Durham Research Online

Deposited in DRO:
13 June 2017

Version of attached file:
Accepted Version

Peer-review status of attached file:
Peer-reviewed

Citation for published item:

Further information on publisher’s website:
https://doi.org/10.1093/gji/ggx260

Publisher’s copyright statement:
This article has been accepted for publication in Geophysical Journal International ©: 2017 The Author. Published by Oxford University Press on behalf of The Royal Astronomical Society. All rights reserved.

Additional information:

Use policy

The full-text may be used and/or reproduced, and given to third parties in any format or medium, without prior permission or charge, for personal research or study, educational, or not-for-profit purposes provided that:

- A full bibliographic reference is made to the original source
- A link is made to the metadata record in DRO
- The full-text is not changed in any way

The full-text must not be sold in any format or medium without the formal permission of the copyright holders.

Please consult the full DRO policy for further details.
Structural variability of the Tonga-Kermadec forearc characterised using robustly constrained geophysical data

Authors: Funnell M.J.¹, Peirce, C.¹ and Robinson, A.H.¹

¹ Department of Earth Sciences, Durham University, South Road, Durham, DH1 3LE, UK.

Accepted in original form

Abbreviated title: Tonga-Kermadec forearc structural variability

Corresponding author:
Matthew Funnell
Department of Earth Sciences
Durham University
DH1 3LE
Durham
United Kingdom
Email: matthew.funnell2@durham.ac.uk
Structural variability of the Tonga-Kermadec forearc characterised using robustly constrained geophysical data

Funnell M.J.\(^1\), Peirce, C.\(^1\) and Robinson, A.H.\(^1\)

SUMMARY

Subducting bathymetric anomalies enhance erosion of the overriding forearc crust. The deformation associated with this process is superimposed on pre-existing variable crustal and sedimentary structures developed as a subduction system evolves. Recent attempts to determine the effect and timescale of Louisville Ridge seamount subduction on the Tonga-Kermadec forearc have been limited by simplistic models of inherited overriding crustal structure that neglect along-strike variability.

Synthesis of new robustly tested seismic velocity and density models with existing datasets from the region, highlight along-strike variations in the structure of the Tonga-Kermadec subducting and overriding plates. As the subducting plate undergoes bend-faulting and hydration throughout the trench-outer rise region, observed oceanic upper- and mid-crustal velocities are reduced by \(-1.0\) km s\(^{-1}\) and upper mantle velocities by \(-0.5\) km s\(^{-1}\). In the vicinity of the Louisville Ridge Seamount Chain (LRSC), the trench shallows by 4 km and normal fault throw is reduced by \(>1\) km, suggesting that the subduction of seamounts reduces plate deformation. We find that the extinct Eocene frontal arc, defined by a high velocity (7.0-7.4 km s\(^{-1}\)) and density (3.2 g cm\(^{-3}\)) lower-crustal anomaly, increases in thickness by \(-6\) km, from 12 to \(>18\) km, over 300 km laterally along the Tonga-Kermadec forearc. Coincident variations in bathymetry and free-air gravity anomaly indicate a regional trend of northward-increasing crustal thickness that predates LRSC subduction, and highlight the present-day extent of the Eocene arc between 32°S and \(-18°\)S. Within this framework of existing forearc crustal structure, the subduction of seamounts of the LRSC promotes erosion of the overriding crust, forming steep, gravitationally unstable, lower-trench slopes. Trench-slope stability is most likely re-established by the collapse of the mid-trench slope and the trenchward side of the extinct Eocene arc, which, within the framework of forearc characterisation, implies seamount subduction commenced at \(~22°\)S.

Key words: Controlled source seismology; Crustal imaging; Crustal structure; Subduction zone processes.

1 INTRODUCTION

Island arc-backarc systems record the characteristics of subduction over time as changes in sedimentary and crustal structures on the overriding plate (Dickinson & Seely 1979; von Huene & Scholl 1991; Reagan et al. 2013). Variations in the strength and roughness of the subducting plate are manifest as changes in the rate of frontal and basal subduction erosion (Clift & Vannucchi 2004; von Huene et al. 2004), and in the dominant stress regime acting on the overriding plate (Bonnardot et al. 2007). Although the trench-proximal regions of erosive
The subduction of large bathymetric anomalies, such as seamounts and aseismic ridges, results in accelerated erosion and complex deformation patterns (Lallemand et al. 1992; von Huene et al. 1997; Dominguez et al. 1998; Zeumann & Hampel 2015). Recent research into the nature and effects of LRSC subduction along the Tonga-Kermadec margin has generated a range of theories regarding the subducted ridge strike and location of collision onset. A westward bend in the strike of the chain, to varying degrees, has been proposed at the current trench-ridge intersection by Timm et al. (2013) and Bassett & Watts (2015), whilst others suggest or assume linearity of the bathymetric trend of the ridge into the subduction zone (e.g. Ruellan et al. 2003; Bonnardot et al. 2007; Stratford et al. 2015). Further uncertainty surrounds the onset location of LRSC subduction, which is proposed to be either at the northern end of the subduction zone (e.g. Ruellan et al. 2003; Bonnardot et al. 2007), or to the north of Horizon Deep Bight, ~22.5°S, where the Tonga trench exhibits an ~80 km trench offset (von Huene & Scholl 1991; Lallemand et al. 1992; Wright et al. 2000; Contreras-Reyes et al. 2011; Stratford et al. 2015). These contrasting theories imply that there are significant gaps in our understanding of how LRSC-subduction alters and deforms existing forearc structures.

In order to provide better constraint on the variability of existing forearc structures, and the effect LRSC subduction has on them, a comprehensive margin-wide forearc structural model is required. We present new wide-angle (WA) and multichannel seismic (MCS) data that image sedimentary, crustal, and mantle structure across the LRSC and Kermadec forearc, perpendicular to seamount chain strike. These data are synthesised with results from previously published geological and geophysical studies (e.g. Crawford et al. 2003; Contreras-Reyes et al. 2011; Meffre et al. 2012; Funnell et al. 2014; Stratford et al. 2015) as well as satellite-derived gravity anomaly data (Sandwell et al. 2014) to: i) constrain variations in the structure and deformation of the Pacific plate crust around the Tonga-Kermadec trench-LRSC intersection; ii) characterise along-strike variations in the inherited (pre-LRSC subduction) Tonga-Kermadec forearc crustal structure; and using these classifications, iii) determine the temporal and spatial extent of forearc deformation caused by LRSC subduction.
2 GEOLOGICAL SETTING

