 6), the continuity of the sequence and its consistent thickness suggests that convergent motion across the Gondwana margin had ceased by the time Sequence Y was deposited.

4.2.2. CEN: Cenozoic Sedimentary Unit

[26] Above Horizon Y, late-Oligocene to Recent sediments (CEN) (Figure 15) across the Hikurangi Plateau are, on the basis of Sites 1123 and 1124 (Figures 3, 13, and 14) [*Carter et al.*, 1999], interpreted as nanofossil chalks with interbeds of tephras and clay deposits. Although generally well laminated these sediments are frequently disrupted by erosional channels, unconformities and sediment slump/compaction deposits. This is particularly pronounced within the southern 200 km of the Southern Basement High and North Chatham Basin area (Figure 6). Deep Western Boundary Current transport and erosion [*Carter et al.*, 2004], channeled turbidite deposition of sediment derived from the post-25 Ma developing compressional plate boundary, onshore New Zealand, and slumping of sediment from the North Chatham Slope have combined with the deposition of deep-water calcareous oozes to form the CEN sedimentary unit.

4.2.3. MES

[27] Beneath Sequence Y is approximately 200–1300 ms of relatively low amplitude reflectivity,

well-laminated, sedimentary section (MES, Figures 5–10 and 15). This section thins to near zero at the northwestern Hikurangi Plateau and is thickest in the North Chatham Basin and within grabens on the Chatham Rise. There are at least two characteristic high-amplitude reflection horizons within MES that are recognizable on the Chatham Rise and the southern Hikurangi Plateau. The MES unit generally mantles and infills topographic relief on the volcanic basement and is relatively undisturbed by channeling or faulting except above the northern toe of the North Chatham accretionary prism. It is interpreted to correspond to the Cretaceous MES2 unit of *Wood and Davy* [1994]. The infilling, ponded nature of the sediment suggests that it is redeposited clastic sediment rather than pelagic drape. The absence of channeling suggests deposition in a low-energy marine environment.

4.2.4. HKB: Hikurangi Basement (Volcaniclastic/Limestone/Chert)

[28] Underlying unit MES is a ~100 ms TWT (~150 m) thick unit (unit VB) (Figure 15) of high-amplitude reflecting horizons that form the upper section of a ~300–1500 ms TWT (500–2000 m) thick unit (Unit HKB, Figure 15) of laterally disturbed, discontinuous and faulted reflectors. Unit HKB is near the deeper limit of horizon resolution on line HKDC1. Within and near the base of the resolvable unit HKB are several regionally correlatable high-amplitude reflection horizon units (e.g., Figure 9). Unit VB is principally identified within the southern Northern Volcanic Region at the upper surface of unit HKB (Figures 7–9).

[29] Multichannel seismic stacking velocity analysis gives a velocity for Unit HKB of 2400–3500 m/s. A strong reflecting horizon (horizon B) interpreted as marking the base of Unit HKB and highlighted in Figures 6–9 has underlying interval velocities of >4000 m/s. Despite being somewhat discontinuous the reflectors within unit HKB display areas of strong lamination disrupted by widespread faulting (e.g., Figures 7–9 and Figure S1). As an LIP the basement of the plateau is expected to be basaltic. A seismic velocity of 2400–3500 m/s, as interpreted for unit HKB on the Hikurangi Plateau, is too low to be the bulk velocity of a basalt sequence [*Planke and Cambray*, 1998; *Planke et al.*, 1999]. Such velocities would be more characteristic of volcaniclastic sediments or limestone/chert sediment similar to that sampled on Ontong Java and

Manihiki plateaus. Abundant volcaniclastic rocks (much of it being hyaloclastite) were dredged from the upper portion of the Northern Volcanic Region margin (Rapuhia Scarp and southeast of it) during the SO168 cruise [*Hoernle et al.*, 2003]. Volcaniclastic sandstone/siltstones have been sampled over a 300 m interval on the Manihiki Plateau. Many of these sediments are hyaloclastite in nature [*Jenkyns*, 1976]. The compressional wave sound velocity over this interval is only ~2000 m/s peaking to ~2500 m/s over a 50 m interval [*Boyce*, 1976].

[30] The upper ~one third of unit HKB has a similar seismic reflection-derived interval velocity, ~2500 m/s, and a similar thickness, ~300–600 ms (TWT) to that of the Cretaceous sandstone/mudstone unit OJ2 interpreted across the Ontong-Java Plateau [*Phinney et al.*, 1999]. Compilation of sonic logs and lab-based core measurements for DSDP/ODP drill sites on the Ontong Java Plateau by *Gladczenko et al.* [1997] identified a ~350 m thick layer of Cretaceous-Eocene limestone and chert interbeds, with an average velocity of 3500 m/s, extending across most of the plateau. A similar 350 m thick layer of Maastrichtian-Aptian age limestone and volcaniclastic sandstone/siltstone is interpreted in the DSDP 317 drill hole data, Manihiki Plateau [*Schlanger et al.*, 1976] however this unit has a lower 2500–2600 m/s velocity. Although the 350 m thick, 3500 m/s limestone/chert velocity from the Ontong Java Plateau is typical of the lower two thirds of unit HKB the latter sequence is much thicker (~1.5 km). The lower two thirds of unit HKB is heavily normal faulted consistent with such deformation having occurred prior to, or during, OJMHP breakup at ~120 Ma. We thus interpret much of this lower unit HKB to be volcaniclastic sediment overlain by, and possibly interspersed with, limestone and chert sequences.

[31] The underlying, ~1 s TWT (~1.5 km) deeper, strong reflecting horizons (e.g., horizon B and below) with velocities > 4000 m/s are more likely to be basalt flow deposits.

4.3. Chatham Rise Subduction

4.3.1. Seismic Reflection Data

[32] The subducting plate carrying the Hikurangi Plateau can be traced on seismic reflection profiles south for over 150 km beneath the Chatham Rise (Figures 4–6). Thrust faults and folds in the rocks of the North Chatham Slope overlying the subducted plateau are interpreted to reflect deforma-

tion associated with subduction of the plateau. The structure imaged on HKDC1 (Figure 6) is very similar to that imaged in the accretionary prism offshore North Island where the Hikurangi Plateau LIP is presently being subducted [*Lewis and Pettinga*, 1993; *Henrys et al.*, 2006]. This seismic section is the first direct evidence for an accretionary prism associated with subduction at the North Chatham Rise. Sediment from unit HKB, on line HKDC1, is interpreted as being thrust into the accretionary prism. The accretionary prism is recognizable in the North Chatham Slope up to 50 km south of the base of the slope. The southern boundary of the recognizable accretionary prism is marked by a major thrust fault (labeled NCT in Figure 6) that soles out at the interpreted subducted Hikurangi Plateau basement interface, horizon B. The slope of the subduction interface increases from 4.5° to 7° immediately south of the intersection with the above thrust fault. It is possible that this intersection marks a leading edge of the Hikurangi Plateau crust on Line HKDC1. The major thrust fault, NCT, displaces interpreted Cenozoic sediment, including Sequence Y, at the southern limit of the accretionary prism suggesting some localized movement in the Cenozoic (Figure 6). The Cenozoic sedimentary section overlying the interpreted accretionary prism is interpreted to have undergone slumping, displacing sediment above Sequence Y.

[33] Although the accretionary prism is obvious extending ~50 km south within the North Chatham Slope, it is not recognizable south of the major thrust fault NCT. In this sense the margin on line HKDC1 is similar to that observed off Hawke Bay in the North Island [*Henrys et al.*, 2006] and north of East Cape [*Collot and Davy*, 1998]. In the latter example, *Collot and Davy* [1998] have traced a major fault scarp, the Awanui Fault. Lying 20–40 km inboard of the active subduction trough, this fault extends ~200 km southward from the Kermadec Trench/Rapuhia Scarp intersection and separates the lower transcurrently faulted, eroding, ex-accretionary prism from a deforming backstop. South of fault NCT the Chatham Rise basement is relatively featureless, contrasting strongly with the sediment-filled half-graben structures cut into it (Figure 5).

[34] Late Cretaceous sediments comprising parts of Sequence Y and unit MES are interpreted on seismic sections, including HKDC1, to be widely distributed on the North Chatham Basin, the North Chatham Slope and in half-grabens on the Chatham Rise crest. On line HKDC1 (Figures 4 and 7) the MES unit has ponded within topography above the HKB unit forming a filled basin up to ~700 ms TWT (~800 m) deep and extending 130 km north-south. At its northern limit MES pinches out against the Southern Basement High (Figure 7). At its southern limit MES is interpreted to pinch out above the fossil accretionary prism (Figure 6). On HKDC1, unit MES is interpreted as being deposited postaccretionary prism development and is not incorporated into the accretionary prism. This contrasts with the interpretation further east on line HKDC3 where the MES unit is up to 1500 ms TWT (~1.5 km) deep and forms a basin extending ~125 km north-south. On line HKDC3 the MES unit appears to have been at least partially incorporated into the accretionary prism (Figure 10). This may indicate a variation in subduction cessation timing, or the mode of subduction, along the margin, possibly associated with a separate subduction history east and west of the Rekohu Embayment. On HKDC1 unit MES is also ponded within the topography above the accretionary prism and interpreted within at least one half-graben on the Chatham Rise crest (Figure 5). Most of unit HKB, which immediately underlies unit MES on the plateau, is conversely interpreted as having been either subducted beneath the margin or possibly incorporated into the accretionary prism. A separate unit VB (Figures 7 and 8) is not recognizable at the upper surface of HKB in, or south of, the Central Basin.

