The role of protothrusts in frontal accretion and accommodation of plate convergence, Hikurangi subduction margin, New Zealand

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ABSTRACT

Protothrusts mark the onset of deformation at the toe of large subduction accretionary wedges. They are recognized in seismic reflection sections as small-displacement (tens of meters) faults seaward of the primary frontal thrust fault. Although assumed to reflect incipient accretionary deformation and to mark the location of future thrusts, few studies discuss their displacement properties, evolution, and kinematic role during frontal accretion and propagation of the subduction décollement.

We analyze high-quality geophysical and bathymetric images of the spectacular 25-km-wide Hikurangi margin protothrust zone (PTZ), the structure of which varies along strike north and south of the colliding Bennett Knoll seamount. We provide a quantitative data set on protothrust scaling relationships and fractal fault population characteristics. Our analyses lead us to speculate on the importance of stratigraphic heterogeneity in structural development, and highlight the role of protothrust arrays in the formation of the frontal thrust. We document a migrating wave of protothrust activity in association with forward advancement of the décollement and deformation front. Shortening east of the present frontal thrust, calculated from displacements on seismically imaged faults and from subseismic faulting derived from power law relationships, reveal the significant role of the PTZ in accommodating shortening. There is possibly as much as ~7.4 km and ~4.0 km of shortening with forward advancement of the décollement and deformation front. Shortening east of the present frontal thrust, calculated from displacements on seismically imaged faults and from subseismic faulting derived from power law relationships, reveal the significant role of the PTZ in accommodating shortening. There is possibly as much as ~90% of the total strain is observable as fault displacements in seismic reflection sections, implying that protothrust zones are characterized by significant diffuse deformation at scales below the vertical and horizontal resolution of airgun-source seismic sections. Protothrust zones also appear to define regions of tectonic compaction prior to the onset of thrusting (Bray and Karig, 1986). Lateral changes in porosity derived from seismic velocities have been used to estimate the amount of tectonic shortening across the protothrust zone (PTZ) and frontal thrusts at Nankai Trough (e.g., Morgan et al., 1994; Morgan and Karig, 1995; Moore et al., 2011) and Cascadia (Adam et al., 2004). Such analyses suggest that <20% of the total strain is observable as fault displacements in seismic reflection sections, implying that protothrust zones are characterized by significant diffuse deformation at scales below the vertical and horizontal resolution of airgun-source seismic sections.

Mesoscale and microscale structures also have been observed in drill cores from the Nankai Trough PTZ, and are thought to accommodate this diffuse shortening. They include features variously described as kink bands, shear bands, diffuse deformation, and fractal fault population characteristics. Our analyses lead us to speculate on the importance of stratigraphic heterogeneity in structural development, and highlight the role of protothrust arrays in the formation of the frontal thrust. We document a migrating wave of protothrust activity in association with forward advancement of the décollement and deformation front. Shortening east of the present frontal thrust, calculated from displacements on seismically imaged faults and from subseismic faulting derived from power law relationships, reveal the significant role of the PTZ in accommodating shortening. There is possibly as much as ~7.4 km and ~4.0 km of shortening with forward advancement of the décollement and deformation front. Shortening east of the present frontal thrust, calculated from displacements on seismically imaged faults and from subseismic faulting derived from power law relationships, reveal the significant role of the PTZ in accommodating shortening. There is possibly as much as ~90% of the total strain is observable as fault displacements in seismic reflection sections, implying that protothrust zones are characterized by significant diffuse deformation at scales below the vertical and horizontal resolution of airgun-source seismic sections. Protothrust zones also appear to define regions of tectonic compaction prior to the onset of thrusting (Bray and Karig, 1986). Lateral changes in porosity derived from seismic velocities have been used to estimate the amount of tectonic shortening across the protothrust zone (PTZ) and frontal thrusts at Nankai Trough (e.g., Morgan et al., 1994; Morgan and Karig, 1995; Moore et al., 2011) and Cascadia (Adam et al., 2004). Such analyses suggest that <20% of the total strain is observable as fault displacements in seismic reflection sections, implying that protothrust zones are characterized by significant diffuse deformation at scales below the vertical and horizontal resolution of airgun-source seismic sections. Protothrust zones also appear to define regions of tectonic compaction prior to the onset of thrusting (Bray and Karig, 1986). Lateral changes in porosity derived from seismic velocities have been used to estimate the amount of tectonic shortening across the protothrust zone (PTZ) and frontal thrusts at Nankai Trough (e.g., Morgan et al., 1994; Morgan and Karig, 1995; Moore et al., 2011) and Cascadia (Adam et al., 2004). Such analyses suggest that <20% of the total strain is observable as fault displacements in seismic reflection sections, implying that protothrust zones are characterized by significant diffuse deformation at scales below the vertical and horizontal resolution of airgun-source seismic sections.

INTRODUCTION

Protothrusts represent incipient deformation in clastic trench sequences at the toe of subduction accretionary wedges (Fig. 1). They are most notable at

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ductile shear zones, deformation bands, and brittle faults (e.g., Moore et al., 1986; Lundberg and Moore, 1986; Karig and Lundberg, 1990; Maltman et al., 1993). Recent analyses of cores identified polyphase deformation indicative of different deformation mechanisms during the evolution of the PTZ. For example, relatively high angle (≥45°), early stage ductile deformation bands are overprinted by lower angle (<45°) faults (e.g., Ujiie et al., 2004; Morgan et al., 2007). These high-angle structures are now generally interpreted as compactive shear bands developed in association with porosity reduction, fabric collapse, and dewatering (Bésuelle, 2001; Ujiie et al., 2004; Morgan et al., 2007). The localization of compaction and shear may be facilitated by the collapse of intergranular bonding or cementation, leading to rapid failure, and transient increases in pore pressures followed by consolidation. The lower angle cross-cutting faults in Nankai Trough cores are preferentially developed in areas of the PTZ that show higher strain, and are consistent with brittle Mohr-Coulomb failure (Karig and Lundberg, 1990; Morgan et al., 1994, 2007). PTZs are therefore complex structural systems characterized by deformation recorded at a variety of scales. To date, however, the relationships between the core-scale and seismic-scale structures have not been established. Moreover, their kinematic role in the evolution of frontal thrust development and accretion remain poorly understood.

The spectacular, seismically imaged PTZ at the front of the Hikurangi accretionary wedge, New Zealand, reaches 25 km in width (Fig. 2) (Davey et al., 1986; Lewis and Pettinga, 1993; Barnes and Mercier, 1997; Barnes et al., 2010; Ghisetti et al., 2016). Previous studies of the PTZ described its general characteristics and spatial extent, recognizing that it coincides with the widest part of the accretionary wedge. Ghisetti et al. (2016) characterized the dip distributions of these faults, interpreted an older series of PTZs now accreted into the wedge, and reconstructed their spatial development associated with forward growth of the wedge over the past ~2 m.y. Despite consistent seaward vergence of the major thrusts (Fig. 2B), the predominant dip of the PTZ varies along strike (e.g., Barnes et al., 2010). While many have discussed the possible factors that contribute to fault vergence in accretionary wedges worldwide (Davis et al., 1983; Byrne and Hibbard, 1987; Dahlen, 1990; MacKay et al., 1992; Byrne et al., 1993; Lallemand et al., 1994; MacKay, 1995; Bonini et al., 2000; Gutscher et al., 2001; Underwood, 2002; Smit et al., 2003; Adam et al., 2004; Buiter et al., 2006; Zhou et al., 2007; Moeremans et al., 2014; Cubas et al., 2016), the cause of along-strike variation in the predominant dip of the Hikurangi PTZ remains unclear and is beyond the scope of this study.

Available seismic images offer the opportunity to further characterize the Hikurangi PTZ and improve understanding of its kinematic role. In this study we make use of excellent imaging of the PTZ in seismic reflection profiles and in new 30 kHz multibeam bathymetric data. We focus on the active PTZ east of the primary frontal thrust (Fig. 2B), provide the first substantial quantitative data on protothrust population attributes, and discuss their role in thrust fault growth processes. We discuss the role of protothrusts in frontal accretion, including their kinematic response to trenchward propagation of the plate interface décollement and frontal thrust. A conceptual model of deformation evolution is presented that may be applicable to other subduction margins with similar characteristics. We estimate the tectonic shortening accommodated by the active PTZ east of the frontal thrust from measurements of seismically imaged fault displacements and estimates of subseismic-scale faulting derived from power law relationships. These data, combined with consideration of PTZs landward of the frontal thrust, allow us to evaluate the surprisingly important role of the protothrust system in the accommodation of regional plate convergence at the central Hikurangi margin. We compare our estimates of tectonic shortening with results from some other similar subduction margins.

**HIKURANGI SUBDUCTION, SEDIMENTARY SEQUENCE, AND ACCRETIONARY WEDGE**

The Hikurangi margin is located above the subducting Pacific plate along the eastern margin of North Island, New Zealand (Fig. 2A). The vector of relative motion between the Pacific and Australian plates is oblique to the margin and decreases in magnitude from ~48 mm/yr in the north off the Raukumara Peninsula to ~43 mm/yr offshore of northern Wairarapa (Beavan et al., 2002). These changes are accompanied by backarc extension in the Taupo Volcanic Zone and clockwise rotation of the forearc region. The combined effects of these result in forearc margin-normal tectonic shortening decreasing from close to 60 mm/yr in the north offshore of the Raukumara Peninsula (Fig. 2A), to ~30–50 mm/yr in the study area offshore of northern Wairarapa (Fig. 2B), and close to 20 mm/yr offshore of southern Wairarapa (Wallace et al., 2004).