The Pacific plate subducts to the west along the ~2700 km-long, 020° trending, Tonga-Kermadec trench axis (Fig. 1). Evidence of volcanism throughout the Eocene (Bloomer et al. 1995), and as early as ~51 Ma (Meffre et al. 2012; Michibayashi et al. 2012), suggests that subduction initiation at this margin was concurrent with that along the Izu-Bonin-Mariana margin (Reagan et al. 2013). At the time of inception, volcanism was focussed along the now extinct Tonga arc, which forms the present-day forearc ridge (Fig. 1). Between 45 and 34-32 Ma, the forearc and arc subsided rapidly from a sub-aerial setting to ~3 km below the sea surface, causing the trenchward tilting of sedimentary intervals (Clift & MacLeod 1999). Subsequent to the initial period of volcanism, the Lau-Colville Ridge became the active arc and the South Fiji Basin opened until ~17 Ma (Herzer et al. 2011). During this period, seamount subduction is attributed with the arcward tilting of sediments and generating an erosional unconformity, prior to 16.2 Ma, on the Tonga forearc (MacLeod 1994; Clift & MacLeod 1999). At ~5.5 Ma, a shift in tectonic forces caused the opening of the Lau Basin and the Havre Trough (Ruellan et al. 2003), resulting in an apparent ~2 Myr volcanic hiatus and the migration of volcanism to the currently active Tofua and Kermadec arcs (Clift et al. 1994; Bloomer et al. 1995), which are roughly parallel to the trench. Collision of the LRSC with the Tonga-Kermadec subduction system is suggested to have initiated 5-4 Ma or >7 Ma (Ruellan et al. 2003 and Timm et al. 2013, respectively), and is observed at ODP site 841 as an increase in the trenchward rotation of the forearc at ~1.5-0 Ma (Fig. 1 – MacLeod 1994; Clift & MacLeod 1999).

Stresses associated with Pacific plate subduction along the Tonga-Kermadec margin generate large extensional faults throughout the downgoing oceanic crust and upper mantle (Pelletier & Dupont 1990; Contreras-Reyes et al. 2011). Similar normal extensional faults observed along other trenches, e.g. Chile and Costa Rica, are proposed to enable hydration of the upper mantle (Ranero et al. 2003; Ivandic et al. 2010; Moscoso & Grevemeyer 2015), facilitating a reduction in the flexural rigidity of the subducting plate (Billen & Gurnis 2005; Arredondo & Billen 2012). The oceanic crust subducting along the Tonga-Kermadec trench formed along the Osbourn spreading centre (Fig. 1) roughly perpendicular to the present-day convergence direction (DeMets et al. 2010), prior to ~90 Ma (Billen & Stock 2000; Downey et al. 2007). Shortly after crustal formation, hotspot magmatism generated the bathymetrically prominent LRSC (Lonsdale 1988; Vanderkluysen et al. 2014). The relatively young host-plate age is associated with the highly intrusive nature of the seamount cores and limited crustal thickening observed at the northwestern end of the seamount chain (Contreras-Reyes et al. 2010; Richards et al. 2013). Volcaniclastic sedimentation supplements the <200 m of sediment cover that is typical for the southwestern region of the Pacific plate (Burns & Andrews 1973).

The structure of the overriding plate varies significantly with distance from the trench, and along the length of the Tonga-Kermadec margin (Fig. 1). Limited sediment cover (<100 m) and a steeply dipping (10–24°), highly irregular, basement characterises the lower-trench slope (Karig 1970; Ballance et al. 1999). Mid-slope basins are observed along the entire length of the subduction system, although they are less prominent directly north of the present-day LRSC-trench intersection point (e.g. Karig 1970; Ballance et al. 1999; Funnell et al. 2014; Stratford et al. 2015). Low seismic velocities of 3.5-4.0 km s\(^{-1}\) up to 5 km below seafloor, and pervasive
normal faulting observed in cores at ODP site 841 (Clift et al. 1994; Contreras-Reyes et al. 2011; Stratford et al. 2014) suggest that the lower- and mid-trench slopes are subject to significant extensional regimes controlled by subduction erosion (von Huene et al. 2004; Sallares & Ranero 2005). Between the mid-trench slope and the active volcanic ridge the extinct Eocene arc, known as the Tonga Ridge, is bathymetrically expressed from Vava’u, the northernmost island in the Tongan island group, to ~28°S (Ballance et al. 1999; Contreras-Reyes et al. 2011). South of ~28°S, the forearc is defined by ~2 km deeper, sediment filled basins (Karih 1970; Funnell et al. 2014). The active Tofua and Kermadec arcs lie 170-200 km west of their respective trenches (Ballance et al. 1989), and distribute volcanioclastic material into the surrounding forearc and backarc basins (Karih 1970). Active backarc extension causes the separation of the Tonga and Kermadec arcs from the Lau-Colville backarc, with higher spreading rates and more complex plate structure in the northern Lau Basin (Zellmer & Taylor 2001; Ruellan et al. 2003; Conder & Wiens 2011).

There are a number of major along-strike boundaries that delimit major structural and deformational variations. South of 32°S an anomalous increase in trench depth is associated with a westward step in the forearc and deepening of the mid-slope terrace (Karih 1970; Pelletier & Dupont 1990). Ballance et al. (1999) noted that these features also coincided with a marked change in backarc basin structure, with steeper basin-bounding faults to the south and the seafloor reaching ~500 m deeper. This feature has been attributed to different mechanisms: the initial point of Hikurangi Plateau subduction (Collot & Davy 1998), increased subduction erosion (Pelletier & Dupont 1990; Ballance et al. 1999), and the southern extent of the extinct Tonga Ridge (Bassett et al. 2016).

Further north, subduction of the LRSC enhances subduction erosion (Ballance et al. 1989), causing the Tonga trench to shift westward by ~80 km (Clift & MacLeod 1999). This process is thought to have formed Horizon Deep Bight, the deepest part of the Tonga trench, and the steepest lower-trench slope, >10°, immediately north of the present-day seamount subduction region (Fig. 1; Lonsdale 1986; Contreras-Reyes et al. 2011). The active region of LRSC subduction-associated forearc deformation lies roughly arcward of Osbourn seamount, the oldest bathymetrically expressed seamount in the chain. The deformation is indicated by a shallowing of the trench axis by ~4 km (Ballance et al. 1989), forearc uplift of 300 m (Stratford et al. 2015), and the presence of a gap in seismicity (Bonnardot et al. 2007).