[35] The subducted plate interface beneath Chatham Rise is interpreted extending down to ~7.0 s TWT (~18 km) below the seafloor at the southern limit of line HKDC1. Several seamounts ~0.5–1.0 s (2–3 km) high and ~5–20 km in diameter are tentatively interpreted as irregularities on the dipping plate interface. One set of such peaks underlies the main horst structure at the Chatham Rise crest on line HKDC1. Another seamount is interpreted at the downdip limit of the fault NCT. Given the observed volcanic topography, particularly on the northern end of line HKDC1, and the scatter of seamounts penetrating the seafloor on the Hikurangi Plateau, such subducted seamounts are not unexpected, but suggest that subduction continued after at least some seamounts were formed. A sequence of north dipping reflectors is interpreted to extend from the southern limit of line HKDC1 down almost to the subducting plate interface. A similar set of dipping interfaces is observed 250 km further east on Line HKDC3. The origin of the reflectors is unknown, but they
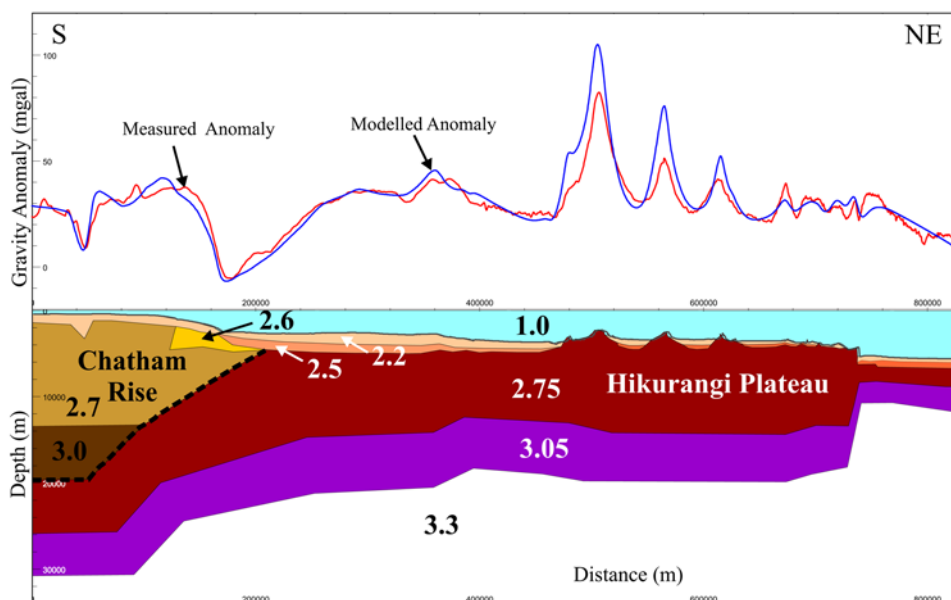
Figure 16. Two-dimensional gravity model along line HKDC1. Observed gravity is ship-borne data. Location in Figure 1. Densities in gm/cc used in the model are labeled. The dashed line marks the fossil plate interface interpreted from seismic reflection data. Ocean crust, modeled as 4–6 km thick lies north of the Hikurangi Plateau. Without velocity constraint the plateau thickness is highly nonunique. Plausible alternate density distributions can yield plateau crust up to 23 km thick. The ~40% thinning beneath the Chatham Rise is, however, common to all models.

may be displacement or intrusion surfaces that linked extension in the Bounty Trough south of the Chatham Rise crest with the subducting plate.

4.3.2. Gravity Model of the Subducted LIP

[36] Two-dimensional gravity modeling by *Davy and Wood* [1994] suggested a 15-km-thick crust of the Hikurangi Plateau abuts against 25 km thick Chatham Rise crust. Unlike similar models for the Ontong-Java Plateau [*Gladczenko et al.*, 1997] velocity information is lacking for most of the crustal section beyond the stacking velocity resolution of the 6 km streamer used to survey HKDC1. A revised gravity model (Figure 16) with lower crustal/mantle densities of 3.05 and 3.3 g/cc, the same densities as used by *Gladczenko et al.* [1997], and an expanded oceanic crustal density distribution [*Hussong et al.*, 1979] implies a slightly (~1 km) thicker plateau than the earlier *Davy and Wood* [1994] model. The model base of crust has been guided by seismic reflection interface information on this and nearby lines, although many of these reflector interpretations are inconclusive. The Hikurangi Plateau lower crust remains unconstrained by refraction velocity data and it is possible, perhaps even likely, that such information could constrain the Hikurangi Plateau to crustal thicknesses closer to the estimates for thicker sections

of the Manihiki and Ontong-Java plateaus. Alternate plausible density distributions [*Gladczenko et al.*, 1997] and thicker oceanic crust north of the Hikurangi Plateau can yield model results with plateau crustal thickness up to 23 km.

[37] Assuming an interval velocity of ~6.5 km/s below ~3 s TWT within the Chatham Rise, then the ~5–7 s TWT depth of the plate interface beneath the basement upper surface on the Chatham Rise indicates that the lower 30–40% of the rise is composed of rocks of the subducting plate. Gravity modeling of such a configuration indicates that using either 3.05/3.3 (Figure 16) or 3.0/3.4 g/cc lower crust/mantle density contrasts and the seismic reflection plate-interface constraint, the Hikurangi Plateau thins by about one third from 16 to 23 km to ~10–15 km beneath the Chatham Rise. The implied thinning of the Hikurangi Plateau crust under the Chatham Rise is interpreted to mark a leading edge to the plateau. This interpretation is reinforced by the observation of a set of reflectors ~3s TWT (~10 km) beneath the subducted basement and extending ~50 km beneath the North Chatham Slope (marked as "Base of plateau crust?" in Figure 6). These reflectors dip to the north beneath the fossil accretionary prism until they are no longer recognizable beyond the toe of the Chatham Rise slope. A ~10 km thick crust for the subducted plate

beneath the Chatham Rise would be consistent with the gravity modeling (Figure 16) and with this being subducted "normal" oceanic crust at the southern edge of the Hikurangi Plateau. In contrast Line HKDC3 (Figure 10) extends almost twice as far (~100 km) south to beneath the crest of the Chatham Rise before a possible leading edge of the plateau is observed ~30 km from the southern limit of line HKDC3.

4.4. Southern Basement High: Central Basin

[38] On line HKDC1 (Figure 7) the Southern Basement High is dominated by a series of faulted volcanic basement highs at the northern limit of the area. To the south of these highs, the MES unit thickens toward the foot of the North Chatham Slope onlapping and infilling the underlying basin. Conversely the underlying HKB unit thins southward from ~1 s TWT thickness adjacent to the basement highs.

[39] Reflecting unit HKB and horizon B step down to the north by ~0.5 s across these basement highs. Conversely Sequence Y which is only broken in continuity in a few places across these basement highs, drops by ~1 s and comprises a mixture of small <100 m down-thrown faults and north dipping bedding. It is inferred that a northward dip was already in existence when Sequence Y was deposited but that dip was subsequently accentuated by further down-faulting. Within the Southern Basement High bathymetric mapping [*Hoernle et al.*, 2003] has revealed guyots with interpreted wave-cut tops presently ~1500–2000 m below sea level. If the upper surface of unit HKB was deposited synchronously with seamount formation, and immediately prior to guyot erosion and planation across the plateau, then the Southern Basement High area will have similarly subsided from a depth 2–2.5 km at the time of unit HKB deposition.

[40] Within the Central Basin region Unit MES thickens toward the basement highs at the southern boundary of the area. Unit HKB is interpreted to climb toward the boundary with the Northern Volcanic region.

4.5. Northern Volcanic Region

[41] This region is characterized by shallow volcanic basement, relatively thin (<1 s) sedimentary cover, numerous volcanic intrusions and large volcanic features (~25 km diameter, 1000–

1500 m high above seafloor). The southern boundary of the Northern Volcanic Region is very abrupt on Line HKDC1 and lies between Palmer Seamount and the Hikurangi Channel (Figures 2 and 11).