The Pacific plate subducting beneath North Island comprises the oceanic Hikurangi Plateau, a large igneous province with crustal thickness of ~12–15 km (Davey et al., 2008). The plateau is studded by scattered seamounts, more common offshore Hikurangi, but including Bennett Knoll seamount in the south (Fig. 2B). The lithostratigraphy of the plateau cover sequence has been interpreted from seismic sections tied to borehole data [Ocean Drilling Program (ODP) Site 1124; Shipboard Scientific Party, 1999] east of the trench (Davey et al., 2008), and correlated to the Hikurangi Trough where there are currently no borehole data available (Barnes et al., 2010). The plateau cover sequence is thought to contain a lower unit of possible Cretaceous (100–70 Ma) clastic sedimentary rocks (as yet unsampled), overlain by a condensed sequence of Late Cretaceous–early Cenozoic (70–32 Ma) pelagic chalks, nannofossil oozes, and mudstones, and late Cenozoic (mainly Miocene) nannofossil oozes, mudstones, and tephra. Beneath the 2500–3500-m-deep Hikurangi Trough, at the front of the accretionary wedge, the plateau cover sequence is overlain by an ~1–6-km-thick onlapping wedge of Pleiocene–Pliocene turbidites associated with the evolution of the Hikurangi Channel (Lewis and Pettinga, 1993; Barnes and Mercier de Lépinay, 1997; Lewis et al., 1998, 2002; Lewis and Pantin, 2002; Barnes et al., 2010; Plaza-Faverola et al., 2012; Bland et al., 2015).
Figure 2. (A) Regional tectonic setting of the Hikurangi subduction zone, North Island, New Zealand. White arrows are relative Pacific-Australian plate motion vectors (Beavan et al., 2002). TVZ—Taupō Volcanic Zone; RP—Raukumara Peninsula. (B) Structure and morphology of the central Hikurangi margin accretionary wedge. Yellow shaded areas include the currently active protothrust zone largely southeast of the primary frontal thrust (modified from Barnes et al., 2010). The north protothrust zone (PTZ) and the south PTZ are separated by the Bennett Knoll seamount (bold dashed purple lines represent the base of seamounts on the subducting Pacific plate; modified from Barnes et al., 2010). Bold black arrow is the Pacific-Australian plate motion vector. Bold numbers along the deformation front are margin-perpendicular convergence rates (mm/yr) from Wallace et al. (2004). Thrust faults labeled F11—F18 were discussed in Ghisetti et al. (2016). Blue contours are water depths (m). Bold red lines are seismic sections illustrated in Figures 3–5.
Pettinga, 1993; Barnes and Mercier de Lépinay, 1997; Barker et al., 2009; Barnes et al., 2010; Ghisetti et al., 2016; Plaza-Faverola et al., 2016). The low taper has been considered to be consistent with high pore pressures and/or a plate interface that is weak relative to the wedge (e.g., Wallace et al., 2009; Fagereng, 2011; Skarbek and Rempel, 2017). The décollement beneath the frontal wedge has a regional dip of −2°–3.5°W and is mainly developed in the upper part of the pelagic cover sequence, >5–6 km below sea level.

Upper plate thrust faults at the Hikurangi margin are predominantly northwest dipping (seaward verging), and within the frontal wedge are typically characterized by maximum displacements of 0.8–1.5 km (Barnes et al., 2010; Ghisetti et al., 2016). Active fluid seepage is widespread on the crests of thrust hanging-wall anticlines, particularly along the middle and upper slope, attesting to fault-controlled fluid flow (Lewis and Marshall, 1996; Greinert et al., 2010; Barnes et al., 2010; Pecher et al., 2010; Plaza-Faverola et al., 2014). Contemporaneous thrust faulting, folding, and sedimentation has resulted in growth sedimentary sequences in slope basins and on the limbs of anticlinal ridges (Lewis and Pettinga, 1993; Barnes and Mercier de Lépinay, 1997; Pecher et al., 2010; Barnes et al., 2010; Plaza-Faverola et al., 2012, 2014; Ghisetti et al., 2016). The timing of the onset of growth stratigraphy combined with restoration of fault displacement and folding was used in Ghisetti et al. (2016) to progressively reconstruct the propagation of the frontal wedge; from the analysis, they found that the accretionary wedge has widened and advanced seaward by as much as 60 km in the past −2.0 ± 0.8 m.y., and that major thrust faults accommodate not more than 20%–45% of the total predicted margin-normal tectonic shortening. Accreted protothrust zones were recognized in thrust-bound panels as much as 35 km landward of the deformation front.

In this paper we present a structural analysis of the youngest, active PTZ beneath the basin floor east of the primary frontal thrust.

The major thrust faults are associated with bathymetric ridges that have relief of ∼500 m and length of ∼30 to >100 km (Fig. 2B). The active PTZ in the Hikurangi Trough is as much as 25 km wide, and extends 230 km along strike. The PTZ is split into two parts by Bennett Knoll seamount, the northern part of which is underthrusting the deformation front (Barnes et al., 2010; Ghisetti et al., 2016). In this study, we refer to these separate regions of the PTZ as the south PTZ and north PTZ, respectively.

## DATA AND ANALYTICAL METHODS

### Multibeam Bathymetric Data

We present new Kongsberg EM302 multibeam echo-sounder data collected by the National Institute of Water and Atmospheric Research (NIWA) in 2015 on R.V. Tangaroa survey TAN1510. This echo sounder operated at a frequency of 30 kHz, with an angular sector of 140°, across track coverage of as much as 5 times water depth, and maximum ping rate of 10 Hz. The transmit beams were electronically stabilized for roll, pitch, and yaw, and the received beams were stabilized for roll movements. Sound velocity profiles were obtained with an AML Oceanographic Smart Probe P0835 system (https://aml-oceanographic.com/) to calibrate the multibeam echo-sounder data. An Anplaxin (https://www.applanix.com/) POS MV V4 system provided latitude, longitude, velocity, heading, pitch, roll, and heave data. The multibeam data were processed using CARIS HIPS and SIPS V8.1.12 software (http://www.caris.com/products/hips-sips/), and were tidally corrected using an area-specific tidal curve generated from a NIWA tidal model, with all soundings reduced to mean sea level. The final processed bathymetric data were used to construct a 30 m grid resolution digital elevation model (DEM), which was imported into ArcGIS V.10.3 (desktop.arcgis.com/en/arcmap/) for spatial analysis and mapping of tectonic geomorphology.

### Seismic Reflection Data and Stratigraphy

We utilize the eastern parts of the same seismic reflection profiles used in Ghisetti et al. (2016) for structural analysis and restoration. Profile 05CM-38 joined to SO191-4 crosses the frontal wedge and south PTZ immediately west of Bennett Knoll (Figs. 2B and 3); profile SO191-6 crosses the south PTZ a further 28 km to the southwest (Figs. 2B and 4); profile SO191-1 crosses the frontal wedge and the center of the north PTZ (Figs. 2B and 5); these sections are oriented 110°, 130°, and 120°, respectively, −17°, 3°, and 7° oblique to the inferred shortening direction normal to the major frontal thrust faults (127°).

Profile 05CM-38 was acquired by GNS Science and the New Zealand Government in 2005 with a 4140 in³ source array and 12-km-long hydrophone streamer (960 channels) (Barker et al., 2009). This profile is mainly west of the south PTZ (Fig. 3), and was reprocessed by Plaza-Faverola et al. (2016) using a pre-stack depth migration (PSDM) workflow. The SO191 profiles were acquired on R.V. Sonne in 2007 using a 2080 in³ source volume and short (32 channel) hydrophone streamer, and processed as post-stack time migrated sections with common depth point interval of 12.5 m (e.g., Barnes et al., 2010). These sections were depth converted in Ghisetti et al. (2016) using velocity data from the 05CM-38 PSDM.

We adopt the same seismic reflection stratigraphy as in Ghisetti et al. (2016), including horizons R0–R7, essentially based on that in Barnes et al. (2010) and modified by Plaza-Faverola et al. (2012). Reflection R7 is thought to mark the top of a sequence of inferred Late Cretaceous–early Oligocene (70–32 Ma) pelagic chalks, nanofossil oozes, and mudstones. This horizon coincides with the basal décollement south of Bennett Knoll seamount (Figs. 3 and 4). Reflection RSB is a prominent erosion and onlap surface that has been interpreted to mark the top of the Hikurangi Plateau pelagic cover sequence of inferred early Pliocene or late Miocene age (older than 3.5 Ma). This surface coincides locally with the basal décollement north of Bennett Knoll seamount (Fig. 5). Reflections R5, R4–R5, R4, and R3 mark four prominent horizons within the overlying turbidite wedge, which is ∼3 km thick at the deformation front in this area. Although not validated by borehole data,
Figure 3. Part of depth-converted seismic reflection profiles 05CM-38 adjoined to SO191-4 across the frontal part of the central Hikurangi accretionary wedge, including south protothrust zone (PTZ), and the western margin of Bennett Knoll seamount (modified from Ghisetti et al., 2016). (A) Uninterpreted enlargement with vertical exaggeration, VE = 2, of the area indicated in B. (B) Interpreted seismic section with VE = 2. (C) Cross section with no vertical exaggeration based on data in B. Seismic stratigraphy is from Ghisetti et al. (2016), modified from Barnes et al. (2010) and Plaza-Faverola et al. (2012). F13–F18 identify thrust faults traced between seismic sections using seafloor morphology, as shown in Figure 2B. (D) Legend for A–C. H is horizontal.
Figure 4. Part of depth-converted seismic reflection profile SO191-6 across the frontal part of the central Hikurangi accretionary wedge, including south protothrust zone (PTZ). (A) Uninterpreted enlargement with vertical exaggeration, VE = 2, of the area indicated in B. (B) Interpreted seismic section with VE = 2. (C) Cross section with no vertical exaggeration based on data in B (from Ghisetti et al., 2016). Note that Ghisetti et al. (2016) smoothed reflector R7, top of Pacific plate, and basal décollement, compared to B, to account for velocity artifacts related to depth conversion and to facilitate progressive structural restorations with decompaction. F15–F17 identify thrust faults traced between seismic sections using seafloor morphology, as shown in Figure 2B. (D) Legend for A–C. H is horizontal.
Figure 5. Part of depth-converted seismic reflection profile SO191-1 across the frontal part of the central Hikurangi accretionary wedge, including north protothrust zone (PTZ). (A) Uninterpreted enlargement with vertical exaggeration, VE = 2, of the area indicated in B. (B) Interpreted seismic section with VE = 2. (C) Cross section with no vertical exaggeration based on data in B (from Ghisetti et al., 2016). Note that Ghisetti et al. (2016) smoothed reflectors R5B, R6, R7, top of Pacific plate, and basal décollement, compared to B, to account for velocity artifacts related to depth conversion and to facilitate progressive structural restorations with decompaction. F12–F16 identify thrust faults traced between seismic sections using seafloor morphology, as shown in Figure 2B. The green marker in the R4-R5 sequence is used in our analysis of tectonic shortening. (D) Legend for A–C. H is horizontal.
the ages of reflections R5, R4, and R3 have been inferred, from correlation to dated seafloor samples in the southern reaches of the margin, as 2.0 ± 0.8 Ma, 1.0 ± 0.5 Ma, and 0.6 ± 0.2 Ma, respectively (Barnes and Mercier de Lépinay, 1997; Barnes et al., 2010; Plaza-Faverola et al., 2012; Ghisetti et al., 2016).

Based on the velocity structure (Plaza-Faverola et al., 2016; Ghisetti et al., 2016) and frequency content we estimate the vertical resolution ($\lambda/4$, where $\lambda = V/F$; $\lambda$ is seismic wavelength, $V$ is velocity, and $F$ is frequency) of the SO191 profiles to range from ~5 to 10 m for the stratigraphic interval between reflectors R3 to R4, ~7 to 17 m for the interval between reflectors R4 to R5, and ~14 to 20 m for the interval between reflectors R5 to R7 (largely below the PTZ).