At ~18°S a much smaller aseismic discontinuity coincides with the northern extent of the bathymetric Tonga Ridge (Crawford et al. 2003; Bonnardot et al. 2007). Dredged rock ages decrease northwards across the boundary (Meffre et al. 2012), and the backarc basin becomes increasingly complex with extension migrating from the Eastern Lau Spreading Centre to the Central Lau Spreading Centre and Fonualei Rift (Zellmer & Taylor 2001; Ruellan et al. 2003; Conder & Wiens 2011).

3 DATA ACQUISITION AND PROCESSING

Profile B (Fig. 2), a coincident MCS and WA seismic line, was acquired during R/V Sonne cruise SO215 (Peirce & Watts 2011). East to west, the profile crosses Pacific crust and Canopus seamount, at ~65° from the palaeo-spreading direction (Billen & Stock 2000), as well as the Kermadec forearc and arc south of the present-
day LRSC-trench intersection. Swath bathymetry and gravity data were acquired contemporaneously. Data from Profiles A and D (Fig. 1), which were also acquired during cruise SO215, are further discussed in this paper (Funnell et al. 2014; Stratford et al. 2015).

3.1 Swath bathymetry data

A hull mounted SIMRAD EM120 multibeam echosounder acquired depth soundings throughout the transit and shooting phases of cruise SO215. This swath bathymetry dataset was cleaned to remove bad soundings using MB-System (Caress & Chayes 1996), merged with swath data from previous cruises, and gridded at ~50 m node spacings to provide good constraint on the depth of the seabed (Fig. 2). Expendable bathymetric-thermographs, deployed at ~100 km intervals, recorded the seismic velocity of the water column to a depth of 1.8 km.

3.2 Seismic data source

Contemporaneous acquisition of the WA and MCS data required a 60 s shot interval to prevent water wave wrap-around on the ocean-bottom seismographs (OBSs), whilst still achieving a reasonable MCS fold. The seismic source consisted of a 12 G-gun array, with a combined volume of 5440 in³ (84 l), towed 7.5 m below the sea surface and fired at 3000 psi (210 bar).

3.3 MCS data acquisition and processing

MCS reflection data were recorded using a 3 km-long, 240-channel streamer with 12.5 m group interval, that was towed at 10 m depth. At the average ship speed of ~4.5 kn, a shot spacing of ~150 m yielded a maximum theoretical fold of 10. Given the low fold, 25 m super-bins were used for data processing, which increases the average fold of cover from 10 to ~40 and significantly improves the reflectivity of the intra-sediment and basement reflectors. Data were recorded at a 2 ms sampling interval in 59 s trace lengths.

A bulk time-shift of 548 ms corrected for the shot-receiver datum as well as aim point and shot triggering delays. A Butterworth filter (bandpass: 3-10-100-120 Hz) reduced noise generated by the significant swell encountered during SO215, increasing the SNR in the usable frequency bandwidth.

Initial processing of the MCS data involved detailed velocity analysis, undertaken at 50 CMP intervals using semblance values as well as constant velocity gathers and stacks to generate a stacking velocity model with 1.25 km lateral resolution. Following the application of normal moveout correction and stacking, a finite difference time-migration operator removed the strong scattered signal, particularly at the seabed in heavily faulted regions. A seabed mute removed signal in the water column for display purposes.

3.4 WA seismic dataset

WA seismic data were recorded by 26 OBSs deployed ~15 km apart along Profile B, with a 30 km gap at the Kermadec trench due to the maximum instrument depth rating of 8 km (Peirce & Watts 2011; Fig. 2a). A hydrophone, vertical geophone, and two horizontal geophone components output analogue signals that were
digitised and recorded at 4 or 5 ms sampling intervals on the LC-type and KUM deep-water type instruments respectively.

Instrument-seabed coupling and variations in geological structures between source-receiver pairs control the SNR and the characteristics of seismic phases recorded by each instrument. Along Profile B the WA data are divided into three groups according to significant changes in these characteristics: OBSs 01-10 on the Kermadec forearc, 11-15 in the trench, and 16-27 on the Pacific plate.

3.4.1 Kermadec forearc OBSs 01-10

Sediment turning P-waves, $P_t$ (2.2-3.5 km s$^{-1}$), are only observed on the westernmost instruments (01-04) on the Kermadec forearc, suggesting significant sediment accumulations are only located proximal to the Kermadec arc (Fig. 3). Further east, crustal arrivals, $P_g$ (4.0-7.0 km s$^{-1}$), are observed as the shortest offset refractions (from ~4.5 km), suggesting poor sediment cover on the trench slopes. The SNR of crustal refractions on forearc instruments is highly variable, although, if they are observed, the arrivals are typically found at much greater offsets than those of the Pacific plate (up to 70 km offset). Mantle refractions, $P_n$ (7.5-8.0 km s$^{-1}$), are less clear and are typically observed between 50-100 km offset.

3.4.2 Kermadec trench OBSs 11-15

Instruments located in the Kermadec trench display the most complex pattern of arrivals and the lowest SNR (e.g. Fig. 4). Crustal refracted arrivals indicate abnormally low velocities, between 3.5 and ~6.5 km s$^{-1}$, on the lower slope of the forearc. On the outer trench slope crustal velocities remain closer to those observed by the other Pacific plate OBSs (4.0-7.0 km s$^{-1}$). Incoming shots to trench-located instruments are observed to typical offsets for a given phase only if they pass through the plate on which the OBS is located. Signal that passes through the subduction interface is significantly attenuated, resulting in the rapid degradation of arrival clarity beyond the trench apex. As such, mantle refractions are rarely observed at this point along the profile and, although some rays appear to be transmitted between the overriding and subducting crustal units, these have a very high noise component.

3.4.3 Pacific plate OBSs 16-27

OBSs located on the Pacific plate recorded low-velocity (4.0-5.0 km s$^{-1}$) crustal turning rays as the shortest offset refractions (Fig. 5). No sediment arrivals were observed on the Pacific plate instruments due to the relatively thin sediment cover. Instruments located close to, or on, Canopus seamount display crust and mantle arrivals that are perturbed in time by the highly undulatory seafloor. Mantle turning arrivals are observed between 30 and <190 km offset from the instruments.

3.4.4 Pick method and uncertainty assignment

First and secondary arrivals were picked on either the hydrophone or vertical geophone record for each instrument, depending on which component displayed the higher SNR. Pick phases were assigned based on the offset and apparent velocity of the arrivals.
Pick uncertainties were calculated independently for each OBS based on instrumentation and location errors, as well as the qualitatively assessed SNR for a given pick. Assigned uncertainty values typically increase with shot-receiver offset, and thus turning depth. Instruments located in and around the trench had particularly high location errors due to the extreme water depth and strong water column current, and a low SNR due to the increased sub-seabed complexity (Table 1).