[42] Line HKDC1 crosses three volcanic peaks, the southern two of which are guyots and the northern peak is interpreted as a near crossing of a nearby seamount. The southernmost of these peaks, Palmer Seamount was swath mapped by the SO 168 survey [*Hoernle et al.*, 2003] (Figures 11 and 12). The swath bathymetry reveals a 20 km long, 500–800 m high, arcuate volcanic ridge initially extending south, but later curving southeastward, from Palmer Seamount. This ridge is interpreted as a constructional volcanic rift system, a feature commonly associated with the shield stage of volcanism forming intraplate ocean island volcanoes such as Hawaiian and the Canary Island volcanoes. Line HKDC1 crosses the southern tip of the volcanic ridge extending south from Palmer Seamount (Figure 12). Swath bathymetry was not collected to the northwest of Palmer seamount but the trends on the bathymetry data collected combined with satellite gravity structure over the seamount suggest an arcuate volcanic ridge symmetric to the ridge on the opposite side of Palmer Seamount such that the volcano forms a reversed "S" in plan view. Swath bathymetry data from Lange Seamount ~100 km to the east-northeast suggest this seamount exhibits similar arcuate volcanic ridges extending from this seamount (Figure 12). Such volcanic ridges are likely to mirror either inherited basement grain or lines of strain in place during eruption of the seamounts. Many of the guyots, such as Palmer Seamount, are polygonal in structure and for the majority of guyots swath-mapped across the Hikurangi Plateau there is a long axis trending northwest to north-northwest. Complementary to this long axis many of the guyots have shorter linear boundary segments trending northeast. The northwest trend of the seamount structure, along the long axis of the volcano body, suggests this as one of the main basement grain directions.

[43] The OJMHP is interpreted [*Neal et al.*, 1997] to have erupted onto oceanic crust of approximately magnetic anomaly M29-M0 (157–125 Ma) age [*Nakanishi et al.*, 1992]. The northeast oriented spreading ridge grain, revealed by the magnetic anomalies north of Ontong-Java Plateau, is approximately orthogonal to the dominant northwest, and

parallel to the secondary northeast, structural grain of the Hikurangi Plateau basement.

[44] Approximately 80 km southwest of the Rapuhia Scarp, the seafloor on line HKDC1 descends onto an 80- to 100-km-wide, 4-km-deep terrace, hereafter referred to as the North Hikurangi Terrace, along the northern plateau limit (Figures 9 and 11). Horizon VB steps down at the southern boundary of this terrace to be ~5.7 s TWT below sea level across the terrace. The step in basement appears part of an extensional rotated block, down-thrown to the southwest, but masked by overlying volcanics. Ridge-like volcanoes (see below) in the Northern Volcanic Region occur preferentially along such basement steps presumably erupting along normal faults formed during Manihiki-Hikurangi Plateau rifting.

[45] Approximately 60 km southeast of line HKDC1 the bathymetry and underlying stratigraphy steps down to the southeast by ~1 km onto a terrace hereafter referred to as the Northeast Hikurangi Terrace (Figure 11). Although still in the Northern Volcanic Region, seamounts Lange, Katz and Muldoon occur in the Northeast Hikurangi Terrace region, where the unit VB, interpreted from low-fold seismic data, lies at ~6.5 TWT below sea level, approximately 1.2 s TWT deeper than VB further west. Some horizon reflectivity within unit HKB is apparent to ~1 s below unit VB between the three major seamounts on line HKDC1. However, it is north of these seamounts, extending to within 25 km of the Rapuhia Scarp, that ~1.0–1.5 s TWT of high-amplitude laminar reflectivity is apparent within HKB. These laminar horizons are heavily faulted. At the level of fault displacement within Unit B the faulting appears extensional in character.

4.6. Rapuhia Scarp: Rift Margin?

[46] The Rapuhia Scarp, the northern boundary of the Hikurangi Plateau with the deep ocean floor of the South Pacific basin, is a steep, sharp ~1 km high margin that extends about 200 km southeast from its intersection with the Kermadec Trench. Further to the southeast, between longitude 177°W and 175.5°W, the margin is still sharp but is buried by sediments. Morphologically the margin in the southeast is broader and more diffuse than to the northwest.

[47] Swath bathymetry data collected in the region of the northern Hikurangi Plateau margin (Figures 11 and 12) reveal a number of the "ridge-like" seamounts occurring within ~70 km of the Rapuhia Scarp [Hoernle et al., 2004]. These ridge-like

seamounts, which are oriented subparallel to the Rapuhia Scarp margin, generally formed at the northeasternmost edge of terraces and have sharp crests showing no evidence of seafloor erosion. These "ridge-like" seamounts form, in common with guyot seamounts, part of northeast oriented seamount triplets (Figure 3) along the northern Hikurangi Plateau. For instance the Muldoon "ridge-like" seamount forms a triplet with Lange and Katz Seamounts (Figure 12).

[48] The seafloor on line HKDC1 steps down ~400 m (~0.6 s TWT) from the North Hikurangi Terrace onto a 30 × 20 km terrace that protrudes northeast from the general southeast trend of the Rapuhia Scarp margin (Figures 9, 11, and 12). The seismic data indicate that the basement of this lower terrace is, like the basement step at the southern margin of the North Hikurangi Terrace, back-tilted to the southwest. Unit B, or deeper, rocks are interpreted as comprising most if not all the Rapuhia Scarp at this location. Unit VB is interpreted to crop out within ~5–10 km of the Rapuhia Scarp and there is little (<100 m) overlying sediment above this unit across the terrace. Given the manner in which this terrace protrudes from the margin it appears likely that currents of the Deep Western Boundary Current System [Carter et al., 2004] have swept the terrace free of significant sediment accumulation. Beneath unit VB, unit HKB is ~0.75 s TWT thick compared with 1–1.5 s TWT 30 km or more to the south.

[49] While extensional rotated block fault features are characteristic of the northern Hikurangi Plateau margin as far east as line HKS2 (Figures 12 and 17), further southeast line CR3057–8 (Figures 11 and 18) reveals a change in the nature of the plateau margin. The northern limit of the plateau coincides with an edge to a ~0.5–1 s TWT (0.6–1.7 km) thick layer of unit HKB (interpreted volcaniclastics/limestone/chert) deposited on top of a layer of strongly laminated reflectors (horizon B?). The deeper horizons B and below are tentatively interpreted as basement lava flows of the LIP. Unit HKB thickens to the southwest as the strongly laminated layer deepens. Although some of the apparent shallowing of the HKB-B interface may be due to the higher velocity of unit HKB (2400–3500 m/s) versus the adjacent sedimentary layer (2000–3000 m/s), the contrast is not considered sufficient to explain much of the apparent shallowing. Both unit HKB and the underlying horizon B appear disrupted by normal faulting. This contrasts with the sedimentary layers adjacent
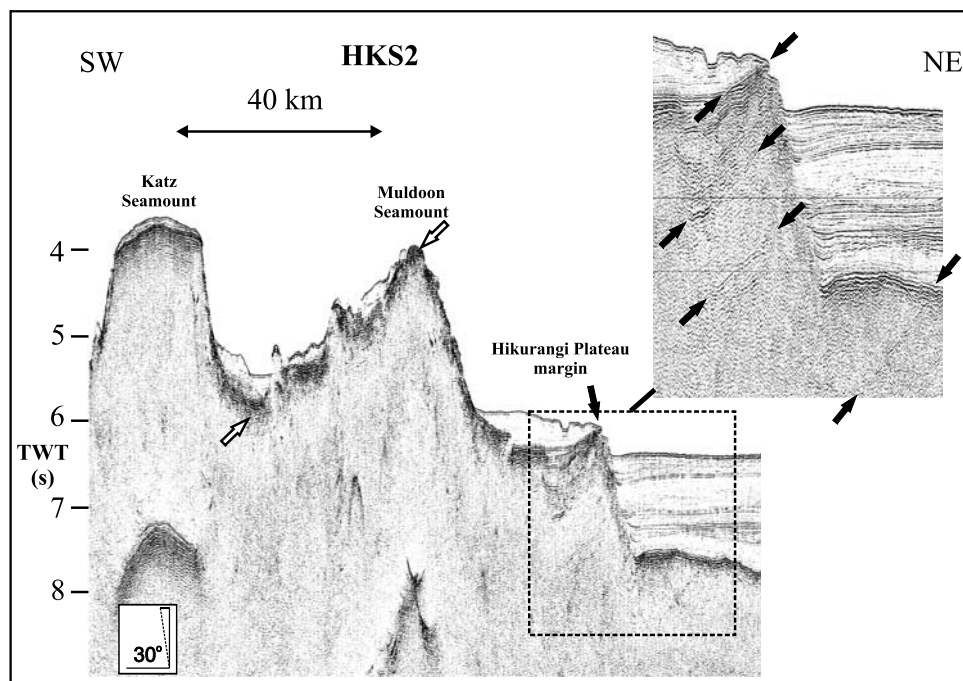
Figure 17. Seismic section on profile HKS2. Location as in Figures 11 and 12. Southwest dipping units are highlighted by solid arrows. A similar basement dip in Muldoon Seamount is highlighted by the hollow arrows. Bottom left inset shows the true slope as per Figure 5.

to the margin and those overlying unit HKB. Laterally coherent reflectors are not recognizable in the oceanic crust northeast of the interpreted margin. The interpreted normal faulting is tentatively associated with Hikurangi-Manihiki Plateau rift breakup at the margin.