**Structural Analysis**

We interpreted the depth sections using Move software (Midland Valley; https://www.mve.com/software), and captured geometric and structural data in spreadsheets for statistical analysis. We present a range of structural measurements, which are defined in Figure 6. These include fault dip, length, width, spacing, heave, and separation. The fault length was measured directly in map view (e.g., Fig. 7), while all other parameters, including the apparent fault spacing, were measured on the seismic profiles. The fault width is the downdip distance measured along the fault trace in the section. The net reverse displacements on the faults (also known as the slip) cannot be measured on the two-dimensional profiles, because the slip vectors on the faults remain unknown relative to the profile orientations. Separation is the correct term for the apparent displacement, and is defined as the distance measured in the plane of the section between the fault footwall and hanging-wall cutoffs of the same horizon (e.g., Hobbs et al., 1976; Groshong, 1999; Davis et al., 2012).

**RESULTS OF THE ANALYSIS**

### South PTZ

**Structural Geometry, Separation Data, and Scaling Relationships**

The seismic sections reveal a densely spaced array of protothrusts west and south of Bennett Knoll seamount (Figs. 2B, 3A, 3B, 4A, and 4B), where the south PTZ is 25 km wide (Figs. 7C, 7D), narrowing southwest to <5 km east of Aorangi Ridge (Fig. 2B). The protothrusts are well imaged as dipping lineaments of weak reflectivity coinciding with reverse offsets of seismic reflections. They are preferentially concentrated in the turbidite sequence above R5B; the densest part of the array is visible between reflection horizons R5 and R3 (Figs. 3, 4, and 8A–8D). The faults do not produce fault-plane reflections, but local bright spots in seismic reflections in the immediate hanging wall of some protothrusts (Figs. 8A, 8C) likely indicate fluid and/or gas migration up the faults. They dip predominantly to the northwest (Figs. 8A and 9A), and a subordinate set dips southeast, forming conjugate structures (Figs. 8I, 8J, and 9C). The majority of the protothrusts have an apparent dip (i.e., measured in the plane of the sections; Fig. 6C) of 40°–50° (Table 1) (Ghisetti et al., 2016). These dips are similar to those of the upper sections of the major frontal thrust faults. A plot of protothrust dip measured on faults offsetting horizon R4-R5 versus distance east of thrust fault F16 (red curve in Fig. 10) reveals that dips lower than average were measured within the regions as much as 3 and 6 km east of thrust faults F16 and F17, respectively.

The multibeam bathymetric data reveal geomorphology associated with the PTZ on the basin floor of the Hikurangi Trough (Figs. 7C, 7D). Linear scarps and ramps related to monoclinal folds confirm that the protothrusts strike sub-parallel to the deformation front (Barnes and Mercier, 1997; Lewis et al., 1998).
Figure 7. (A) Interpretation of fault structure across the deformation front in the north protothrust zone (PTZ) on a background hillshade image derived from Kongsberg EM302 multibeam bathymetry data (30 m grid digital elevation model). Contours in blue are water depths (m). White dashed lines indicate the base of the Bennett Knoll seamount buried beneath Hikurangi Trough sediments, identified from seismic sections. Locations of the maps are shown in Figure 2B. (B) Uninterpreted bathymetry of corresponding area in A. (C) Interpretation of fault structure across the deformation front in the south PTZ. (D) Uninterpreted bathymetry of corresponding area in C.
Figure 8 (on this and following page). Interpreted and corresponding uninterpreted depth-converted seismic sections with no vertical exaggeration, illustrating structural details of the active protothrust zone. Color coding of fault structures reflect apparent fault dips in the plane of the sections. (A–D) Northwest-dipping array in the south protothrust zone (PTZ), southwest of Bennett Knoll seamount. V—vertical, H—horizontal. (E, F) Southeast-dipping array in the north PTZ, north of Bennett Knoll. (G, H) Conjugate protothrusts in the north PTZ. (I, J) Conjugate protothrusts in the south PTZ. Section locations are shown in Figures 4 and 5. Seismic stratigraphy as in Figures 3–5.
Conjugate protothrusts

Stratigraphic horizons
- R3: 0.6 ± 0.2
- R4: 1.0 ± 0.5
- R4-R5: 2.0 ± 0.8
- R5: > 3.5
- R6: 15 ± 5

Inferred age (Ma)
- 2.0 ± 0.8
- 1.0 ± 0.5
- 0.6 ± 0.2
- > 3.5

Fault dips (°)
- 20 - 30
- 30 - 40
- 40 - 50
- 50 - 60

Figure 8 (continued).
Table 1. Structural parameters associated with the Hikurangi South and North protothrust zones

<table>
<thead>
<tr>
<th>Parameter</th>
<th>South protothrust zone</th>
<th>North protothrust zone</th>
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<tbody>
<tr>
<td>Protothrust zone width (km) (Fig. 2)</td>
<td>25</td>
<td>25</td>
</tr>
<tr>
<td>Protothrust geomorphic length in multibeam echo sounder data (km) (Figs. 6 and 7)</td>
<td>Range 4–24</td>
<td>4–28</td>
</tr>
<tr>
<td>Mean 12</td>
<td>11</td>
<td></td>
</tr>
<tr>
<td>Protothrust dip ('') (Ghisetti et al., 2016) (Fig. 6)</td>
<td>Range 20–70</td>
<td>20–70</td>
</tr>
<tr>
<td>&gt;50% measurements</td>
<td>20–70</td>
<td>20–70</td>
</tr>
<tr>
<td>Protothrust spacing across protothrust zone strike (i.e., map view) (m) (Figs. 6 and 11)</td>
<td>Frequency modes on horizons R4 and R5 combined</td>
<td>100–200 100–200</td>
</tr>
<tr>
<td>Mean 12</td>
<td>11</td>
<td></td>
</tr>
<tr>
<td>Horizon R4 Frequency modes on horizons R4 and R5 combined</td>
<td>Range 50–2800</td>
<td>50–4960</td>
</tr>
<tr>
<td>Median 320</td>
<td>290</td>
<td></td>
</tr>
<tr>
<td>Standard deviation 450</td>
<td>760</td>
<td></td>
</tr>
<tr>
<td>Horizon R5 Median 670</td>
<td>820</td>
<td></td>
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<tr>
<td>Standard deviation 650</td>
<td>800</td>
<td></td>
</tr>
<tr>
<td>Maximum separation measured on individual protothrusts (m) (Figs. 6, 13, and 15)</td>
<td>Range 3–233</td>
<td>3–137</td>
</tr>
<tr>
<td>Mean 15</td>
<td>12</td>
<td></td>
</tr>
<tr>
<td>Median 10</td>
<td>10</td>
<td></td>
</tr>
<tr>
<td>Standard deviation 19</td>
<td>8</td>
<td></td>
</tr>
<tr>
<td>Protothrust width (m) (Figs. 6 and 15)</td>
<td>Range 40–5500</td>
<td>75–2550</td>
</tr>
<tr>
<td>Mean 670</td>
<td>500</td>
<td></td>
</tr>
<tr>
<td>Median 350</td>
<td>290</td>
<td></td>
</tr>
<tr>
<td>Standard deviation 760</td>
<td>480</td>
<td></td>
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</table>
The seismic sections show that geomorphic lineaments are associated with either singular protothrusts or closely spaced clusters of protothrusts, some arranged in vertically segmented and left-stepping geometries (e.g., Figs. 8A–8D, 9D, and 9E). The geomorphic lineaments have lengths ranging from 4 to 24 km (Table 1). We also recognize two linear features interpreted here as left-lateral strike-slip faults, crossing the deformation front and the south PTZ southwest of Bennett Knoll seamount (Figs. 7C, 7D). These newly discovered features will be presented in detail elsewhere and are not discussed further herein.

The seismic sections reveal a large range in protothrust spacing across the strike of the south PTZ (Figs. 3 and 4). The frequency of occurrence of fault spacing measured along each of the two southern seismic sections is presented in Figure 11 (green bars). Values are corrected for the obliquity of each profile (i.e., ~17° and 3°) relative to the vector normal to the frontal thrusts F16 and F17 (i.e., 127°). The predominant frequency modes are 100–200 m and 300–450 m spacing on horizons R5 and R4. The spacing appears to increase over ~1 km depth in the Pliocene–Pleistocene turbidite sequence (from R4 to R6) (Table 1), and there is a general increase in spacing in the outer reaches of the PTZ.

Notwithstanding differences in the vertical seismic resolution (Δ/4) with depth, reverse separations on the protothrusts are measurable in the sections to about half a seismic wavelet (e.g., Fig. 12). In Figure 13 we plot the apparent maximum separation measured on individual protothrusts against the frequency of observations, with measurements corrected for the obliquity of each profile relative to the inferred shortening direction normal to the frontal thrusts. Note, however, that these corrected values are apparent maximum separations, because in map view the profiles do not necessarily cross each structure at the location of its maximum displacement. The data, excluding the incipient thrust fault F17 (Fig. 3), show that the apparent maximum separation in the south PTZ ranges from ~3 to 233 m; ~50% of measurements are <10 m and 98% are <70 m (Table 1).

The majority of protothrusts visible in the seismic sections appear to be largely confined to the Pliocene–Pleistocene turbidite section (R0 to R5B) and do not penetrate down-dip into the inferred pelagic sequence below horizon R5B, i.e., into the sequence hosting the outer wedge décollement (e.g., Figs. 3, 4, and 8A–8D). While this may reflect some loss of vertical seismic resolution, horizon R5B scarcely shows deformation by either reverse faulting or folding across the PTZ. Moreover, detailed measurements of separation down the dip of the visible protothrusts reveal a reduction in separation toward their apparent lower tip (Figs. 14C, 14F). Normalized plots of the separation down the dip of the protothrusts are presented in Figures S1D and S1H. Some structures show a slightly asymmetric decay with a maximum separation on individual structures toward their apparent lower tip.

The measured protothrust widths (Fig. 6C) range from 40 to 5500 m, with a median value of ~350 m (Table 1). These measurements may not all represent true maximum widths because the profiles may not cross each protothrust at the location of its largest width dimension. Nevertheless, the larger widths measured are commonly associated with protothrusts within
Figure 12. Protothrust separation measurements (Fig. 6B) made in seismic sections. (A) Illustration of measurable offsets of negative and positive polarity seismic wavelets. (B, C) Examples of protothrusts in depth-converted seismic data, showing enlargements with reflection offsets (m) highlighted by yellow lines. VE—vertical exaggeration.