3.5 Gravity data acquisition

Gravity data were acquired port-to-port using a Lacoste & Romberg Air-Sea gravimeter. These data were processed, incorporating Eötvös and drift corrections, and tied to absolute stations to calculate the absolute free-air anomaly (FAA). Crossover analysis gives an average root mean square misfit, and thus dataset uncertainty, of ~10 mGal which, although relatively high, is most likely generated by the high gravity gradients in the region (Peirce & Watts 2011).

4 2D MODELLING

4.1 Forward velocity modelling

Forward ray tracing was chosen as the primary velocity modelling technique for the OBS dataset (Table 1) because of the extreme lateral heterogeneity expected and typical of subduction zone velocity structure. Profile B OBS locations, shotpoints, and bathymetry data were projected from geographic coordinates into distance along profile (Fig. 6), for modelling using rayinvr (Zelt & Ellis 1988; Zelt & Smith 1992).

4.1.1 Model initialisation and modelling method

The velocity model was initialised with the water column thickness and velocity along the profile, determined from high-resolution swath bathymetry and expendable bathymetric-thermograph data respectively. OBS instrument locations were refined in the model space by relocating instruments laterally and vertically to reduce the misfit between the modelled and observed water wave arrivals. Initial sediment layer velocity and thickness were derived from the first-pass processed stacking velocities and the interpreted basement reflector from the MCS data (Fig. 2b), and draped beneath the model seabed. Below this, a simple velocity gradient was applied down to the bottom of the model.

Improvements to the fit between the calculated and observed arrival times were achieved by iteratively changing either the depths of model layer boundaries or the velocities at the top and bottom of that layer, following the method of Zelt & Smith (1992). This approach was first applied to the shallow layers (i.e. the sediment column), improving the fit on a subset of instruments in a rolling window along the length of the model before moving onto deeper layers. Additional boundary nodes and layers were only added to the model when an acceptable fit could not be achieved with the existing model elements. Nodes are typically spaced at 2.5 km intervals for the seabed and sediment column, 5 km for the intracrustal layers, and >10 km at the Moho. In regions where the ray coverage was poor, such as through the sediment column or at the model edges, the model was kept as close to the background structure as possible.
### 4.1.2 Misfit minimisation and the final velocity model

Following the achievement of a reasonable qualitative agreement between the modelled and observed datasets, the travelt ime root mean square residual, $T_{\text{RMS}}$, and uncertainty-weighted factor, $\chi^2$, enabled further model improvements to be made within the uncertainty limits of observed data. The $T_{\text{RMS}}$ and $\chi^2$ misfits for specific phases in the major regions of the model are presented with the pick uncertainties in Table 1. Areas of the model with higher misfit, such as the trench-proximal region ($T_{\text{RMS}}=113$ ms) and the lower crust of the forearc ($T_{\text{RMS}}=83$ ms) coincide with groups of instruments that exhibit poor ray coverage or image more geologically complex structures.

The final velocity model, shown in Fig. 6b, has a total $T_{\text{RMS}}$ misfit of ~85 ms, and a $\chi^2 < 2$. The model has nine layers that represent distinct geological units: three crustal layers and a mantle block, for each of the overriding and subducting plates, and a laterally continuous sedimentary layer.

### 4.2 Forward model sensitivity testing

Forward model fit sensitivity was tested by adjusting individual layer thickness, upper and lower boundary velocity (variable velocity gradient), and bulk layer velocity (constant velocity gradient) until a predefined misfit threshold is exceeded. A single threshold value for the entire model was found to allow regions of relative geological simplicity and better fit, such as the Pacific plate, more variability before exceeding the limit and thus appearing less sensitive than areas of the model that fit less well. This result can be misleading, so an improved understanding of model sensitivity was obtained by also testing the three main regions of the model separately, with more realistic thresholds chosen to match the set of instruments represented.

The results of the sensitivity tests, shown in Table 2, indicate that intra-crustal regions of both the subducting and overriding plates are sensitive to relatively small changes in layer depth (~0.3 km), upper or lower boundary velocity (~0.3 km s$^{-1}$), and bulk layer velocity (0.2 km s$^{-1}$). Although the forearc sediments are similarly sensitive, the lack of turning arrivals through the Pacific plate sediments gives minimal control on the shallow velocity structure (up to a 0.5 km s$^{-1}$ shift in upper or lower seismic velocity). Conversely, the Pacific plate region of the model is almost twice as sensitive to changes in Moho depth and upper mantle boundary velocity (0.5 km and 0.3 km s$^{-1}$ respectively) when compared to the forearc (0.7 km and 0.8 km s$^{-1}$ respectively). Model sensitivity in the measurable characteristics of the trench region is found to be roughly half that of the analogous features in the overriding and subducting plates. For the remainder of this paper, significant layer thicknesses and velocities determined from the forward velocity model will be quoted with the relevant layer sensitivity to indicate confidence in the results.