[50] A similar edge to unit HKB is observed further east on line CR3057-1 (Figures 11 and 18) where the line crosses the plateau margin into the Rekohu Embayment. Unit HKB and deeper layers near this margin also appear disrupted by mainly normal faulting in contrast to the overlying sequence Y.
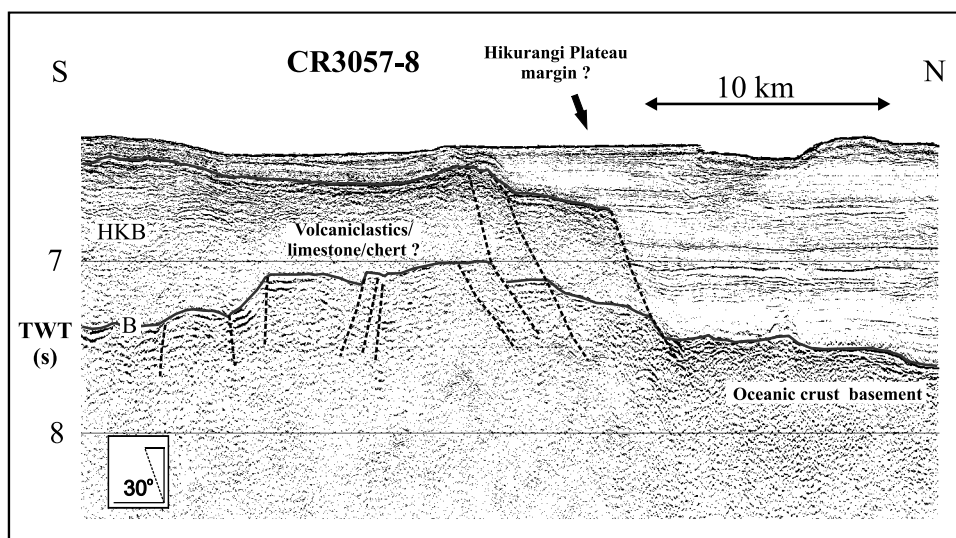
Figure 18. Seismic section on profile CR3057-8 showing faulting at the edge of the Hikurangi Plateau. Location as in Figure 11. HKB, Hikurangi Basement (volcaniclastic/limestone/chert); B, Hikurangi Basal Interface (lavas?). Faults marked as dashed lines. Seismic data provided by the National Institute of Water and Atmosphere (NIWA). An uninterpreted version of this line is available as Figure S9. Inset shows the true slope as per Figure 5.

The interpreted oceanic crust within the Rekohu Embayment is highly irregular with a major step in the basement near the projected southeast extension of the Rapuhia Scarp (Figure 18).

5. Discussion

5.1. Plateau Formation and Breakup

[51] One model of Large Igneous Province formation involves excess volcanism occurring above the plume interaction with a spreading ridge (e.g., Iceland [*Coffin and Gahagan*, 1995]). The excess volcanism and flood basalts associated with such interaction form the LIP and the spreading ridge dynamics lead to splitting and rifting apart of two or more sections of the LIP. While the formation of the proposed Ontong-Java-Hikurangi-Manihiki Plateau (OJHMP) LIP [*Taylor*, 2006; Hoernle et al., manuscript in preparation, 2008; B. Davy, From Ontong Java/Manihiki/Hikurangi Plateau breakup to New Zealand–Antarctica breakup—The role of LIP subduction/collision in plate motion evolution, manuscript in preparation, 2008] probably closely preceded its rifting apart, *Taylor* [2006] noted that the dates of formation of all three plateaus are too synchronous, across such a large area (∼1% of the Earth's surface), for them to have formed via the Iceland spreading ridge model. Although alternative models of plateau formation such as bollide impact [*Ingle and Coffin*, 2004] have been proposed a satisfactory explanation of the apparently rapid, voluminous and widespread volcanism is still to be established. LIP formation and rifting are closely separated in time and close in location to the M0 magnetic anomaly (125 Ma) that *Taylor* [2006] has inferred lies beneath the Danger Islands Trough that extends NNE across Manihiki Plateau.

[52] Seismic reflection data have revealed block faulting in the Hikurangi Plateau basement consistent with northeast directed rifting apart of the Hikurangi-Manihiki plateaus at the Rapuhia Scarp and its southeastern extension (Figure 17). This margin is however highly variable in its character. Two major terraces to the north and east of the Northern Volcanic Region have possibly evolved from more uniform LIP crust by subsidence following extension and block faulting. The Northern Hikurangi Terrace has several sets of northwest oriented ridge-like seamounts within and bordering it consistent with such northeast oriented extension (Figures 11 and 12). The location of the ridge-like seamounts on terrace boundaries and their asym-

metric, possibly block-rotated structure, favors an interpretation of these seamounts forming during OJMHP breakup ∼120 Ma.

[53] Further east the Hikurangi Plateau margin is increasingly dominated by the Central Basin structure cored by the Rekohu Embayment (Figures 2, 3, and 11). The Northeast Hikurangi Terrace represents an intermediate step down to the Central Basin, parallel to the northeast alignments of both guyot and ridge-like seamounts within the Northern Volcanic Region. The center of the Rekohu Embayment is cored by an interpreted 100 × 60 km northeast oriented rectangular embayment in the margin of the Hikurangi Plateau. Oceanic crust is interpreted within this embayment, reinforced by gravity models (not shown) of the ∼20 mgal high in the embayment (Figure 3). The oceanic crust within the embayment is ∼600 m shallower than ocean crust north of the plateau possibly indicating formation at a different period or simply a manifestation of basement shallowing close to the plateau as seen on many seismic reflection profiles (e.g., Figure S1 and Figure 17).

[54] If the Hikurangi Plateau and Manihiki plateaus rifted apart the interpreted normal faulting within horizon B and unit HKB may record this extensional episode. The Rapuhia Scarp and other terrace steps in basement topography near the northeastern boundary also show (on some of the lines that cross the margin and bordering seamounts (Figure 17)) rotated fault block formations possibly associated with plateau rifting. The almost-buried northeastern Hikurangi Plateau margin on line HKS2 (Figure 17), and possibly Muldoon Seamount, illustrate such basement rotation features and contrast with the Katz Seamount guyot further south on the same line. Minor volcanic cones on its southern flank and its asymmetric basement structure suggest Muldoon Seamount may be formed from a combination of rotated tectonic crustal blocks and later volcanism. The northern lower terrace that line HKDC1 crosses before descending to oceanic crust appears to be a single back-tilted crustal block.

[55] Plate reconstruction modeling between Manihiki and Hikurangi plateaus (Davy, manuscript in preparation, 2008) has traced transform fault and spreading ridge evolution in satellite gravity data and swath bathymetry data as well as matching the interpreted morphological fit of the two plateau margins. The resultant close 120 Ma fit of the Manihiki Plateau against the Hikurangi Plateau (Figure 19) (Davy, manuscript in preparation,
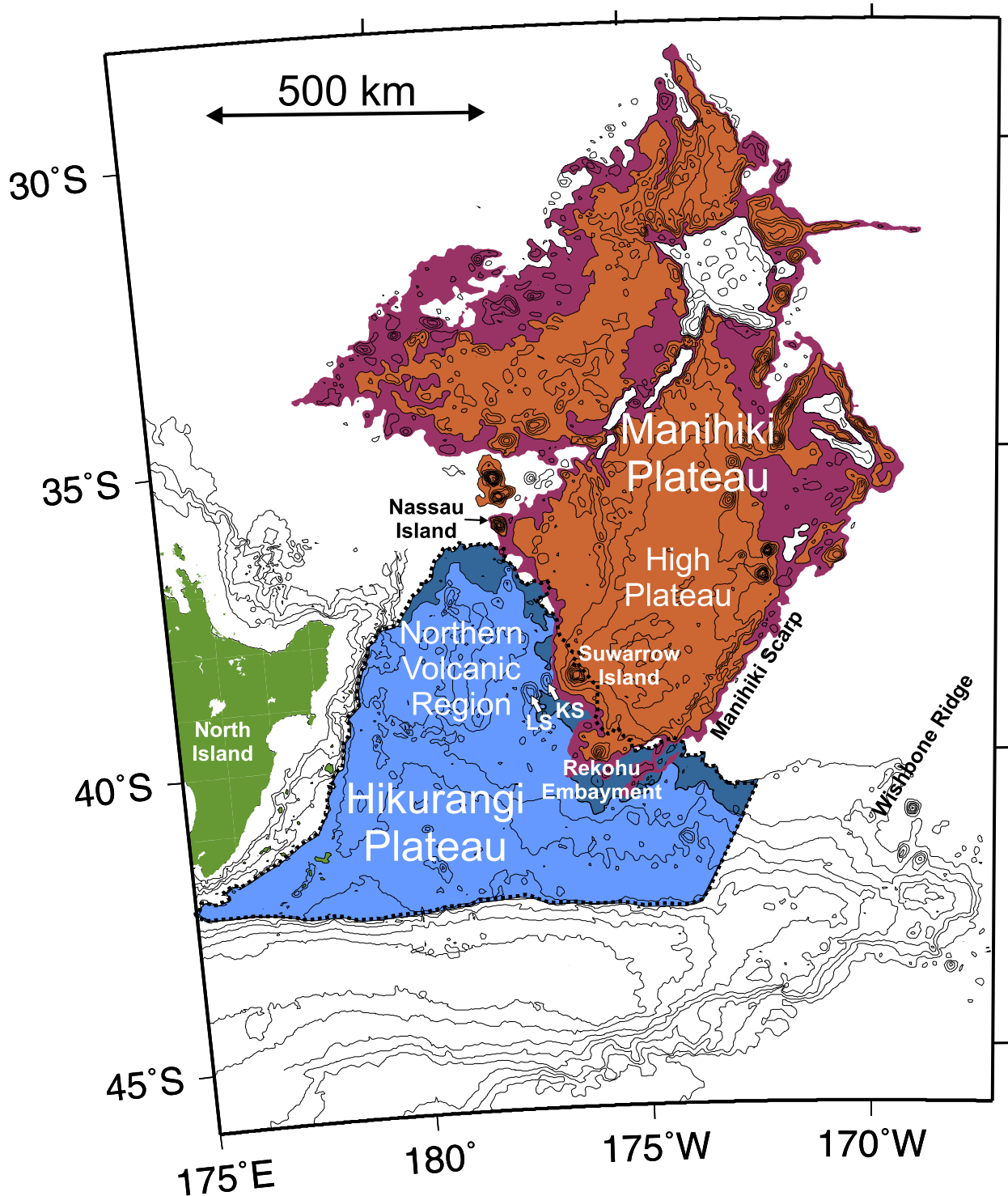
Figure 19. Prebreakup match of the Hikurangi and Manihiki plateaus (from Davy, manuscript in preparation, 2008), with Hikurangi Plateau kept fixed. Water depths between 4500 and 4000 m on the Hikurangi and Manihiki plateaus have been shaded dark blue and maroon, respectively, and water depths shallower than 4000 m have been shaded light blue and orange, respectively. Contour interval is 500 m. The black dashed line of the northeastern 4500 m contour (edge of dark blue) for the Hikurangi Plateau has been overlaid to indicate the overlap of the repositioned two plateaus. KS, Katz Seamount; LS, Lange Seamount.