Figure 13. Frequency distributions of the maximum separation (Fig. 6C) on individual protothrusts interpreted in the seismic sections across the north protothrust zone (PTZ; number of measurements, n = 229) and south PTZ (n = 285). The separation measurements were corrected for the obliquity of each profile relative to the vector normal to the frontal thrusts F16 and F17 (i.e., 127°).
Figure 14. Separation data plotted against distance along the fault (referred to elsewhere as distance-displacement plots; e.g., Williams and Chapman, 1983; Hughes and Shaw, 2015) for the frontal thrust fault, incipient thrust, and selected (bold) protothrusts in each of the three seismic sections. Note that for comparative purposes, the measurements are made on protothrusts representing the predominant protothrust zone (PTZ) fault dip across each section. (A) Section SO191-6. (B, C) Data for faults highlighted on section shown in A. (D) Composite section 05CM-38–SO191-4. (E, F) Data for faults highlighted on composite section in D. (G) Section SO191-1. (H, I) Data for faults highlighted on section in G. The plots in B, C, E, F, H, and I show the reverse separation of reflections plotted against the updip distance along individual faults (separate colored curves), with measurements taken from reflector R5 as the origin, positive separations updip from R5 and negative separations downdip from it. Bold red curves in B, E, and H correspond to major frontal thrusts (F17 in A, F16 in G), and an incipient thrust (F17 in D) labeled in the upper panels. C, F, and I are enlargements of the protothrust separation data compressed in B, E, and H. Dashed curves in C, F, and I are individual segments and composite separation data for vertically segmented protothrusts (see examples in Figs. 8A–8D and 9). Figure S1 (see footnote 1) compares these data with separation diagrams normalized against the maximum separation for each structure, thus highlighting general similarities and differences in the shapes of the respective curves on different sections.
the inner (landward) part of the PTZ (Figs. 3 and 4). We find that the widths scale positively with the apparent maximum separation measured in the profiles, with most of the fault population having separation:width ratios of 0.1–0.001 (Fig. 15). A breakdown of the separation:width data by seismic line and protothrust dip direction are presented in Figure S2 (see footnote 1). The regression lines are slightly more gently dipping than the average global regressions (Fig. 15) (e.g., Scholz and Cowie, 1990; Kim and Sanderson, 2005). Note, however, that the global data incorporate along-strike and downdip fault measurements from crustal faults with a larger range of separation and/or displacement and fault dimensions compared to the protothrusts analyzed here. Despite bathymetric imaging of protothrusts with lengths of 4–24 km, the limited seismic data available do not allow comparable displacement-length relationships to be established.

**Relationship to the Décollement and Major Thrusts**

The south PTZ is divided into domains bounded by thrust faults (Fig. 3). From south to north, the discontinuous series of major thrust faults along the landward edge of the south PTZ includes thrusts F15 (Aorangi Ridge), F17, and F16 (Fig. 2B). In contrast to most protothrusts, these major thrusts penetrate the entire sequence to the décollement (with widths of ~6.5 km; cf. Fig. 6C). These faults ramp upward from the décollement at dips of <20°, but they steepen upsection, typically to 40°–50° in the upper 2 km (i.e., with the same dip as the predominant protothrusts) (Ghisetti et al., 2016) (Figs. 3 and 4). The décollement is located in the inferred pelagic section on horizon R7 (Barnes et al., 2010; Ghisetti et al., 2016; Plaza-Faverola et al., 2016). In profile SO191-6 (Fig. 4), thrust F17 is clearly the major frontal thrust with almost 800 m of total separation, and is associated with a substantial bathymetric ridge (Fig. 7C). As shown in Figure 14B, the reverse separation diminishes toward the upper tip, with an interval of reduced separation between horizons R5B and R5 (Fig. 14B) (Ghisetti et al., 2016). Within the accretionary wedge west of this fault, older northwest-dipping protothrusts are visible in profile SO191-6 in the uplifted and tilted thrust-bound panels (Fig. 4). East of thrust F17 beneath the southern part of the south PTZ, there is no evidence for slip localized on the regional décollement horizon (R7).

In adjoined profiles 05CM-38–SO191-4 (Fig. 3) F16 is the primary frontal thrust, and F17 is an incipient thrust within the inner part of the south PTZ. Both structures ramp from the décollement horizon (R7), offsetting the inferred pelagic sequence below the inner protothrusts. Thrust F17 has ac-
cumulated ~300 m of total separation (Fig. 14E), but is associated with only subdued seafloor geomorphology (Fig. 7C). The displacement profile of this fault also reveals a separation low at horizon R5B (Fig. 14E).

In figure 6A of Barnes et al. (2010) it was shown that fault F18 is a major thrust that carries the northeast part of the south PTZ onto the subducting Bennett Knoll seamount (Fig. 2B). This fault is associated with a significant bathymetric scarp (Fig. 7C), but it is largely developed in between seismic profiles 05CM-38-SO191-4 and 38-SO191-1 (Fig. 2B), and is thus not a significant structure on any of the profiles presented herein. This thrust does not extend southwest to profile SO191-6 (Fig. 4), and is barely recognizable as a major fault on profile 05CM-38-SO191-4 (Fig. 3), which crosses the southern part of the fault at a stepover in its surface trace (Fig. 7C). The north-south strike of thrust F18 differs from the more common southwest-northeast strike of the primary frontal thrusts elsewhere along the landward edge of the PTZ. This reflects the orientation of Bennett Knoll seamount (Fig. 2B). Based on the position of thrust F18, however, incipient slip may be currently localized on the décollement horizon east of the incipient thrust F17 in profile 05CM-38-SO191-4 (Figs. 3 and 14D).

**North PTZ**

**Structural Geometry, Displacement Data, and Scaling Relationships**

The north PTZ reaches 25 km in width (Figs. 2B and 7A), and is characterized by a densely spaced array of protothrusts that also are preferentially concentrated in the RSB-R0 turbidite sequence (Figs. 5 and 8E–8H). In marked contrast to the south PTZ, seismic profile SO191-1 reveals that the north PTZ consists predominantly of southeast-dipping protothrusts (Barnes et al., 2010; Ghisetti et al., 2016). Conjugate structures associated with an antithetic fault set dipping northwest are widespread (Figs. 5, 8G, and 8H). In Ghisetti et al. (2016) it was shown that a large proportion (>55%) of the protothrusts dip at angles of 40°–50°, although the scatter in dip measured for the entire population ranges from ~20° to 70°. Figure 10 (tan curve) shows that protothrusts offsetting horizon R4-R5 between 12 and 17 km east of the deformation front have dips lower than average. This region coincides approximately with the location of an incipient thrust fault associated with growing folds (Fig. 5; see following). The north PTZ also is well expressed in the seafloor bathymetry (Figs. 7A, 7B), indicating that the protothrusts here also strike subparallel to the regional thrust front (F16). Southwest-northeast–striking geomorphic lineaments associated with singular protothrusts or closely spaced clusters of protothrusts have lengths ranging from 4 to 28 km (Table 1). An east-west–striking lineament crosses the north PTZ, and is inferred to be a right-lateral strike-slip fault. (The details and discussion of the significance of this feature will be presented elsewhere.)

Seismic profile SO191-1 reveals a large range (50–4960 m) of protothrust spacing across the strike of the north PTZ (Fig. 11, red and pink columns). The fault spacing frequency measured on horizons R5 and R4, and corrected for the profile obliquity of 7° relative to the inferred shortening direction normal to the frontal thrust F16 (i.e., 127°), is polymodal with the predominant modes at ~100–200 m and 300–400 m spacing. As with the south PTZ, the fault spacing increases over a depth of ~1 km in the turbidite sequence, and generally into the outer reaches of the PTZ (Table 1).

Relatively few protothrusts in the north PTZ appear to penetrate down dip into the inferred pelagic sequence below reflection horizon R5B (Figs. 5 and 8E–8H). They reveal a reduction in separation toward their apparent lower tip (Fig. 14I), with separation profiles being relatively more symmetrical compared to the south PTZ. The maximum apparent separation is centered approximately at marker horizon R4-R5 (Figs. 14G, 14I). Composite profiles derived from vertically segmented protothrusts have shapes similar to singular structures.

The apparent maximum separation measured on individual protothrusts in section SO191-1, corrected for the obliquity of the profile relative to the inferred shortening direction normal to thrust F16, ranges from ~3 to 55 m, with a median value of 10 m (Table 1). The protothrust widths range from 75 to 2550 m, with a median value of ~290 m. The widths also scale positively with the maximum separation measured in the profiles, with most of the fault population having separation:width ratios similar to the south PTZ (Fig. 15). There is no significant difference between the separation:width ratios of southeast-dipping versus northwest-dipping structures within the north PTZ, although the widths of northwest-dipping protothrusts seldom exceed 1000 m (Fig. S2; see footnote 1).

**Relationship to the Décollement and Major Thrusts**

In profile SO191-1 the major frontal thrust (F16) along the landward side of the PTZ is northwest dipping, has almost 900 m of reverse separation, and is associated with a pronounced bathymetric ridge (Fig. 7A). This fault ramps at an apparent dip of <20° from the primary décollement located just above reflection R5B in this area, but steepens to ~50° in the upper turbidite section. Reverse separation along the fault diminishes toward its upper tip, and locally reduced separation occurs close to horizon R5 (Fig. 14H). The other major thrusts in the wedge further west also have northwest dips (i.e., seaward vergence; Fig. 5). Despite this, all protothrust arrays across the frontal wedge as much as 35 km landward of the deformation front have predominantly southeast dips, subparallel to the north PTZ.

As in the south PTZ, there is subtle evidence for an incipient thrust developing east of the deformation front beneath the north PTZ, albeit with substantially less separation than thrust F17 in profile 05CM-38-SO191-4. On profile SO191-1 there are actively growing, opposite-verging folds developing 9 and 14 km east of the primary frontal thrust (Figs. 5 and 16). These folds converge at depth and affect the sequence bounded by reflections 5A and 6, consistent with our proposed interpretation that a thrust fault (inferred from...
Figure 16. (A–C) Cumulative shortening measured along seismic profiles (fault heave, Fig. 6B) plotted against distance east of thrust F16 for the southern protothrust zone. (D–F) For the northern protothrust zone. A and D show the markers used for this analysis (R3 is in orange, R5 is in red, and marker R4-R5 is in green). B and C compare the cumulative shortening on reflection R3 and the location of post-R3 growing folds (black filled triangles) with horizons R4-R5 (i.e., the stratigraphic level of maximum protothrust shortening in the southern PTZ) and R5. E and F compare the cumulative shortening on reflection R3 and the location of post-R3 growing folds with marker horizon R4-R5 (i.e., the stratigraphic level of maximum protothrust shortening in the northern PTZ; Fig. 14I). Note that for clarity of details in E, the y axis is plotted at 2x vertical exaggeration relative to the y axis in B. Steeply dipping dashed gray lines identify the location of structures on the respective cumulative shortening curves that relate to the corresponding active (post-R3) folds. Bracketed numbers above the curves in B and E, separated by short black lines normal to the curves, represent the percentage of total shortening accounted for by the corresponding segment of the respective curves.
northwest-dipping reflections) has started developing beneath these folds in the pelagic cover sequence, ramping not from horizon R5B, but from horizon R7. This may indicate that décollement development is starting to localize on horizon R7 beneath the inner part of the north PTZ, the same as further south and west of thrust F13 (Fig. 5) landward of the north PTZ. If so, the décollement is now propagating at a deeper level than that beneath the outer two major thrust panels in the outer wedge landward of the north PTZ (i.e., horizon R5B between thrusts F13 and F16) (Fig. 5). A deeper anticline located 16 km east of the deformation front in profile SO191-1 is associated with the inferred strike-slip fault that crosses the north PTZ (Fig. 7A).