### 4.3 Inversion modelling

We used the inversion method of Zelt & Barton (1998) as an independent modelling procedure to test the reproducibility of long-wavelength velocity structure, the degree of modeller-bias, and resolution of the final best-fit WA model (as in Zelt et al. 2003). Model perturbations were limited by a set of velocity constraints to prevent geologically unrealistic alterations, and smoothed to minimise the acquisition geometry influencing the outcome.
The starting model had a 1.5 km-thick region of high velocity gradient, from 1.5-6.0 km s$^{-1}$, draped beneath the seabed representing the sediment column and shallow crust (Fig. 7a). Below this velocities increase, reaching 8.0 km s$^{-1}$ at 15.0 km depth to represent a constant-depth Moho, although no first-order velocity discontinuity is imposed. This starting model, which was chosen because of the variability in crustal thickness along the profile, enabled the greatest modelling freedom within geological reality. We found that the optimal modelling process used two runs of five non-linear inversion iterations, with the inversion cell size reduced between the runs. During run 1, 10 km by 1 km inversion cells improved the larger-scale velocity structure of the model (Fig. 7b). This reduced the misfit to ~100 ms ($\chi^2$=6.8), with the greatest misfit remaining on the overriding plate and in the trench. For run 2, 5 km by 0.5 km inversion cells refined small-scale and shallow structures, most notably in the overriding crust (Fig. 7c), resulting in a further reduction of the misfit to 70 ms ($\chi^2$=3.1). The final inversion velocity model (Fig. 7c) displays a velocity structure that is broadly similar to the forward model (Fig. 7d) and, often noted in inversions, acquisition geometry-related smearing is absent. The Pacific crust and upper mantle are represented by typical oceanic-velocity structure, except for a poorly defined Moho (Fig. 7c). As with the forward model, Canopus seamount exhibits low velocities (<6.0 km s$^{-1}$) within the volcanic cone. In the overriding plate, the inversion model displays thick regions of low velocity in the shallow crust (<8.0 km-below seabed) and throughout the lower-trench slope crust (170-220 km-offset). Regions with poor ray coverage, such as the lower forearc slope (Fig. 7e), remained poorly fitting following run 2 but are broadly comparable to the forward model and remain within the indicated sensitivity values. The relatively high misfit of these regions is most likely due to model smoothing factors and the inversion cell sizes significantly exceeding the true scale of subsurface velocity variability (e.g. Zelt et al. 2003). Although the forearc Moho is very vertically diffuse, it appears to be ~2 km shallower between 70 and 150 km model offset for the inverse model relative to the forward model. The final forward velocity model remains our ‘preferred’ velocity model, because it is better constrained by additional data (MCS information and WA-reflected phases), is better resolved, and is more geologically representative. The inversion velocity model indicates the minimum structure required by the first arrival dataset and, thus, highlights the most robust features that are shared by both models (e.g. Zelt et al. 2003).

4.4 Inversion model resolution testing

Checkerboard testing of the final inversion velocity model, using a 5% velocity perturbation, determined the smallest structures resolvable by the acquisition geometry and model parameterisation (Zelt 1998). Fig. 7f indicates that velocity anomalies larger than 30 km by 3 km should be well resolved throughout regions of the model with >200 rays per forward node (Fig. 7e). In regions with particularly high ray coverage, such as in the mid-crustal region of the Pacific plate, structures as small as 20 km by 2 km may be resolvable (Fig. 7g). Ray coverage and the recovery of the input checkerboard are consistently poorest between 150 and 220 km offset, in the trench region. These results indicate that the internal velocity structure of LRSC seamounts may be recoverable with acquisition geometries similar to that along Profile B. Despite this, the checkerboard tests suggest that the velocity perturbations within the lower-trench slope region of the overriding crust are so poorly resolvable, due to the high pick errors and low ray density, that a velocity anomaly would have to be larger than
that expected to be generated by an entire intact seamount to be determinable. To achieve the accurate and robust imaging of such a feature, greater ray densities would be required, either through the addition of more instruments, or the addition of more shots in the region.

4.5 MCS reprocessing and fit to WA data

Although of a lower lateral and vertical resolution than the MCS stacking velocity model through the sediment column (1.25 km), the final WA forward velocity model (~2.5 km) should better represent true subsurface velocities. Conversion of the robustly tested and well resolved velocity model from depth to the time domain enabled the restacking and migration of the MCS dataset to produce a more geologically representative seismic section (e.g. Fig. 2d and f).

Restacking focused a band of deeper reflectivity beneath the flat-lying segments of the Pacific crust, at 9-10 s TWTT, that was previously unseen in the initial stack based on velocity analysis alone (compare Fig. 2e and f). This reflector fits to within 0.5 km of the forward velocity model crust-mantle boundary, suggesting it represents the Moho. The Moho reflector is not observed beneath the forearc due to the high amplitude water column multiple energy. The top-crust reflector fits within 0.1 km of the respective forward model boundary throughout the section except in the vicinity of the thicker upper forearc basin, which is most likely due to the scale of basement variations being below the lateral sensitivity of the forward modelling method. Despite the reduced lateral resolution and slight mismatch in the depth of the top-crust, there is no evident loss of the intra-sediment reflector information in the restacked MCS reflection data (Fig. 2d). This therefore provides further confidence that the WA velocity model is a good representation of the velocity structure of the crust along the profile.

4.6 Gravity anomaly modelling

The observed gravity anomaly provides an independent test of velocity model uniqueness and gives additional constraint on poorly resolved regions. The final forward velocity model was, therefore, converted into a density model consisting of 2D polygons. The expected free-air gravity anomaly was calculated from these using grav2d, based on the algorithm of Talwani et al. (1959).

Model edge effects were reduced by extending the density structure for 1000 km beyond the model edges, draped beneath the bathymetry extracted from the GEBCO 30 arc-second dataset (IOC et al. 2003). The inclusion of the subducting slab and mantle structure to 100 km depth, based on the extension of the velocity model using expected slab dips from slab1.0 (Hayes et al. 2012), removed the initial long-wavelength gravity anomaly mismatch.

A layered crustal block-model, where the average velocity through each layer of the forward velocity model determined the density of its respective block, generated the best-fitting gravity anomaly (Fig. 8). Standard velocity-density relationships were applied for the sediment column (Nafe & Drake 1957) and the subducting Pacific oceanic crust (Carlson & Raskin 1984), with a good fit found for the subducting plate. Continental (Christensen & Mooney 1995) and oceanic crustal velocity-density relationships were tested for the overriding
crust and, although both failed to generate the high-amplitude (~200 mGal) forearc anomaly, the oceanic crustal relationship was found to produce the closest fit (~30 mGal). This suggests that the forearc crust may be of a predominantly mafic, rather than felsic, composition. The addition of a high-density anomaly (3.2 g cm$^{-3}$) at the base of the crust, coincident with the region of high velocity between 50 and 150 km model offset, significantly decreased the misfit of the entire model to 12.9 mGal.

5 COMBINED MODEL DESCRIPTION AND INTERPRETATION

5.1 Background Pacific plate

Along Profile B the Pacific plate displays highly variable basement reflector depths (Fig. 2b). Sediment thicknesses typical for the region of 100-200 m are only observed at ~410 km model offset and instead reach >1 s TWTT (~1.5 km at 3.0 km s$^{-1}$) in the downthrown segment of a normal fault at ~415 km offset (Burns & Andrews 1973; Funnell et al. 2014). Despite this, typical oceanic crustal velocities are observed at the northeastern end of Profile B (Fig. 9b). The oceanic crust is ~6.0 ± 0.3 km thick, consisting of an ~1.5 ± 0.3 km thick upper- and middle-crust (3.5-6.5 ± 0.2 km s$^{-1}$), and a 4.5 ± 0.3 km thick lower crust that reaches 7.1 ± 0.3 km s$^{-1}$ at the Moho. Upper mantle velocities are consistently low, ~7.6 ± 0.3 km s$^{-1}$, compared to typical Pacific plate velocities of ~8.2 ± 0.3 km s$^{-1}$ (Fig. 9b; Spudich & Orcutt 1980; White et al. 1992). The ~65$^\circ$ orientation of Profile B to the palaeo-spreading, or fast seismic wave propagation direction of the Osbourn spreading centre (Fig. 1), may cause these upper mantle seismic velocities to be close to the lower limit of what may be expected from typical azimuthal mantle anisotropy of 4-7%, which is controlled by preferential olivine crystal alignment (Kodaira et al. 2014).