Table 1. Plate Motion Changes, Related Tectonic Events, and Spreading Ridge Orientation/Rate

Pacific Plate Motion Changes[a]	Tectonic Event	Spreading Ridge Analysis[b]
124 Ma: 60° clockwise	123–120 Ontong Java/Manihiki/Hikurangi Plateau formation and breakup	
124-110 Ma		>7 cm/a, 2400 km, 15–20°, 17–24 cm/a
110 Ma: 45° anticlockwise	Convergence obliquity change at Gondwana margin: extension begins onshore NZ (back-arc opening?)	
110–105 Ma		>7 cm/a, 900 km, 2–5°, 18 cm/a
105 Ma	Initial Hikurangi Plateau collision with Gondwana margin	
105–100 Ma	Hikurangi Plateau subduction slows	2–6 cm/a, 200 km, 2–5°, 5 cm/a
100 Ma	Hikurangi Plateau subduction stops, Osbourn Trough spreading stops	
99–89 Ma	HIMU volcanism throughout Hikurangi Plateau, Chatham Rise, onshore South Island	
96 Ma: 75° clockwise	North Chatham subduction jammed (slab detachment?); extension within NZ and Gondwana switches from NE to NW	
85 Ma: 20° clockwise	Seafloor spreading begins between Chatham Rise/Campbell Plateau and Marie Byrd Land	

[a] *Kroenke et al.* [2004].
[b] *Downey et al.* [2007]. Entries are as follows: inferred full-spreading rate, ocean crust formation, spreading orientation, and full-spreading rate implied by *Kroenke et al.* [2004] dates.

2008) provides reinforcement of the argument for their formation as part of a common OJMHP LIP and subsequent rift separation. In the fit shown in Figure 19, the Rekohu Embayment matches with the volcanic promontory at the southeastern limit of the Manihiki Plateau and the Manihiki Scarp intersects the Hikurangi Plateau at the eastern margin of the Rekohu Embayment. Despite the deformation of the plateau margins by the rifting processes the match between the respective 4000 and 4500 m contours outlining each plateau is very good. Notable is the match between the Rapuhia Scarp and the Nassau-Suwarrow Scarp and the complementary nature of the elevated Northern Volcanic Region in the west of the Hikurangi Plateau with the elevated High Plateau in the east of the Manihiki Plateau. This partitioning of the elevation between the two plateaus segments has implications for the volcanism and rift mechanisms involved in OJMHP formation and breakup. There remains a strong NE–SW grain to both the Manihiki and Hikurangi plateaus, most obvious in the satellite gravity data (Figure 3) which we attribute to the M29-M0 seafloor spreading grain upon which much of the superplateau was deposited [*Taylor*, 2006]. The alignment of the Suwarrow Island on the Manihiki Plateau with the Lange-Katz-Muldoon Seamounts may be coincidence but is suggestive of formation along a common zone of weakness, possibly related to the underlying crustal structure,

which has been exploited during plateau breakup and later.

[56] *Mortimer et al.* [2006], *Taylor* [2006], and Hoernle et al. (manuscript in preparation, 2008) have proposed that breakup of the Manihiki-Hikurangi Plateau occurred ~120–115 Ma, soon after plateau formation and that the Hikurangi Plateau collided with the Chatham Rise margin of Gondwana ~105–100 Ma. *Downey et al.* [2007] examined spreading ridge fabric between Manihiki and Hikurangi plateaus and identified two periods, differing by ~15° in spreading orientation, of >7 cm/a spreading that produced 3300 km of separation between the two plateaus. The 1200–1800 km spreading between the Osbourn Trough and the Hikurangi Plateau over a 10–20 Ma period implies a half-spreading rate for the Osbourn Trough spreading center of 6–18 cm/a. There is however considerable uncertainty in this timing as the breakup time is assumed to be close to plateau formation age and the collision with the Hikurangi Plateau is linked to the onset of extension within the New Zealand sector of Gondwana [*Bradshaw*, 1989]. The degree of synchronicity of these events remains to be established.

5.2. Gondwana Subduction Cessation and Dynamics

[57] The Hikurangi Plateau lies beneath the Chatham Rise and extends north from it. Along the western
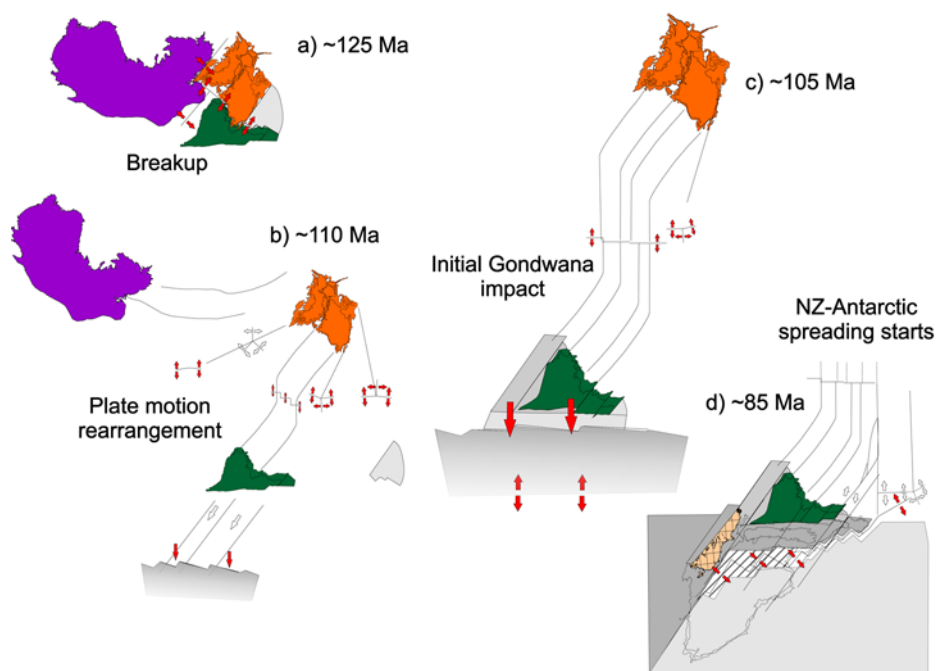
Figure 20. Simplified schematic of mid-late Cretaceous plate motion stages associated with Chatham Rise–Gondwana subduction. See Table 1 for description of plate motion change events. Empty arrows indicate earlier motion patterns.

part of the rise, between the Hikurangi Trough and the Chatham Islands, the junction between the Chatham Rise and the Hikurangi Plateau is quite sharp and 2,000–2,500 m deep (Figure 2). East of the Chatham Islands the northern margin of the Chatham Rise becomes broader and more complex, and deepens to 4,000–5,000 m at the extreme eastern end.

5.2.1. Cretaceous Gondwana Subduction at North Chatham Rise

[58] The Chatham Rise is interpreted to have formed between about 300 and 100 Ma as part of the New Zealand portion of the Gondwana margin. Convergence along this margin during that time resulted in the repeated accretion of terranes to the continent [*Mortimer*, 2004]. The degree of metamorphism of the Chatham Island schist indicates uplift and erosion of up to 5 km between about 195 Ma and 165 Ma (Late Jurassic and Early Cretaceous) [*Adams and Robinson*, 1977]. This predates subduction of the 120 Ma Hikurangi Plateau and is probably the result of earlier terrane accretion.