Measurement of Shortening across the Protothrust Zone

To evaluate the total amount and spatial distribution of seismically resolved shortening across the active PTZ east of the frontal thrust, we applied two methods. (1) We calculated cumulative fault heave for the marker horizon R4-R5 on profiles 05CM-38–SO191-4 (Fig. 3) and SO191-1 (Fig. 5), which approximates the stratigraphic position corresponding to maximum protothrust separation in each PTZ (Fig. 16). For the south PTZ we also compared the measurements on marker R4-R5 with measurements on horizon R5, given that the separation profiles of some larger individual protothrusts appear to have maximum values close to this horizon (Fig. 14F). Fault heave for each protothrust and incipient major thrust offsetting these horizons was calculated from the reverse separation and average apparent fault dip. In Figures 16B and 16E, the cumulative heave measured is plotted against distance east of the frontal thrust on each section. The results show that across the south PTZ, a total cumulative heave of ~1097 m is measurable on all faults that offset horizon R4-R5 (green curve in Fig. 16B; Table 2), with 43% of the total being landward and inclusive of the incipient thrust F17. Despite the separation profiles of some individual protothrusts revealing an apparent maximum separation close to horizon R5 (Figs. 14C, 14F, and S1 [see footnote 1]), total deformation on this horizon does not represent the maximum cumulative shortening across the south PTZ. Horizon R5 (red curve in Fig. 16B) reveals 965 m of cumulative heave, with ~60% landward and inclusive of thrust F17. In comparison, across the north PTZ, a total cumulative heave of ~525 m is measured on all faults that offset the marker R4-R5 (Fig. 16E; Table 2), which is ~50% of the maximum in the south PTZ.

(2) We derived shortening of ~1095 m and ~600 m for marker horizon R4-R5 in the south and north PTZs, respectively, by restoring the horizon to a putative horizontal template (restoration performed using section analysis in Move software) with shortening defined as the original line length, L

TABLE 2. SUMMARY OF TECTONIC SHORTENING DATA FOR THE NORTH AND SOUTH PROTOTHRUST ZONES, DERIVED FROM SEISMIC MEASUREMENTS OF CUMULATIVE HEAVE, SECTION ANALYSIS, AND ONE-DIMENSIONAL CUMULATIVE POWER LAW REGRESSION DATA SPECIFICALLY FROM MARKER HORIZON H R4-R5 CORRESPONDING TO THE MAXIMUM SHORTENING IN EACH PROTOTHRUST ZONE

<table>
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<tr>
<th>Component</th>
<th>Power law heave truncation (m)</th>
<th>X value (fauls with heave ≥ X value)</th>
<th>Y value (number of faults in each fault size class)</th>
<th>Total heave contributions from subseismic faults on H R4-R5 (m)</th>
<th>Total heave contributions above power law heave truncation (m)</th>
<th>Cumulative heave actually measured from all faults on H R4-R5 (Fig. 16) (m)</th>
<th>Shortening estimated from linear restoration (m)</th>
<th>Total shortening on H R4-R5 in the plane of each profile (m)</th>
<th>Total shortening on H R4-R5 corrected for profile obliquity* (m)</th>
<th>Component of shortening imaged seismically (%)</th>
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Note: PTZ is protothrust zone. Section analysis used Move software (Midland Valley software; https://www.mve.com/software). Bold font in values within columns highlights the estimates of tectonic shortening.

*Subtotal (subseismic shortening from power law cumulative regression).
+Subtotal (seismic-scale cumulative heave calculated from measurements of individual proto-thrust apparent separation on horizon and average dip, for faults with heave > power law truncation.
+Total shortening on horizon from above two subtotals.
**Relative to the average margin normal bearing associated with strike of the frontal thrusts in this region (127°).
placement ≥d can then be written \( N \sim d^{-c} \), where on fault displacement show that the cumulative number of faults (\( N \)) with displacement (e.g., fault displacement, fault length) commonly display power law scalings, consistent with fractal characteristics of the fault population (e.g., Scholz and Cowie, 1990; Walsh et al., 1991; Walsh and Watterson, 1992; Marrett and Allmendinger, 1991, 1992; Yielding et al., 1996; Bonnet et al., 2001). Data sets on fault displacement show that the cumulative number of faults (\( N \)) with displacement ≥d can then be written \( N \sim d^{-c} \), where \( c \) characterizes the fractal dimension of the population, with increasing values indicating a larger proportion of smaller fault displacements relative to larger fault displacements. Power law distributions have straight-line segments on plots of \( \log N \) versus \( \log d \), with values of \( c \) typically of ~0.5–1.0 for one-dimensionally sampled populations (e.g., one or multiline measurements along boreholes or horizons in a section), and 1.3–1.7 for two-dimensionally sampled populations (e.g., cross sections, maps, satellite data).

The fractal size implies that that the number of faults with relatively large displacement (or length) in a power law population (e.g., faults mapped in the field or detectable in seismic profiles) can be used to estimate the number of faults otherwise undetectable at the scale of a given analysis (e.g., microscopic faults, or faults below the limits of seismic resolution).

Power law regression lines on distributions of fault parameters are derived only from the component of the population that satisfies the power law relationship. A variety of factors may influence power law regressions, including the number of samples measured and the bandwidth of the measurements required to give a statistically reliable characterization of the population (Bonnet et al., 2001). Ideally, a range of values over 2–3 orders of magnitude should be sampled from at least 200 measurements to give the best definition of a power law exponent; from a global review, however, Bonnet et al. (2001) demonstrated that in practice, 2–3 orders of magnitude are rarely met. Three common issues contribute to limiting the range of scale. (1) The specific resolution of a given analysis determines the size scale of under-sampled faults (truncation). This invariably leads to a shallowing of the slope of the power law distribution at the lower end of the scale range. (2) Faults with sizes comparable to the sampled region may be incompletely sampled (censored) because they extend outside the boundary of observation. This may lead to a steepening of the slope of the distribution at the upper end of the scale range. (3) There may be finite size effects within the population that are related to fault processes and lithological properties. Despite these issues, measurements over a limited range of approximately 1 order of magnitude, from <200 measurements, have been shown to produce reliable power law exponents. Nicol et al. (1996) showed that scale may also influence power law regressions, with \( c \) values derived from outcrop data commonly in the range of ~0.4–0.6, compared to values of 0.8–1.0 more typical for seismic data. Limiting the extrapolation to no more than 1 order of magnitude may avoid potential differences between submeter-scale fault populations and distributions derived from seismic data.

For each of our seismic profiles we plot the cumulative distribution for the apparent maximum fault separation for the frontal thrust faults and all protothrusts east of the frontal thrust (Fig. 17A). The apparent maximum separation on individual faults was measured from a range of stratigraphic position. The log versus log curves for these two-dimensionally sampled data are nonlinear, with reduced slopes toward low separations, resulting from the under-sampling of small displacement faults at the scale of the seismic lines. The slopes of the curves also shallower toward the largest separations associated with the major frontal thrusts. Figure 17B shows that the apparent maximum separations measured on all structures within in the PTZ display a power law relationship over >1 order of magnitude, with sample size of ~100–200 faults per section, and truncation at measured separation of 13–15 m. The distributions are characterized by high power law exponents, \( c \), of 1.5–2.2.

One-dimensional fault populations derived from displacements along a marker horizon in a section may be used to estimate the contribution of faults unresolved at the seismic scale (subseismic faults) to total shortening or extension (e.g., Walsh et al., 1991; Marrett and Allmendinger, 1991). To estimate the contribution of subseismic faulting to total shortening across the Hikurangi PTZ east of the primary frontal thrust, we plot the cumulative distribution for apparent fault heave from all structures displacing the marker horizon R4-R5 (Fig. 17C). This horizon approximates the stratigraphic position of the maximum measured shortening in both PTZs (Fig. 16). Data plotted in Figure 17C are consistent with a power law relationship, although we can only account for <100 measurements per section and scale ranging across ~1 order of magnitude. The cumulative power law regressions have high exponents of ~1.7 (south PTZ) and ~2.0 (north PTZ), respectively. Although the regression lines are subparallel, the regression for the south PTZ is offset to the right of that for the north PTZ, reflecting a different protothrust displacement population with a larger number of faults for a given heave value. This is consistent with the larger estimates of tectonic shortening measured in the south PTZ at seismic scale (Fig. 16). The population truncation for fault heave, defined at the inflection point where the slopes of the curves becomes shallower, are 12 m (south PTZ) and 7 m (north PTZ), respectively (Fig. 17C).

The extent to which the linear trend of the power law distribution can be reliably extrapolated to the smallest heave values remains unknown. Extrapolation by about one order of magnitude (i.e., to faults with heave ~1 m) may well represent the upper limit of reliability for estimates of heave from small-
Cumulative distributions plotted against protothrust and frontal thrust apparent maximum separation and heave (Fig. 6C). PTZ—protothrust zone. (A) Apparent maximum separation curves (color coded for the respective seismic sections) include protothrusts from all stratigraphic intervals and the principal frontal thrust in each section (i.e., these are two-dimensional data). Thin gray lines indicate the protothrust range of measurements from seismic data. (B) Measurements as in A, but excluding the frontal thrusts. Power law regressions (with their respective equations and R squared values) are derived for the straight sections of the curves to the right of the protothrust population separation thresholds. Dashed lines extrapolate the regressions below the resolution of the seismic sections, indicating that large numbers of closely spaced, small-displacement protothrusts exist at subseismic scales. (C) Cumulative distribution of protothrust apparent maximum heave measured on marker horizon R4-R5 in both the north and south PTZs. All data are from east of the frontal thrust F16. Power law regressions (with their respective equations and R squared values) are derived from only the straight sections of the curves to the right of the protothrust heave truncations. Dashed lines extrapolate the regressions to ~1 order of magnitude in heave below seismic resolution. See text for explanation, and Table 2 for calculation of the maximum possible shortening from subseismic faulting below the heave truncations.