5.2 Canopus seamount (LRSC)

Sediments in the moat around Canopus seamount reach ~1.5 km thick, with reflectors typically dipping away from the seamount. Although the sediment layer thickness was insufficient to observe or model turning rays, the inferred average layer velocity of ~2.5 km s$^{-1}$, and internal reflectivity, suggests that the thick sediment accumulation consists of volcaniclastic basalts and glass associated with seamount emplacement (Koppers et al. 2012; Vanderkluysen et al. 2014). Along Profile B, the internal velocity structure of Canopus seamount increases from 3.0 to 5.5 ± 0.3 km s$^{-1}$ over the first 5 km below the peak. These relatively low velocities suggest that the shallow seamount structure imaged comprises highly porous extrusive and intrusive basaltic rocks more typical of upper oceanic crust (White et al. 1992; Koppers et al. 2012). Contreras-Reyes et al. (2010) observed velocities of 6.4 km s$^{-1}$ less than 2 km beneath the summit of the seamount immediately southeast of Canopus, and proposed that they represent a highly intrusive mafic core. Although such high internal velocities are not observed on Profile B, their absence could be explained as an artefact of sampling location and limited intrusion width, as Profile B crosses the saddle between the two peaks of Canopus seamount (Fig. 2a). Elevated velocities of ~7.4 ± 0.3 km s$^{-1}$ at the base of the crust beneath the seamount are comparable to the seismic structures imaged at adjacent seamounts (Contreras-Reyes et al. 2011; Stratford et al. 2015). These anomalously high velocities, and
corresponding slightly thickened crust, extend over 50 km from the seamount summits, suggesting that the LRSC seamounts are supported by broad lower-crustal intrusions (Contreras-Reyes et al. 2011; Richards et al. 2013).

5.3 Trench-outer rise region

Between Canopus seamount and the trench, bend-related normal faulting is manifest as <0.1-0.3 km trenchward-dipping reflector offsets that preserve a sub-horizontal seabed (Aubouin 1989; Ranero et al. 2003). This deformation pattern is significantly different to the 2.0 km-throw horst and graben structures observed on Profile D that coincide with an increased plate dip of ~5° (Fig. 10a & b; Funnell et al. 2014). Coincident with this region of faulting, between 0-55 km perpendicular distance from the trench, upper- and mid-crustal velocities are reduced by ~1.0 km s\(^{-1}\), lower-crustal velocities remain unchanged, and upper mantle velocities decrease by ~0.3 km s\(^{-1}\) to ~7.4 ± 0.3 km s\(^{-1}\).

5.4 Lower- and mid-trench slope

Large fault blocks and a rough seabed define the lower trench slope (Fig. 2a; Karig 1970). Sediment accumulations are limited to <0.1 s TWTT thick and are, for the most part, absent due to an ~10° slope. This region exhibits the lowest crustal seismic velocities in the model, increasing from 4.2 ± 0.6 km s\(^{-1}\) beneath the sediments to just over 6.0 ± 0.6 km s\(^{-1}\) at the base of the ~9 km-thick crust (Fig. 11c). The calculated gravity anomaly for the lower-trench slope only fits the observed data when densities derived from the local seismic velocity structure (i.e. kilometre scale) were ignored and, instead, entire forward model layers were converted into density blocks (Fig. 8), indicating that the velocity-density ratio that defines this model region is not typical of oceanic crust. Similar seismic wavespeeds and poor reflectivity are observed by Contreras-Reyes et al. (2011) and Sallarès & Ranero (2005) at the Tonga and North Chile margins respectively, and are attributed to the pervasive faulting, disaggregation, and hydration of the overriding crust in response to subduction erosion (e.g. Clift & MacLeod 1998; Clift & Vannucchi 2004).

Further west, between 160-170 km offset, an ~1.0 s TWTT (750 m-high) horst-and-graben structure offsets the lower-trench slope from a poorly imaged mid-slope sedimentary wedge. Sediments thicken to ~1.0 s TWTT and sedimentary reflectors tilt more arcward towards the east of the ~25 km-wide basin. Velocities of >6.0 km s\(^{-1}\) are reached just 4 km below the sediment base, and more typical lower crustal velocities are observed e.g. 6.8-7.2 ± 0.3 km s\(^{-1}\) (Holbrook et al. 1999; Contreras-Reyes et al. 2011).

5.5 Upper forearc slope and arc

Along Profile B, the upper forearc slope is defined by an ~50 km-wide ridge and basin in front of the active volcanic arc (Fig. 6a). The sediments beneath the ridge are ~1.0 s TWTT thick and relatively horizontal, although broken up by pervasive small-scale, <0.1 s TWTT throw, normal faults (Fig. 2c & d; Herzer & exon 1985). In the ~40 km-wide basin, an ~2.0 s TWTT thick sedimentary succession is divided by a high amplitude intra-sediment reflector at ~1.0 s TWTT-below seafloor. This high amplitude reflector pinches out onto the basement of the morphological ridge (Fig. 2d), suggesting that the underlying ridge and basin morphology predates the majority of sedimentation along the margin (Herzer & exon 1985). Reflectivity of the ridge and basin
sediments changes from chaotic to horizontal and planar with increasing distance from the present-day arc, which suggests that turbiditic flows transport and emplace volcaniclastic material along the forearc (Herzer & Exxon 1985; Gillies & Davey 1986).

In the forward velocity model, the crust beneath the upper-forearc slope is 10-12 ± 0.7 km thick and displays a laterally variable seismic velocity structure (Fig. 6b). Beneath the upper forearc basin, upper- and mid-crust velocities increase from ~3.0 to 6.2 ± 0.3 km s⁻¹ over 5 km in depth, and reach 6.8 ± 0.3 km s⁻¹ at the base of the crust. Mid-crustal velocities under the morphological ridge are ~2 km shallower than in the basin to the west and increase to ~7.4 ± 0.2 km s⁻¹ just above the Moho. Mantle velocities appear to be consistent beneath the forearc, at ~7.8 ± 0.8 km s⁻¹, although poor ray coverage limits model sensitivity in this region (Table 2).