[59] The structure of the basement and sedimentary sequences along the northern part of the Chatham Rise reflect the final stages of the long period of convergence. This region is interpreted to have been a subduction zone along the Gondwana

margin during the Cretaceous. Deformation along the margin was caused by the collision of the Hikurangi Plateau with Gondwana about 105–100 Ma [*Bradshaw*, 1989; *Davy*, 1992; *Davy and Wood*, 1994].

[60] Absolute plate motion studies for the Pacific [*Kroenke et al.*, 2004], assuming a fixed hot spot frame of reference, predict major changes in plate motion at 124 Ma, 110 Ma, 96 Ma and 85 Ma (Table 1). The hot spotting technique reconstruction of *Kroenke et al.* [2004] is based upon seamount trails within the Cretaceous Pacific Plate crust. The hot spot tracing model is dependent on an assumption that the plates move over hot spots fixed within a framework decoupled from the global convection pattern and ensuing relative plate motions. Paleomagnetic investigations [*Tarduno et al.*, 2003], plate circuit studies [*Stock*, 2003] and geodynamic modeling [*O'Connell et al.*, 2003] of hot spot trails such as the Hawaiian-Emperor Chain have highlighted problems in the hot spot fixativity assumptions with the consequent models generally invoking concepts of "mantle drift" to explain hot spot motion. The use of hot spots to trace relative plate motion are further complicated by conclusions such as those of *Koppers and Staudigel* [2005] that some hot spot trail bends may be related to stress changes within the affected plate.

Regardless of the model for hot spot migration path the relative timing of hot spot trail kinks have been generally related to major plate tectonic events resulting from changes in plate motion or stress.

[61] For this study we are concerned principally with the timing of tectonic events during the Cretaceous Normal Superchron, a period from which seafloor spreading magnetic anomaly dates are not available. When correlated with timing estimates from fracture zone morphology and seafloor spreading hill fabric [*Downey et al.*, 2007], radiometric ages of LIP formation and guyot volcanism [*Parkinson et al.*, 2002; Hoernle et al., manuscript in preparation, 2008], and the interpreted dates of New Zealand–Antarctic rifting [*Davy*, 2006b] the timing of predicted plate motion changes, including those indicated by hot spot trails, are estimated to be within ±3 Ma.

[62] The hot spot tracing of *Kroenke et al.* [2004] has not allowed for seafloor spreading in the Ellice Basin as proposed by *Taylor* [2006], or independent motion of the Hikurangi and Manihiki plate segments [*Downey et al.*, 2007]. Allowing for such motion and placing the possible formative/breakup hot spot source at the M0 location within the OJMH Plateau as suggested by *Taylor* [2006] increases the likelihood that the OJMH Plateau hot spot was originally the Louisville Seamount Chain hot spot. This possibility is consistent with modeled palaeolatitudes for the Ontong Java Plateau [*Antretter et al.*, 2004].

[63] Figure 20 is a simplified schematic of some of the plate motion changes outlined in Table 1. Figure 20a depicts a closely grouped OJMHP. While we infer a close fit of the Hikurangi Plateau against the Manihiki Plateau the close fit of the Ontong-Java Plateau against the Manihiki Plateau, similar to *Taylor* [2006], is more speculative. The thicker more volcanically smooth crust of the Ontong-Java Plateau [*Winterer et al.*, 1974] contrasts significantly with the much more faulted and guyot studded crust of the Hikurangi Plateau. The OJMHP breakup to New Zealand–Antarctic breakup occurs entirely during the Cretaceous Normal Superchron. As a result modeled plate motions such as those in Figure 20 are all dependent on interpreted seafloor spreading fabric with constraining ages and chemistry from rocks between the plateaus largely absent and awaiting future surveys. The change in motion at 124 Ma predicted by *Kroenke et al.* [2004] is close to the emplacement and breakup age (125–120 Ma) for the

OJMHP. This age also marks the onset of the Cretaceous normal superchron, possibly linked to planetary circulation changes that have influenced motion in the Earth's core [*Larson*, 1991] and volcanism at the Earth's surface.

[64] The change in Gondwana margin plate motion at 110 Ma is possibly linked to Ontong-Java-Manihiki spreading cessation although an 82.6 Ma basalt has been reported [*Duncan*, 1985] at 10°S, 179.4°W. The 110 Ma plate motion change will have led to a change in subduction obliquity along the New Zealand sector of the Gondwana margin. Such a change in subduction dynamics may in turn have led to the onset of back-arc extension and rifting in the Gondwana margin consistent with the onset of extension observed from 110 Ma [*Mortimer et al.*, 1999, 2002]. *Downey et al.* [2007] interpreted a ~15° anticlockwise rotation of spreading direction ~400 km either side of the Osbourn Trough. We suggest this change in spreading orientation is correlated with the *Kroenke et al.* [2004] proposed 110 Ma, 45° plate motion change for the Cretaceous Pacific plate. Analysis of abyssal hill fabric south of the Manihiki Plateau led *Downey et al.* [2007] to recognize there must have been a triple junction, similar to the Tongareva Triple Junction [*Larson et al.*, 2002], that originated close to the Manihiki Scarp. They also highlighted the necessity for a further triple junction to the southwest of the Manihiki Plateau.

[65] At ~105 Ma the leading edge of the Hikurangi Plateau is inferred to have entered the Gondwana convergent margin. Gravity modeling (Figure 16) suggests the Hikurangi Plateau extends ~60 km south beneath the Chatham Rise along seismic reflection line HKDC1, equivalent to ~0.6 Ma at a 10 cm/a half spreading rate. The initial response to Hikurangi Plateau arrival in the Chatham Rise Trench will have been compression and uplift of the margin. The segmenting of the northern Chatham Rise margin into at least three sections (Figure 2) suggests that this compression may have occurred differentially along the margin segmented by motion parallel to the NE-SW grain of the Hikurangi Plateau and the transform fault fabric of the seafloor northeast of the Hikurangi Plateau. The plate convergence rate may have been much less than 10 cm/a at the time of plateau collision, different sections of the margin (e.g., HKDC3) will have experienced varying amounts of plateau underthrusting and there may have been erosion and compression not preserved in the seismic record. Similarly the southern margin of the Hikurangi

Plateau is likely to have been thinner (e.g., gravity model Figure 16) with tapered volcanic flow units in contrast to the rifted northern margin. Consequently the time duration between first arrival of the Hikurangi Plateau in the Gondwana trench (~105 Ma) and the jamming of the margin is tentatively estimated at ~1–5 Ma. This estimate compares closely with the 2–6 Ma, 100-km-wide zone of extension either side of the Osbourn Trough that *Downey et al.* [2007] have interpreted formed following Hikurangi Plateau collision. A similar "hard docking" period of ~4–5 Ma has been interpreted for Ontong-Java Plateau collision with the Solomon Island Arc [*Phinney et al.*, 1999].

[66] As subduction of the buoyant crust of the Hikurangi Plateau slowed and then jammed in the Gondwana margin at ~100 Ma, the extension previously realized by seafloor spreading in the Osbourn Trough will have also slowed then ceased as motion was increasingly transferred through the Chatham Rise via transform motion along the Wishbone Ridge to manifest as extension behind the Gondwana margin [*Davy*, 2004, 2006a]. Coincident with the proposed change in relative plate motion at 100 Ma, *Larson et al.* [2002] have interpreted a change in the migration path of the Tongareva Triple Junction and a change in Pacific-Phoenix spreading direction from 171° to 164°.

[67] Volcanism at ~99–89 Ma across the Hikurangi Plateau and in the northern South Island may be the consequence of Hikurangi Plateau collision stalling within the Gondwana margin as well as detachment of the subducting slab (Hoernle et al., manuscript in preparation, 2008) and subsequent mantle upwelling. The mechanism associated with such a detachment would suggest a mantle source encroaching from south of the paleoconvergent Chatham Rise margin. Rebound uplift of the Chatham Rise paleoconvergent margin would be expected to follow such slab detachment in a similar manner to that proposed for the Timor margin [*Milsom and Audley-Charles*, 1986]. The eroded material from such an uplifted and aerially exposed margin will have formed much of the source material for unit MES. The diminishing thickness of unit MES sediment with distance from the Chatham Rise suggests that the unit is clastic sediment eroded from the uplifted Chatham Rise during and subsequent to subduction of the buoyant Hikurangi Plateau. The unit exhibits some disruption above the northern limit of the accre-

tionary prism that is absent in the same basin further north.

[68] The proposed ~75° change in plate motion at ~96 Ma [*Kroenke et al.*, 2004] is interpreted as coincident with a switch in orientation within the New Zealand region from northeast oriented extension to northwest oriented extension, orthogonal to the Wishbone Ridge orientation, but parallel to the eventual New Zealand–Antarctic breakup orientation at ~85 Ma [*Davy*, 2006b; *Cande et al.*, 1989].