Figure 17. Cumulative distributions plotted against protothrust and frontal thrust apparent maximum separation and heave. (A) Apparent maximum separation curves (color coded for the respective seismic sections) include protothrusts from all stratigraphic intervals and the principal frontal thrust in each section (i.e., these are two-dimensional data). Thin gray lines indicate the protothrust range of measurements from seismic data. (B) Measurements as in A, but excluding the frontal thrusts. Power law regressions (with their respective equations and R squared values) are derived for the straight sections of the curves to the right of the protothrust population separation thresholds. Dashed lines extrapolate the regressions below the resolution of the seismic sections, indicating that large numbers of closely spaced, small-displacement protothrusts exist at subseismic scales. (C) Cumulative distribution of protothrust apparent maximum heave measured on marker horizon R4-R5 in both the north and south PTZs. All data are from east of the frontal thrust F16. Power law regressions (with their respective equations and R squared values) are derived from only the straight sections of the curves to the right of the protothrust heave truncations. Dashed lines extrapolate the regressions to ~1 order of magnitude in heave below seismic resolution. See text for explanation, and Table 2 for calculation of the maximum possible shortening from subseismic faulting below the heave truncations.

In comparison with the south PTZ, Table 2 shows that in the north PTZ, across profile 05CM-38–SO191-4, as many as 2833 (i.e., 2871 – 38) subseismic-scale protothrusts may occur on marker horizon R4-R5 with heave of between 1 and 12 m. Almost 2000 of these faults would have heave of 1–2 m, indicating an average spacing of ~13 m across the 25 km width of the PTZ for protothrusts of this displacement size class. Protothrusts with heave of between 1 and 12 m could therefore potentially contribute as much as 6570 m of apparent shortening. When added to the apparent shortening value of 971 m derived from seismically imaged faults with heave ≥12 m on marker R4-R5 (i.e., above the power law heave truncation; Fig. 17C), the maximum total apparent tectonic shortening in the plane of the profile can be estimated to be ~7640 m (~23%, using the restored PTZ width as a reference length). Correcting for the profile obliquity of 17° relative to the inferred margin normal shortening direction results in tectonic shortening of as much as ~7210 m across the south PTZ (Table 2). If the population is fractal over 1–100 m of heave (Fig. 17C), the implication is that as little as 13% of the total shortening east of the principal deformation front (fault F16) can be measured from faults imaged in the seismic lines. These results are obviously sensitive to the validity of the regression lines in Figure 17C and to the acceptable extent of the extrapolation over one order of heave magnitude, but unfortunately cannot be tested with additional data at smaller scales. For example, extrapolation of the regression lines only to heave displacements ≥5 m would have the impact of removing >5 km of shortening from the estimates in Table 2.

In comparison with the south PTZ, Table 2 shows that in the north PTZ, across profile SO191-1, as many as 1723 subseismic faults (i.e., 1758 – 35) may occur on marker horizon R4-R5 with displacement heave of between 1 and 7 m. If so, these faults would contribute ~3500 m of shortening. When added to the apparent shortening value of 446 m derived from seismically imaged faults on marker R4-R5 with heave greater and equal to 7 m (i.e., above the
DISCUSSION

To our knowledge, the fault population data presented here for the Hikurangi PTZ are the first of their type for subduction protothrust systems. Despite differences in predominant dip along the strike of the margin, both the north and south PTZs share the following general attributes. They appear to be largely confined to the Pliocene–Pleistocene turbidite sequence, with relatively few structures appearing to penetrate downward into the older pelagic cover sequence. The PTZs strike subparallel to the primary frontal thrusts, as supported by geomorphic lineaments associated with protothrusts or clusters of protothrusts, identified at the scale of the multibeam bathymetry data (30 m grid DEM). They consist of closely spaced (~100–400 m) arrays of seismically resolvable protothrusts (sometimes in conjugate sets), commonly with moderate dip (~40°–50°) and small reverse separation (<20–30 m). No evidence is seen for extensional offsets. Measurements of protothrust separation and width scale positively, while power law relationships established for the protothrust separation and heave imply self-similarity of the displacement and fault dimension characteristics over more than one order of magnitude.

Our observations and results allow us to explore in the following sections the growth and mechanics of the PTZ, and to speculate on the influence of possible rheological heterogeneities associated with stratigraphy. We propose an evolutionary model of frontal deformation, and evaluate the role of protothrusts in accommodating plate boundary convergence. We make comparisons with other margins worldwide characterized by PTZs, for which our results and methods may be applied.

Growth of Protothrusts and Their Relationship to Stratigraphy and Propagation of the Décollement

The observation that the PTZ appears to be largely confined to the Pliocene–Pleistocene turbidite sequence above horizon R5B (notwithstanding the reduction in vertical seismic resolution with depth) led Ghisetti et al. (2016) to speculate that rheology-related differences in deformation mechanisms may be occurring with depth. Theoretically, spatial differences in the rheological properties of the host rocks are controlled by lithology, consolidation state, cementation, and/or fluid pressure regime.

The physical characteristics of the sequence hosting the PTZ remain unknown, because the sequence has never been drilled. However, regional seismic data support the interpretation that the relatively highly reflective sequence hosting the PTZ comprises basin-floor turbidites that were deposited from turbidity currents spilling over the levees of the Hikurangi Channel (Lewis et al., 1998; Lewis and Pantin, 2002; Barnes et al., 2010; Plaza-Faverola et al., 2012; Bland et al., 2015). Holocene deposits recovered in short (<6 m) gravity cores typically comprise 50–400-mm-thick graded sand-silt turbidites with low clay contents (Lewis and Pantin, 2002; our data). We infer that any potential similar sequence at greater depth will be undergoing vertical porosity reduction and possibly cementation under sediment loading (e.g., Scretton et al., 2002). This would be consistent with the increasing seismic velocity with depth (Plaza-Faverola et al., 2016; Ghisetti et al., 2016). Seismically resolved porosity reduction and stress can both be expected to increase toward the deformation front and in the vicinity of major thrust faults (e.g., Moore and Bryne, 1987; Morgan and Karig, 1995; Saffer, 2003; Moore et al., 2011). Moreover, the precise profile and lateral distribution of shear strength will likely be influenced by pore pressure conditions and their evolution through time (e.g., Saffer, 2003), with a strong control by heterogeneities in the stratigraphic sediment facies, permeability, and drainage (Saffer and Bekins, 2002). Sedimentation rates at Hikurangi Trough since ca. 2.0 Ma are high, reaching ~0.75–1.0 km/m.y. close to the deformation front (Figs. 3–5). These rates raise the likelihood of pore pressures locally elevated above hydrostatic in sealed layers, although we have no data to confirm this. We speculate that different lithologies likely undergo different deformation mechanisms, with softer sediments likely accommodating significant horizontal shortening and vertical thickening through ductile means (including compaction). Future analyses of seismic velocity based porosity and stress at Hikurangi Trough may enlighten us on this topic further, while future ocean drilling could provide lithological constraints on possible facies heterogeneity and mesoscale deformation mechanisms.

The large range of measured protothrust dips in the Hikurangi PTZ (Table 1) (Ghisetti et al., 2016) may result from different mechanisms of fracturing. The predominance of moderately dipping (40°–50°) protothrusts is consistent with interpretations elsewhere that such features develop as compactive shear bands associated with porosity reduction, fabric collapse, and dewatering (e.g., Ujiie et al., 2004; Morgan et al., 2007). We plotted the dip of protothrusts measured on structures offsetting horizon R4-R5 in the north and south PTZs against distance east of the frontal thrusts (Fig. 10) to evaluate whether any systematic changes in dip can be observed and related to different mechanisms of protothrust nucleation in areas that may be undergoing higher strain (e.g., Morgan and Karig, 1995; Morgan et al., 2007). Two features are apparent from this analysis. There is more scatter associated with lower dipping protothrusts in the higher strain south PTZ compared to the north PTZ (cf. Table 2). The occurrence of lower dipping protothrusts in the PTZ appears...
to be within 3–6 km of major frontal thrust faults and/or the regions where incipient (post-R3) thrusts and folds are developing within the PTZ. Although circumstantial, these observations support the possibility that the scatter in dips results from newly formed brittle structures overprinting earlier formed compaction bands in areas of relative higher strain (Morgan et al., 2007).

Regardless of the predominant PTZ dip, the protothrust spacing increases in the outer reaches of the PTZ (Fig. 11) and over ~1 km depth in the Pliocene–Pleistocene turbidite sequence (Figs. 3–5). The protothrusts appear to nucleate in the turbidite interval R4 to R5 (Fig. 8), implying that the stratigraphy and consolidation state may be factors controlling their mechanical formation. The protothrust separation data in Figure 14 are consistent with fault growth and propagation of the protothrust tips both updip and downdip away from the nucleation region. Some of the separation variability on protothrusts observed in Figures 14C, 14F; and 14I could result from lithological heterogeneity within the turbidite sequence (e.g., Muraoka and Kamata, 1983; Schöpfer et al., 2008). However, we suspect that this variability is mainly caused by the growth of larger protothrusts by merging and coalescence of smaller protothrusts (e.g., Watterson, 1986; Walsh and Watterson, 1988; Bull et al., 2006). This would be consistent with the composite separation profiles derived from vertically segmented protothrusts indicating that the separation is conserved (e.g., Fig. 14C).

Regardless of the precise growth mechanism, the separation-width scaling relationships for the PTZ (Figs. 15 and S2 [see footnote 1] are consistent with a fault propagation where displacement increases with fault size. Our data fit similar scaling relationships derived from a variety of global faults that include a range of separation and/or displacement and fault scale larger than our data set (e.g., Watterson, 1986; Walsh and Watterson, 1988; Kim and Sanderson, 2005; Bull et al., 2006). The sampling dimensions of our fault population are limited by the scale of faults resolvable in the seismic lines. We can estimate that, at the resolution of the available multibeam bathymetric data (30 m DEM), the seafloor geomorphology associated with singular protothrusts or closely spaced clusters of protothrusts becomes visible when structures reach ~4 km length (Fig. 7).