Although the inversion model also converged to a solution with high velocity gradients in the shallow crust and a high-velocity lower crust beneath the morphological ridge (Fig. 7c), mantle velocities, >7.8 km s⁻¹, are found ~2 km shallower than in the forward model (Fig. 6b). This is the only significant discrepancy between the forward and inverse velocity models, and coincides with an unusually high FAA, ~150 mGal, which is significantly undercalculated by the density-converted forward velocity model using typical velocity-density relationships (Carlson & Raskin 1984; Christensen & Mooney 1995). Whilst assuming an ~2 km-thinner forearc crust beneath the morphological ridge improves the fit between the observed and calculated datasets, the misfit remains >30 mGal and forward-modelled Moho reflections no longer fit within the assigned uncertainties (Table 1). The preferred combined model maintains the crustal thicknesses determined by the forward modelling method, and includes an ~5.0 km-high lower-crustal density anomaly of ~3.2 g cm⁻³ to accurately recreate the forearc FAA, with a misfit of ~13 mGal. The discrepancy in crustal thickness between the forward and inverse velocity models, whilst significant, is within the vertical resolution of the inversion model (Fig. 7e), and most likely results from inversion model smoothing, and the inability to invert Moho-reflected arrivals (PₚP phases – Zelt et al. 2003).

Both of these scenarios require relatively thin crust compared to previous results from the Tonga, Izu-Bonin-Mariana, and Aleutian margins, although it is not entirely unusual (Fig. 11b). Kodaira et al. (2010) imaged <10 km thick forearc crust beneath the Bonin Ridge, attributing the thin crust with forearc spreading during subduction initiation. Similar inferences have been made regarding the origin of the Tonga Ridge, which is widely associated with the onset of west-dipping subduction ~51-48 Ma (Herzer et al. 2011; Meffre et al. 2012; Michibayashi et al. 2012). The forearc structural high observed along Profile B is at the southern end of the bathymetrically expressed Tonga Ridge (Korig 1970; Lonsdale 1986; Contreras-Reyes et al. 2011; Stratford et al. 2015), and most likely also formed during the initial stages of volcanism at the margin. Despite a lack of direct sampling of the lithologies and ages at the southern end of the bathymetrically exposed Tonga Ridge (Clift & MacLeod 1998), evidence from the Profile B MCS section suggests that the ridge-and-basin upper-forearc morphology predates the majority of sedimentation at the margin. As such, we expect the forearc structural high and associated lower-crustal velocity and density anomalies imaged along Profile B to be directly formed by, or developed from, the tholeiitic and boninitic volcanism associated with subduction initiation (Clift et al. 1994; Meffre et al. 2012; Michibayashi et al. 2012; Reagan et al. 2013). To achieve the high density and velocity required by the gravity and WA models, the lower crustal structure most likely comprises high-density pyroxenites
that are left as a residual lower-crustal material following initial subduction volcanism (Nakajima & Arima 1998; Meffre et al. 2012). We refer to this morphological ridge with high density and seismic velocity, as the extinct Eocene frontal arc.

6 DISCUSSION

6.1 Variations in subducting crust and upper mantle structure

Subduction-induced bending of the downgoing Pacific plate generates large-scale normal faults that are observed along the Tonga-Kermadec margin as significant seabed offsets of up to 2 km (Fig. 9a; Crawford et al. 2003), and as downthrown sedimentary units in seismic reflection data (Fig. 10a & b; Pelletier & Dupont 1990). Similar bend faulting is observed into the mantle of the Costa Rican margin (Ranero et al. 2003), and perversely throughout the Chilean trench-outter rise (Contreras-Reyes et al. 2008). Along the Chilean margin, the faults coincide with the reduced seismic velocity of the crust, by 0.4-0.7 km s\(^{-1}\), and upper mantle, by up to 0.5 km s\(^{-1}\) (Moscoso & Grevemeyer 2015), which are proposed to be the physical manifestation of oceanic crust hydration and upper mantle serpentinisation (e.g. offshore Costa Rica and Nicaragua - Ranero et al. 2005 and Ivandic et al. 2010, respectively). These pervasive structural and geochemical modifications are attributed with reducing the flexural rigidity of the subducting plate by up to three orders of magnitude over the trench-outter rise region (Billen & Gurnis 2005; Arredondo & Billen 2012). We find that the upper 2 km of the subducting oceanic crust reduces by \(-1\) km s\(^{-1}\) in seismic velocity with increasing proximity to the Tonga-Kermadec trench, over 100-170 km trench-perpendicular distance, whilst no discernable variations are observed in the lower 4 km (Fig. 9b). The seismic velocity of the uppermost mantle reduces by up to 0.5 km s\(^{-1}\) for profiles that are roughly perpendicular to the palaeo-spreading direction of the Osbourn spreading centre. This velocity reduction is comparable to that observed at the Central Chile margin (Moscoso & Grevemeyer 2015), but is at the limit of the resolution for the models presented here (Table 2; Contreras-Reyes et al. 2010; Contreras-Reyes et al. 2011; Stratford et al. 2015), so it is not a conclusive result. We note that this velocity reduction is of a similar magnitude to the decrease observed between the upper mantle velocities sampled by profiles perpendicular (Profile A, B, P02, and P03 - Fig. 9) and parallel (Profile P03 - Fig. 9) to the palaeo-spreading direction of the Osbourn spreading centre (Fig. 1) most likely caused by azimuthal mantle anisotropy (Kodaira et al. 2014).

Typical bend-related faulting along the Tonga-Kermadec subduction zone is characterised by up to 2 km throw horst-and-graben structures that coincide with a seabed dip of 5°, as shown by Profile D (Fig. 10b – Funnell et al. 2014). In the proximity of the present-day LRSC-trench collision zone, the Tonga-Kermadec trench shallows by \(-4\) km over 150 km (Pontoise et al. 1986; Funnell et al. 2014), and is coincident with reduced bend-related deformation, as represented by faults observed along Profile B. Faults in this region predominantly form half graben that dip trenchward and exhibit a \(<0.3\) km throw, resulting in the subducting plate remaining sub-horizontal with \(<1°\) dip (Fig. 10a). This region of decreased deformation extends up to 120 km from the peak of the closest seamount (Fig. 9a), which is significantly broader than the \(-50\) km-wide bathymetric swell and lower-crustal support imaged (Fig. 6b; Ballance et al. 1989; Contreras-Reyes et al. 2010). Although some larger faults
are found on the flanks of Osbourn seamount (Stratford et al. 2015; Robinson et al. in review), the general reduction in subducting plate dip and fault offset suggests that the presence of the LRSC and its corresponding crustal thickening moderates deformation by providing structural and buoyancy-driven support. This also appears to limit the reduction in subducting plate flexural rigidity in the 120 km around the LRSC, resulting in seamounts that effectively subduct on top of a broad raised ridge that would also contribute to deformation of the overriding plate (e.g. Dominguez et al. 1998; Zeumann & Hampel 2015).