5.2.2. Subduction Dynamics

[69] The ultimate extent of plateau subduction beneath both the North Island, and historically south beneath the Chatham Rise, is largely unknown. The leading edge of the subducting Hikurangi Plateau crust may be evident at ~65 km depths in high-resolution seismicity images from beneath the North Island [*Reyners et al.*, 2006]. Buoyancy analysis [*Cloos*, 1993] has indicated that subduction of basaltic crust greater than 18 km thick is dynamically unfavorable. The modern Hikurangi Plateau appears to be only marginally subductable beneath North Island, New Zealand [*Davy*, 1992]. The dynamics of Cretaceous Hikurangi Plateau subduction differ from modern plateau subduction partly because of the contrasting thermal state of the plateau. The seafloor depth is an approximate proxy for the thermal and compositional buoyancy of the underlying crust [*Smith et al.*, 1989]. The ~20 Ma old Hikurangi Plateau basement was 2000–3000 m deep at the time of plateau collision with the Chatham Rise, 500–1000 m shallower (and more buoyant) than today's 120 Ma old crust in the southwest Hikurangi margin. The seismic reflection and gravity data suggest however that, consistent with the *Cloos* [1993] analysis, the more buoyant 20 Ma old Hikurangi Plateau was not subductable.

[70] The Ontong-Java Plateau, which is typically up to 1000 m shallower than the Hikurangi Plateau (and ~35 km thick [*Gladczenko et al.*, 1997]), would by comparison with the Hikurangi margin choke the Solomon Islands Arc convergent margin. *Taira et al.* [2004] and *Phinney et al.* [2004] have imaged thrust faults ~7 km deep within Ontong-Java Plateau crust beneath the Malaita accretionary prism. These thrust faults are inferred to represent delamination of the Ontong-Java Plateau as it impinges on the Solomon Island Arc. Fragments of the Ontong-Java Plateau are also interpreted to have been off-scraped and exposed onshore on the

Solomon Islands. What is not so obvious from this analysis and the accompanying seismicity studies [*Miura et al.*, 2004] is how far the Ontong-Java Plateau extends down the subducting slab or whether just a tapered leading edge has been subducted. Our analysis of the Hikurangi Plateau margins favor the latter alternative.

[71] A number of factors other than buoyancy, such as plate convergence velocity and convergent margin obliquity, may have been crucial to the Gondwana margin subduction dynamics. The presence of a plume influence possibly contributing to Gondwana breakup [*Weaver et al.*, 1994; *Storey et al.*, 1999; Hoernle et al., manuscript in preparation, 2008] and widespread HIMU-type volcanism may also be among the important contributing factors for the switch from subduction to extensional margin formation in the New Zealand region of Gondwana.

5.3. Plateau Volcanism

5.3.1. Units HKB and VB

[72] Volcanism on the guyots in the Southern Hikurangi Plateau has, on the basis of $^{40}Ar/^{39}Ar$ dating of dredge samples, been dated at 99–89 Ma (Hoernle et al., manuscript in preparation, 2008) with similar dated volcanism interpreted to be widespread across the Hikurangi Plateau, Chatham Rise and onshore South Island New Zealand. Unit VB is interpreted as being continuous with volcanism within volcanic peaks of the Northern Volcanic Region and it is inferred that at least (an upper?) part of the volcanic peaks in this region were formed at the same time as VB was deposited. The above horizon correlation is however based upon seismic reflection data, such as line HKDC1, that cross guyots in the northern Hikurangi Plateau, whereas all the guyots dated by Hoernle et al. (manuscript in preparation, 2008) lie in the central and southern Hikurangi Plateau. It is not obvious on existing seismic lines whether the volcanism associated with unit VB is present south of the Southern Basement High (Figure 7) and hence whether it is incorporated into, or subducted beneath, the accretionary complex.

[73] All of the rocks dredged from ridge-like volcanoes or guyots on the Hikurangi Plateau were HIMU volcanic rocks in contrast to the thoeleiitic-to-mildly alkalic plateau basement samples (Hoernle et al., manuscript in preparation, 2008). This suggests that all the Hikurangi Plateau volcanoes were active between 99 and 89 Ma signifi-

cantly later than plateau formation. This contrasts with a plateau formation age from rocks at the northern Rapuhia Scarp of 118–94 Ma (Hoernle et al., manuscript in preparation, 2008). The ~30 Ma interval between plateau formation (~120 Ma) and cessation of the later 99–89 Ma period of volcanism most likely includes the entry of the Hikurangi Plateau into the North Chatham convergent margin and consequent jamming of subduction. Both Lines HKDC1 and HKDC3 (Figures 6 and 10) indicate that Unit HKB, at least adjacent to the North Chatham Slope, was deposited prior to the cessation of subduction in this margin. Much of unit HKB is interpreted as volcaniclastic sediment and part of the volcanic basement across the plateau deposited during plateau formation ~120 Ma. High seismic velocities (3200–3600 m/s) and comparison with the Ontong Java and Manihiki plateaus suggest however that unit HKB also includes a significant limestone and chert component.

[74] Timing models (e.g., Table 1) suggest the Hikurangi Plateau collided with the Chatham Rise at ~105–100 Ma immediately prior to the onset of 99–89 Ma episode of HIMU volcanism (Hoernle et al., manuscript in preparation, 2008). There is no evidence for the unit MES beneath unit VB suggesting that volcanism on the Hikurangi Plateau either predates or occurs in the early stage of uplift and erosion at the Chatham Rise margin. The base of Sequence Y is older than 65 Ma, and by extrapolation from ODP 1124, is interpreted to be ~70 Ma. The presence of MES sediment within one of the half-grabens on the northern Chatham Rise crest indicates that at least some of the extension forming the major half-grabens on the rise closely postdates subduction cessation. It is not clear how much of the variation in the present depth of unit VB is attributable to variations in the original depositional depth, how much is related to thinning associated with plateau breakup and how much (if any) is the result of a separate episode of postguyot formation subsidence.

5.3.2. Guyots and Ridge-Like Seamounts

[75] Two alternative end-member models emerge for guyot and ridge-like volcanism on the Hikurangi Plateau. In the first model (a) the guyots, ridge-like volcanics and units HKB and VB formed contemporaneously or soon after plateau formation/breakup with the guyots and ridge-like peaks being subject to renewed volcanism at ~99–89 Ma. In the second model (b) the guyots, ridge-like volcanics and possibly significant fractions of units HKB and VB formed at 99–89 Ma, possibly
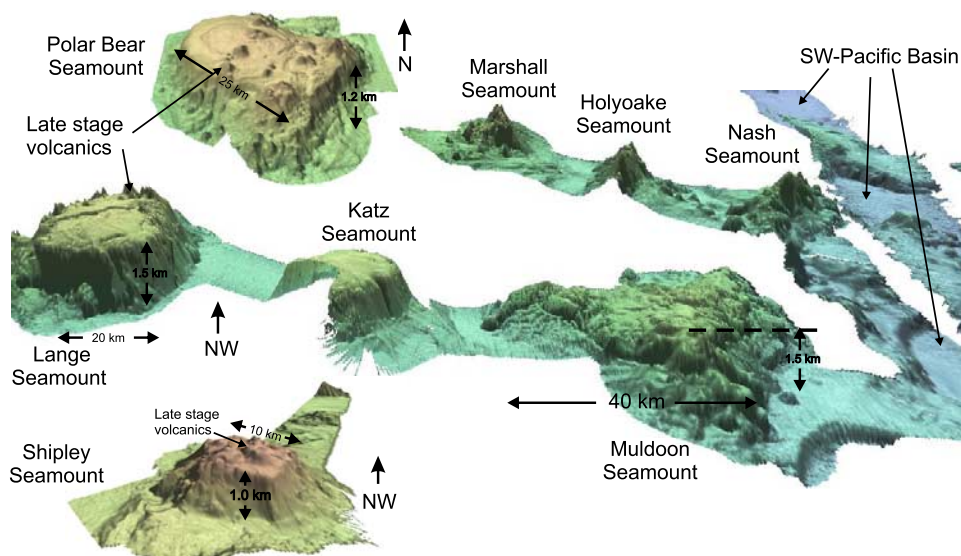
Figure 21. Perspective views of Guyot and Ridge-like seamounts on the Hikurangi Plateau. Locations are indicated in Figures 2 and 11.

contemporaneously with initial impact of the Hikurangi Plateau in the Gondwana convergent margin. Intermediate models are also possible. The uniform, distinctive nature of the HIMU geochemistry in volcanic rocks within the South Island, Chatham Rise and across all the Hikurangi Plateau dredge samples for the 99–89 Ma period lead us to favor the latter model above although the data (discussed below) are not conclusive.

[76] *Hoernle et al.* [2004] suggested the seamounts on the Hikurangi Plateau had formed on extensional faults related to the breakup of the combined Manihiki-Hikurangi Plateau. The simplest assumption, although not necessarily the only possibility, is that the ridge-like volcanism was synchronous with formation of these extensional faults. *Hoernle et al.* [2004] further commented that as the ridge-like seamounts were higher above the seafloor than guyot seamounts within the Hikurangi Plateau, the ridge-like seamounts must have formed postguyot formation and subsidence. However, in the Northern Volcanic Region the guyot seamounts (Palmer, Lange, and Katz (Figure 12)) all occupy a similar height range above the surrounding seafloor (1100–1300 m) to that of the ridge-like seamounts (1000–1300 m). Line HKDC1 and the other deep crustal seismic reflection lines do not intersect the ridge-style seamounts. On low-fold seismic reflection lines that do cross these seamounts (e.g., Katz and Muldoon Seamounts (Figure 17)) it is not possible to distinguish the seamount volcanism (ridge or guyot) from that of the interpreted horizon

VB suggesting that like the guyot seamounts, ridge-like seamounts may have formed synchronously with deposition of Unit VB. The ridge-like Muldoon Seamount is close to (<40 km), and of a similar elevation to, Katz Seamount (Figure 17). If Muldoon Seamount was significantly older than Katz Seamount it would be expected to have undergone greater subsidence. The lack of wave-cut erosion features on Muldoon Seamount suggests however that the ridge seamounts do not significantly predate the guyot seamounts.