Several factors indicate that not all protothrusts east of the frontal thrusts are coeval, but that the PTZ has propagated eastward. Within the turbidite sequence hosting the PTZ, differences in strain localization with depth and through time are evident in the curves of cumulative heave measured across the PTZ (Fig. 16). Notably, in the south PTZ (Figs. 16B, 16C), comparison of the curves for R4-R5 and R3 horizons, combined with the location of post-R3 fault-related folds (black triangles in Figs. 16B, 16C x-axes), reveals that deformation is currently concentrated on the developing thrust F17 and on protothrusts and associated folds east of this structure. We suggest that the dense array of inner protothrusts in the hanging wall of F17 likely developed at an earlier stage and now appears to be deactivating. Increasing activity in the outer zone (as measured by the proportion of total shortening indicated in brackets on each cumulative heave curve in Fig. 16) is yet to be reflected by the closer spacing of seismically imaged protothrusts evident inboard (Figs. 3, 10, and 16A).

These observations imply that eastward propagation of the PTZ has accompanied forward propagation of the décollement and the incipient development of new thrust faults and folds east of the primary frontal thrusts. In the north PTZ, post-R3 folding of the turbidite section 9 and 14 km east of the frontal thrust is associated with a possible incipient thrust developing beneath them in the upper pelagic sequence above horizon R7 (Figs. 5 and 16D). This inferred thrust is at a very incipient stage of development, suggesting that displacement is only starting to localize on the décollement beneath the inner PTZ. Beneath the inner 10 km of the south PTZ, the décollement is more established and thrust F17 is at a more advanced stage of development, having accrued ~300 m of separation (Figs. 3 and 16A).

Thrust F17 in the south PTZ ramps from the décollement at horizon R7 and has broken through the entire accreting (pelagic and clastic) sequence, including the upper pelagic section not typically involved in seismically resolvable protothrust deformation. Similar to the major frontal thrusts, this fault has a listric geometry resulting from low-angle dip (~25°) in the pelagic sequence beneath horizon R5B, and steeper dip upward, where the fault links with a preexisting protothrust (or cluster of protothrusts). Overprinting of a preexisting protothrust is indicated in Figure 14E by the inheritance of a protothrust displacement profile, with preservation of a displacement low stratigraphically localized toward the base of the clastic sequence at horizon R5B (e.g., Ellis and Dunlap, 1988; Ghisetti et al., 2016). Figures 14B and 14H demonstrate that similar displacement lows are retained and enhanced on the more advanced major thrusts at the deformation front even for finite displacements approaching 1 km (i.e., one to two orders of magnitude larger than typical protothrusts). This may indicate some form of fault weakening (strain softening) in the development of protothrusts, whereby their original high-displacement centers continue to accrue relatively more shortening than near their tips even after capture by a major thrust.

In Ghisetti et al. (2016) it was demonstrated that although nonuniquely, trishear fault-propagation mechanisms suitably account for the developing thrust and fold geometry as increasing displacement accrues on the thrust, and through its evolution from incipient and blind to mature and emergent. As slip becomes increasingly localized on the propagating décollement and on the thrust fault that has broken through the PTZ, the accumulated displacement drives the development of a broad fold at the seafloor above the fault tip (e.g., F17 in Fig. 3). The earliest onset of bathymetric relief develops a depositional growth sequence in the overlying turbidites across the hanging-wall anticline and footwall syncline. This reflects the fact that sedimentation and thickness of channel-overbank turbidites depositing in the Hikurangi Trough are highly sensitive to topographic relief developing on the basin floor (Lewis and Pettings, 1993; Barnes and Mercier de Lépinay, 1997; Plaza-Faverola et al., 2012).

**Generic Model for Frontal Accretion and PTZ Migration**

The observations here allow us to build on a generic model of frontal accretion developed in Ghisetti et al. (2016) from progressive retrodeformation at time intervals of ~1 m.y. and less. We propose the following generalized stages of deformation front and PTZ evolution, as conceptualized in Figure 18.
1. In an early stage of development (Fig. 18A) an array of predominantly moderately dipping protothrusts develop, possibly as compaction bands associated with subhorizontal compression and lateral porosity reduction immediately east of and subparallel to the primary deformation front. Regardless of the predominant PTZ dip, they appear to nucleate in the turbidite interval R4 to R5 where factors such as lithology, consolidation state, and cementation could potentially control their mechanical formation. With ongoing contraction, the developing PTZ densifies with self-similar growth in protothrust size and displacement at seismic and subseismic scales. The width of this zone is narrower than the maximum present-day width of the entire PTZ, but may exceed 15 km. The major frontal thrust together with higher displacement protothrusts (or clusters of protothrusts) grow with increasing relative displacement and control seafloor geomorphology. Few, if any, protothrusts penetrate the underlying inferred upper pelagic sequence, which could be shortening by other diffusive mechanisms such as ductile shear and mineral rotations (Morgan et al., 2007). At this stage, however, there may be relatively poor localization of deformation on the décollement horizon beneath the PTZ.

2. With increasing strain, porosity reduction, and protothrust growth, displacement beneath the inner PTZ localizes onto the propagating décollement on horizon R7 and a new thrust fault breaks through the entire accreting sequence within the PTZ, including the upper pelagic units and the PTZ-hosting turbidites (Fig. 18B). A listric thrust geometry results from the deeper low-angle thrust ramp linking with and inheriting steeper preexisting protothrusts. Strain localization onto the décollement and new thrust fault are accompanied by a reduction in the displacement rate of the predominantly moderately dipping protothrust array across the hanging wall of the thrust (e.g., Fig. 18B). Localized higher strains associated with thrust development, however, could potentially be accompanied by contemporaneous onset of lower angle structures in the PTZ (e.g., Fig. 10), which may overprint the relatively steeper dipping, earlier formed protothrusts. The PTZ seaward of the developing thrust fault continues to develop under horizontal compaction, with relatively increased displacement rates and eastward advance of the PTZ front. Seafloor geomorphology continues to develop over the major frontal thrust, the newly developing PTZ thrust, and the higher displacement protothrusts or clusters of protothrusts.

3. Further localization and accumulation of strain on the propagating décollement and the new thrust fault within the PTZ are accompanied by the progressive development of a fault-propagation anticline associated with more prominent seafloor deformation and a fault-growth sedimentary sequence across the fault tip (e.g., Fig. 18C). This more advanced deformation marks the stage at which the inner part of the PTZ is transferred to a major thrust hanging wall, and is accompanied by eastward propagation of the décollement and PTZ seaward of the developing thrust.

4. The emergence and growth of the new thrust fault through the PTZ, accompanied by an increasing displacement rate relative to the previous frontal thrust, ultimately leads to establishment of a new frontal thrust associated with a prominent bathymetric ridge. At this stage, the early-formed inner PTZ has been uplifted and tilted within the hanging wall of the developing thrust. Analogous and now-accreted PTZs are interpreted in the seismic data as much as 35 km landward of the deformation front, within panels bound by thrusts F11–F16 (Figs. 2 and 5). While these PTZs are assumed to be deactivated relative to their formative activity once accreted behind a newly established frontal thrust ridge, some activity in these zones may continue, including selective reactivation of some structures along with other major thrusts as the wedge continues to be shortened (Ghisetti et al., 2016).

We propose that the incremental propagation of the décollement coincides with out-stepping of the frontal thrust, a migrating wave of PTZ development, and the progressive deactivation and accretion of PTZs into the wedge (Fig. 18). This proposed sequence of deformation is in good agreement with the generic model of accretionary wedge growth, proposed in Ghisetti et al. (2016), from retrodeformation of the wedge; those reconstructions invoke progressive eastward propagation of the basal décollement over ~2 m.y., associated with development of new thrust faults that evolve from blind to emergent, migra-
tion of protothrust deformation east of the frontal thrust, and repeated out-of-sequence fault reactivation within the wedge to maintain its critical taper. The demonstrated repetition and outward migration of this sequence of frontal deformation events, facilitated by a thick clastic input sequence, resulted in rapid forward advancement and widening of the accretionary wedge by as much as 60 km over the past 2 ± 0.8 m.y. (Ghisetti et al., 2016). The proposed sequence in Figure 18 is generally consistent with the progressive strain localization, structural inheritance, and frontal thrust development deduced from analogue experiments of accretionary wedge growth (e.g., Adam et al., 2005; Dotare et al., 2016) and from numerical simulations (Ellis et al., 2004).

Role of Protothrusts in Accommodating Plate Boundary Shortening

The power law regressions for apparent maximum protothrust separation from the Hikurangi PTZ seismic profiles indicate that the PTZ likely comprises a dense network of hundreds to thousands of small displacement faults and fractures (Figs. 17 and 18) that remain undetectable at the resolution of the seismic data used for the analysis (subseismic-scale faults).

Our results indicate that as much as 4–7 km of tectonic shortening could potentially be accommodated within the PTZ (Table 2), but we note that these estimates are highly sensitive to the validity of the power law regression lines in Figure 17C and an acceptable level of extrapolation to smaller fault displacement sizes. Inadequate sample size and insufficient amplitude of the scale range sampled by the PTZ one-dimensional population may be factors contributing to the steepness of the regression lines. The resulting estimates of total shortening could therefore be substantially reduced (e.g., by several kilometers) from the maximum values listed in Table 2. However, if our regressions and extrapolation are valid, it is possible that as much as ~90% of the total shortening across the PTZ occurs on structures below the resolution of the seismic profiles. Moreover, these estimates account only for faults with heave ≥1 m, and do not account for the pervasive submeter and microscale deformation that can be expected in porous sediments undergoing compaction. Despite the large uncertainties in the estimates of total tectonic shortening, we propose that PTZs play a significant role in accommodating plate boundary convergence over the long term. Our findings appear to be consistent with seismoc velocity– and porosity-based estimates of tectonic shortening (Morgan et al., 1994; Morgan and Kariş, 1995), as well as observations of widespread mesoscopic deformation bands, and microscopic zones of rotated clay minerals in ODP cores from the Nankai Trough PTZ (e.g., Moore et al., 1986; Lundberg and Moore, 1986; Maltman et al., 1993; Ujiie et al., 2004; Morgan et al., 2007).

The maximum inferred tectonic shortening rate accommodated by the PTZ east of the primary frontal thrust can be estimated by dividing the upper range of total shortening by the time period assumed to represent the duration of the current PTZ formation. Notwithstanding the absence of precisely dated markers in the Hikurangi Trough and the currently poorly constrained age estimates of Plaza-Faverola et al. (2012), Ghisetti et al. (2016) inferred late Pleistocene development of the primary frontal thrusts and frontal PTZ from section restorations incorporating growth faulting, and estimated, for example, that thrust F16 and the north PTZ appeared to have formed since 0.6 ± 0.2 Ma (post–reflector R3), and possibly much later (i.e., within 0.1 ± 0.05 m.y.; Fig. 10 therein). Late Pleistocene development is consistent with subiced fault growth stratigraphy associated with minor ant clinonal and monoclinal folding across the PTZ being essentially limited to the uppermost sedimentary sequence above reflector R3. The results of this study, however, suggest that thrust F16 and the north PTZ together could possibly accommodate as much as 4.6 km of total shortening. If so, this shortening must have taken place over a period significantly longer than 0.1 m.y., otherwise the average rate of shortening on these structures alone would approach or exceed the total margin-normal plate convergence rate of ~40–50 mm/yr at this latitude (Fig. 2) (Wallace et al., 2004).

Taking a conservative estimate of 0.6–0.3 Ma for the age of the Hikurangi PTZ, the maximum estimate of ~4 km of shortening inferred east of the frontal ridge across the north PTZ (Table 2) could have occurred at an average rate of ~7–13 mm/yr. This would represent as much as ~14%–32% of the total margin-normal convergence rate of ~40–50 mm/yr at this latitude (Fig. 2B). In comparison, the inferred maximum of ~7.4 km of shortening possible east of the frontal ridge across the south PTZ could have occurred at an average rate of ~12–25 mm/yr, thus representing ~30%–62% of the total margin-normal convergence rate of ~40 mm/yr at this latitude.

Despite the uncertainties in shortening magnitude and rate, these findings raise the question of what contribution all PTZs across the frontal wedge have made in accommodating plate convergence. PTZs with predominant southeast dip are clearly imaged in the seismic data between thrusts F11 and F16, as much as 35 km landward (west) of the north PTZ (Figs. 2B and 5). Considering the width of the north PTZ, our estimate of shortening of as much as ~4 km equates to an average shortening rate across strike of 160 m/km. If this rate can be applied across all PTZs on profile SO191-1, the northern PTZs collectively could have accommodated shortening of ~5.6 km in addition to the major thrust faults. It follows that the PTZs spanning a total width of 60 km could possibly account for tectonic shortening of ~9.5 km. In comparison, the shortening rate per kilometer across the south PTZ could be as high as ~300 m/km, even though accreted PTZs are not resolved in all of the seismic data there, landward of the frontal thrust (cf. Figs. 3 and 4).

From restoration of the major structures, Ghisetti et al. (2016) considered the outer 35 km of the accretory wedge landward of the northern PTZ to have formed since 2.0 ± 0.8 Ma, and possibly since 1.0 ± 0.5 Ma (Figs. 10 and 13 therein). Dividing this interval by the ~9.5 km of possible PTZ shortening estimated across the wedge represents an average protothrust shortening rate of ~3–8 mm/yr over 2 ± 0.8 m.y., or as much as ~6–19 mm/yr if the deformation is not older than 1.0 ± 0.5 Ma in the northern region. This range of estimates implies that Hikurangi protothrust deformation across the northern part of the study area could possibly account for ~10%–50% of the total margin-normal plate convergence rate of ~40–50 mm/yr over ~1–2 m.y. time periods. These values are consistent with
previous estimates of shortening determined only from restorations of the major thrust faults and folds. In Ghisetti et al. (2016) it was shown that major faults across the frontal wedge account for only ~20%–45% of total margin-normal convergence rate over 2.0 ± 0.8 m.y. When combined with the shortening rate derived from PTZs, these estimates account for ~30%–95% of total margin-normal convergence accommodated by faulting at different scales.

**Comparison of Hikurangi PTZ Deformation with Other Margins**

The forward advancement and widening of the Hikurangi accretionary wedge by as much as 60 km over the past 2 ± 0.8 m.y. represents a rapid rate (~20–50 km/m.y) of frontal accretion (Ghisetti et al., 2016). This rate is comparable to the ~40 km of wedge growth documented at Nankai Trough since 2 Ma (Moore et al., 2001), and the ~20 km growth of the eastern Aleutian wedge since 3 Ma (von Huene et al., 1998). It is substantially slower than the extraordinary rate of >80 km/m.y. inferred at Cascadia since 0.3 Ma (Adam et al., 2004). PTZs appear to be a characteristic of these types of rapidly growing accretionary margins associated with thick turbidite sections that have been deposited at relatively high rates compared to margins with thin sediment cover.

The cumulative power law regression exponents of 1.7–2.0 for one-dimensionally sampled Hikurangi protothrust heave measurements (Fig. 17C) are high relative to values derived from other brittle fault systems (e.g., Scholz and Cowie, 1990; Walsh et al., 1991; Walsh and Watterson, 1992; Marrett and Allmendinger, 1991, 1992; Yielding et al., 1996). The implication that a large array of small-scale faults within the active PTZ accounts for 14%–23% shortening (using restored PTZ length as a reference frame), and that all PTZs across the wedge could accommodate ~10%–50% of the total margin-normal convergence rate compares favorably with other estimates of PTZ and diffuse strains at the front of accretionary wedges calculated from landward thickening of stratigraphy and reduced porosity (e.g., Bray and Karig, 1986; Moore et al., 2011). For example, Adam et al. (2004) estimated 9%–11% shortening across a 5-km-wide PTZ at Cascadia. In Morgan and Karig (1995) it was estimated that 68% of the total 3.3 km of shortening across the PTZ and the first two frontal thrusts along the Muroto transect at Nankai Trough are attributable to diffuse strain associated with porosity reduction. For the same region of the Nankai Trough, Moore et al. (2011) estimated 22% shortening across the PTZ, attributing just 4% to observable fault displacements. These results compare with our estimates that as little as 11%–13% of the total shortening across the active Hikurangi PTZs occurs on the larger scale faults imaged on seismic profiles.

**CONCLUSIONS**

The 25-km-wide Hikurangi PTZ east of the frontal thrust provides an excellent record of the structural growth of PTZs and the various stages of incipient deformation at a turbidite-rich accretionary margin. The PTZ is split into two along strike by the largely buried Bennett Knoll seamount. The two PTZs have opposite predominant dip direction. Both the north and south PTZs are characterized by predominantly closely spaced (commonly 100–400 m) arrays of seismically resolvable faults, commonly with moderate dip (40°–50°), small reverse separations (<20–30 m), and conjugate structures. They both strike subparallel to the primary deformation front, and are largely confined to the Pliocene–Pleistocene turbidite sequence, with relatively few protothrusts apparently penetrating the inferred pelagic cover sequence above the décollement horizon.

Measurements of fault separation in seismic sections reveal a reduction of displacement toward the upper and lower tips of protothrusts, consistent with their nucleation and progressive growth in the turbidite sequence. Their separation profiles are also consistent with updip and downdip migration of the protothrust tips. The observed variability along the displacement profiles in seismic sections may be related to growth of larger protothrusts by merging and linkage of smaller protothrusts, and lithological heterogeneity within the clastic sequence. The larger protothrusts are visible as 4–28-km-long seafloor lineaments in 30 kHz multibeam bathymetry data reduced to 30 m grid resolution.

The formation of protothrusts in the turbidite sequence and their scarcity in the underlying pelagic sequence above the décollement horizon imply that deformation mechanisms change with depth. The predominance of moderate dipping (40°–50°) protothrusts is consistent with their possible formation as compactive shear bands associated with porosity reduction, fabric collapse, and dewatering. Areas of higher strain in the vicinity of the major frontal thrust faults and incipient new thrusts propagating through the PTZ could be characterized by lower angle (<40°) brittle structures that overprint the early compactive deformation features. The upper pelagic sequence underlying the turbidites and above the décollement horizon may be deforming by other diffuse mechanisms.

The apparent maximum protothrust separations measured in the seismic profiles scale positively with width, with separation:width ratios of 0.1–0.001. These data are comparable to other fault populations worldwide sampled at a wide range of dimension and displacement scales. The Hikurangi PTZ displays a power law relationship over >1 order of magnitude for separation and heave measurements, with very high cumulative power law regression exponents of 1.7–2.0 for one-dimensional single horizon data, and 1.5–2.2 for two-dimensional multihorizon profile data. These distributions imply that the PTZ comprises a dense fault and fracture network and that there are large numbers (hundreds to thousands) of small-displacement protothrusts below the resolution of the seismic data.

A general model for frontal accretion at the Hikurangi margin involves a migrating wave of protothrust deformation that advances east of the frontal thrust in response to progressive seaward propagation of the basal décollement and the development of new frontal thrust faults. Newly developing major thrusts comprise low-angle (<30°) basal ramps that link with and inherit the earlier formed protothrusts. This sequence results in listric geometry
and a stratigraphically localized displacement low preserved toward the base of the clastic units. Strain localization onto the newly developing thrust front progressively deactivates the inner portion of the PTZ, and drives new PTZ deformation seaward.

Calculation of tectonic shortening accommodated by the active PTZs east of the present frontal thrust, from measurements of seismically imaged fault separations and estimates of subseismic faulting derived from one dimensionally sampled power law relationships, reveal their surprisingly significant role in accommodating regional convergence. The active PTZ could have accommodated as much as 4 km (north PTZ) and 7.4 km (south PTZ) of shortening, representing 14% and 23% respectively, on faults with have ≥1 m. This indicates that as much as 90% of the total shortening across the PTZ could be occurring at stratigraphic scale, an estimate consistent with other global data incorporating fault displacements and porosity reduction derived from seismic data.

From consideration of the maximum possible shortening rate per kilometer across the active PTZ, combined with estimates of stratigraphic ages and deformation duration, we infer that protothrusts across the total width of 60 km may accommodate 10%–50% of the total margin-normal convergence rate at the Hikurangi margin.

ACKNOWLEDGMENTS

Seismic lines SO191-1, SO191-4, and SO191-6 were acquired in 2007 with funding awarded to Joerg Biasia through expedition SO191 NewVents, BMBF (Federal Ministry of Education and Research) Germany contract 03G0191A, IFM-GEOMAR (Germany). These post-stack time-migrated seismic sections were processed on board R.V. Sonne by Geoffroy Lamarche (NIWA, National Institute of Water and Atmospheric Research) and Ingo Pecher (previously GNS Science, currently University of Auckland, New Zealand). Section 05CM-38 pre-stack depth migration was provided by Stuart Hnns (GNS Science) and Andreia Plaza-Faverola (previously GNS Science, currently University of Auckland, New Zealand). Helen Neil and John Mitchell (NIWA) coordinated and led the acquisition of Kongsberg EM302 multibeam bathymetric data during R.V. Tangaroa survey TAN1510. We appreciate the contribution of the shipboard party involved in the collection and processing of this data. Midland Valley is gratefully acknowledged for providing NIWA with Move software under their Academic Software Initiative.

We thank Andy Nicol (University of Canterbury, Christchurch, New Zealand) for discussion and advice regarding the analysis and interpretation of fault displacements and the treatment of the fault population data. We acknowledge funding from the Royal Society of New Zealand through Marsden grant GNS1204, and NIWA CORE Coasts and Ocean Project COPR1802–1802. We thank Greg Moore, Laura Wallace, and an anonymous reviewer for constructive reviews.

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