[77] While dredge samples from the guyots in the Southern Hikurangi Plateau are dated in the range 99–89 Ma most of the mapped guyots exhibit evidence of minor late stage volcanism which sits atop the planed guyots (e.g., Lange Seamount and Polar Bear seamount (Figure 21)). It is difficult to know if dredge samples have sampled the main guyot edifice or these later volcanics. With the widespread similarity in distinct geochemistry of the 99–89 Ma volcanism it is possible, although unlikely, that all the volcanism sampled on the guyots and ridge-like seamounts is associated with such late stage peaks. This volcanism may result from the impact of the Hikurangi Plateau in the Gondwana margin and/or mantle/plume upwelling following subduction cessation and the ensuing detachment of the downgoing slab. As noted earlier, subduction in the North Chatham margin appears to have ceased following the subduction of some seamounts and subsequent to the deposition of Unit HKB in that region.

[78] There is considerable variation in the subsided depths of the guyot flat tops across the plateau from 1.6 to 3.0 km. Guyot height above interpreted volcanic basement varies on existing seismic data from 1.9 to 2.9 km. *Hoernle et al.* [2004] suggested increasing guyot subsidence with proximity to the Rapuhia Scarp however closer analysis and more recent data suggests this trend probably relates principally to the similar trend in decreasing crustal thickness toward the same margin. Guyot crest depth appears to be relatively constant within the crustal blocks (Figure 3) that make up the Hikurangi Plateau and is interpreted to be principally a function of the crustal thickness of the various blocks, with the thinner more extended blocks undergoing the greater subsidence. Much of the variation in LIP crustal block thickness will have occurred during formation, thermal uplift and/or breakup of the Ontong Java-Manihiki-Hikurangi Plateau at ~120 Ma. Compressive tectonics at ~100 Ma may also have been a factor in guyot height and subsequent subsidence. Such variation has, for instance, resulted in the crests of two guyots (Shipley Seamount and an unnamed guyot 100 km further east) at ~41.8°S (both within 50 km of the paleo-Gondwana margin) being ~1.6 and 3.0 km bsl, respectively versus crests of guyots of the northern volcanic region which lie at ~2.5 km bsl.

[79] Subsidence curves from Ontong-Java Plateau drill holes [*Ito and Clift*, 1998] indicate that seafloor depths for this possibly once-cojoined plateau have remained near 2–3 km over the last 120 Ma, at variance with a simple thermal subsidence model. The ~2 km depths on the Ontong-Java Plateau margins is consistent with guyot heights (1.9–2.9 km) forming at this depth, or deeper, on the Hikurangi Plateau. These guyot peaks now lie 1.6–3.0 km deeper. Similar subsidence has been inferred for ODP site 317 on the Manihiki Plateau [*Ito and Clift*, 1998]. The Hikurangi Plateau experienced renewed volcanism 20–30 Ma after formation complicating the subsidence history. Until further deep core samples are obtained from the Hikurangi Plateau the guyots remain the principal datable indicator of plateau depth.

[80] The similarity in height above volcanic basement for the eroded guyots and the ridge-like seamounts, combined with the inability to separate the volcanic stratigraphy of the two seamount types in existing seismic data is consistent with a similar age for both types of seamounts. Rifting apart of the OJMHP Large Igneous Province is likely to have been responsible for the block rotation faulting observed along the northeastern Hikurangi Plateau and at the core of at least some of the ridge-like seamounts. The linearity and orientation of the ridge-like seamounts perpendicular to the northeastern rift margin favors the interpretation of an extensional origin. Both ridge-like seamounts and guyot seamounts are also aligned in northeast bands (e.g., Figure 12). Most guyot seamounts are highly polygonal in map layout (Figures 12 and 21) with principal axis aligned either northeast or northwest. This geometry is inferred to mirror the basement grain (spreading ridge and fracture zone lines of weakness) of the Mesozoic oceanic crust which the plateau was deposited on. Both plateau formation/breakup and Gondwana collision processes at ~105–100 Ma could have activated such stress distributions.

6. Conclusions

[81] Seismic reflection data (principally line HKDC1) has revealed the volcanic, often volcaniclastic, nature of the Hikurangi Plateau basement. This basement is disrupted by extensional faulting interpreted as occurring during the breakup of the combined Ontong-Java/Manihiki/Hikurangi Plateau.

[82] The Hikurangi Plateau can be fitted back against the Manihiki Plateau prior to rifting apart soon after formation at ~120 Ma within a larger OJMHP super plateau LIP. The southeastern Manihiki Scarp in such a reconstruction matches against the eastern margin of the Rekohu Embayment, corresponding to a major structural break within the Hikurangi Plateau basement. Rotated block basement structure along the northeastern Hikurangi Plateau is consistent with the rifting apart of the two plateaus as is the partitioning of basement elevation and volcanic alignments that result from the breakup of the combined plateau.

[83] Fossil accretionary prism structures along the North Chatham Rise margin overlie the Hikurangi Plateau which has been subducted 50–100 km southward beneath the margin in the Late Mesozoic. A Mesozoic sedimentary unit (MES), up to 1.5 km thick at the base of the North Chatham Rise slope, is interpreted as having been formed by the uplift, above sea level, and erosion of the Chatham Rise following Hikurangi Plateau subduction. This unit which overlies both the North Chatham Rise and the Hikurangi Plateau and thins northward with distance from the Chatham Rise is interpreted

as infilling extensional grabens within the Chatham Rise basement. This unit overlies the North Chatham accretionary prism on line HKDC1 (178.5°W) but is incorporated in the accretionary complex further east on line HKDC3 (175.5°W) indicating variability in subduction cessation timing. Overlying the MES unit is a regionally observed, condensed, Late Cretaceous (~70 Ma)–Early Oligocene (32 Ma), high-amplitude sequence of reflectors, Sequence Y [*Wood and Davy*, 1994]. Sequence Y postdates deformation across most of the region.

[84] The entry of the Hikurangi Plateau into the Gondwana margin at ~105 Ma is interpreted to have led to the subsequent jamming, by the buoyant plateau crust, of the subduction system by ~100 Ma. The contrast between this interpreted choking of the Gondwana subduction system and the subduction of the present-day Hikurangi Plateau beneath North Island suggest a buoyancy cutoff range for LIP subduction consistent with buoyancy modeling [*Cloos*, 1993]. Gravity modeling of the plateau indicates a crustal thickness between 15 and 23 km dependent upon the density distribution assumed. At present there is no deep-crustal velocity information to constrain these models. Regardless of the model chosen the plateau thins by ~one third beneath the Chatham Rise. Subduction extent beneath the Gondwana margin and the timing of subduction cessation is interpreted to be variable along the margin. The Hikurangi Plateau is interpreted to have been subducted a shorter distance (~50 km) and subduction cessation is interpreted as occurring earlier (syn-MES deposition) west of the Rekohu embayment. The transfer of north-south extension, previously occurring at the Osbourn Trough, through the Gondwana convergent margin via strike-slip motion on the Wishbone Ridge Fault Complex, into the back-arc Gondwana region is interpreted to have continued until ~96 Ma. At the end of this period the subducted Pacific Plate slab, bypassed for relative plate motion in the Hikurangi Plateau region, has probably become detached and sunk into the mantle. Plate motion rearrangement following such a slab detachment has led, at ~96 Ma, to uplift along the Chatham Rise, NW-SE oriented extension in the New Zealand region and eventual New Zealand–Antarctic rifting at ~85 Ma.

[85] Seamount volcanism (99–89 Ma) occurred across the Hikurangi Plateau, Chatham Rise and within the South Island of New Zealand during and after Hikurangi Plateau collision with the convergent margin and the postulated slab detachment and change in the extensional orientation within the New Zealand region. Although the HIMU-type alkalic seamount volcanism is compositionally distinct from the EM-type tholeiitic plateau basement (sampled at the Rapuhia Scarp rifted margin), the seamount volcanism is not distinguishable on seismic reflection records from the volcanism interpreted to date from plateau formation at ~120 Ma. This may be a consequence of the 99–89 Ma seamount volcanism occurring as localized sources possibly reactivating zones of basement weakness particularly at the ridge-like volcanoes and guyots of the plateau that formed at ~120 Ma prior to 88–99 Ma reactivation. The absence of samples similar to the plateau basement geochemistry in any of the sampled seamounts favors, however, an interpretation where the seamounts on the Hikurangi Plateau formed at ~99–89 Ma.

Acknowledgments

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