6.2 Characterising the along-strike variability of the Tonga-Kermadec forearc

The 12-13 km thick forearc-arc crust imaged along Profile B is thin compared to global observations of intra-oceanic overriding crust (Fig. 11a & b). Fig. 12 indicates that the thin crust may not be an isolated feature of the Tonga-Kermadec forearc but part of a trend of northward increasing crustal thickness that reaches >18 km over just 300 km distance along strike (Contreras-Reyes et al. 2011). This ~6 km increase is coincident with similar increases in the thickness and width of the high velocity region at the base of the forearc crust, as well as shallowing of the forearc bathymetry. Over the same area, the signature of the FAA (Sandwell et al. 2014) increases in amplitude from ~100 mGal around Profile B to >150 mGal at P02 (Fig. 12c), suggesting that one of the dominant controls on the gravity anomaly along the Tonga-Kermadec forearc is the difference in density between lower crustal material associated with the currently active and extinct frontal arc (Fig. 8).

The general trend of northward-increasing crustal thickness, the FAA, and the prominence of the extinct Eocene arc appears to continue north and south of the region imaged by Profiles B and A, and profile P02 (Fig. 12a & e). To the north, the Eocene frontal arc shoals significantly to form the Tongan island group (Contreras-Reyes et al. 2011), and to the south the bathymetric expression disappears beneath the sedimentary basin of the Kermadec forearc whilst the FAA is significantly reduced (Funnell et al. 2014). Where sampled, the morphology of the forearc crust indicates that its broad structure predates the majority of sedimentation at the margin (Fig. 2; Herzer & Exon 1985). These observations suggest that the margin-wide trend in forearc crustal thickness is controlled by systematic variations in crustal material generation or erosion following formation, but prior to significant sedimentation and seamount subduction. Such observations fit with the proposed tectonic model of Meffre et al. (2012), which infers the southward propagation of subduction initiation, potentially enabling more volcanic material to be generated in the north where subduction has been sustained for a longer period of time.

6.2.1 Southern extent of the extinct Eocene frontal arc

Although the bathymetric expression of the Eocene arc is absent from the Kermadec forearc 50 km south of Profile B, the associated FAA high continues south at a lower amplitude as far as ~31.5°S (Fig. 12e). Using detrended residual FAA data Bassett et al. (2016) show that this termination in gravity signature coincides with the ‘32°S’ tectonic boundary (Ballance et al. 1999). This boundary separates a deep sediment-filled basin, a prominent arc, and a typical <3 km-deep backarc basin to the north from a >10 km-deep trench axis and the westward migration of arc volcanism into a deeper, narrower, and more asymmetric Havre Trough (Pelletier & Dupont 1990; Collot & Davy 1998; Ballance et al. 1999). Whilst this was initially proposed to result from varying...
rates of subduction erosion around the boundary (Pelletier & Dupont 1990; Collot & Davy 1998), more recent research has highlighted the need for a southern limit on the extinct Eocene arc to explain its migration during the opening of the south Fiji Basin (Herzer et al. 2012). Based on the absent bathymetric expression and southern termination of the FAA high, we support the interpretation of Bassett et al. (2016) that the ‘32°S’ boundary represents the present-day southern limit of the extinct Eocene arc along the Tonga-Kermadec forearc (Fig. 12g).

6.2.2 Northern extent of the extinct Eocene frontal arc

North of Profile P02, the FAA widens and increases in amplitude towards the Tongan islands (Fig. 12e). Just beyond the northern tip of these islands, at ~18°S, there is an abrupt reduction in the FAA, shown as a highly negative vertical gravity gradient (Fig. 12f), which also coincides with an ~2 km-deep bathymetric channel and the broadening and deepening of the forearc and arc (Fig. 12a). This sharp delineation coincides with a decrease in age of dredged rock samples, from 51-39 Ma to 35-28 Ma (Meffre et al. 2012), an aseismic discontinuity (Bonnardot et al. 2007), and the southern tip of the Fonualei Rift (Zellmer & Taylor 2001; Conder & Wiens 2011). Bonnardot et al. (2007) proposed that the initial collision of the LRSC, ~5-4 Ma (Ruellan et al. 2003), may have generated this feature, known as the Fonualei Discontinuity (Fig. 1). Although seamount subduction is associated with increased subduction erosion and the removal of significant quantities of material from the overriding plate (Dominguez et al. 1998; von Huene et al. 2004; Clift & MacLeod 2012), the Fonualei Discontinuity is sharper and extends further arcward than is typically expected for such deformation. The seamount subduction hypothesis also fails to explain the age variation across the boundary. We propose that the Fonualei Discontinuity divides the northern extent of the extinct Eocene frontal arc from a younger volcanic frontal arc to the north (Fig. 12g). The significant broadening and reduction in forearc depth north of the discontinuity compared to the south, by ~1 km (Fig. 12a), may be expected of volcanic arcs at different stages of their evolution (Takahashi et al. 2007; Calvert et al. 2008; Kodaira et al. 2010). Extension along the Fonualei Rift in the Lau Backarc Basin terminates just north of the forearc-dividing discontinuity (Zellmer & Taylor 2001, Conder & Wiens 2011), suggesting that changes in crustal structure not only control forearc deformation, but also backarc stresses and processes as the host structure formed prior to Lau Basin spreading (Pelletier & Louat 1989). The observed lack of seismicity along this discontinuity may be caused by the ability of the forearc crust to accommodate significant stresses due to inherited weaknesses along this boundary.

6.3 Implications for seamount subduction along the Tonga-Kermadec trench

Subduction of bathymetrically prominent ridges and seamounts is commonly associated with increased rates of crustal erosion (Ballance et al. 1989; Dominguez et al. 1998; Clift & Vannucchi 2004; Zeumann & Hampel 2015). Although early seamount subduction models suggested that flexure of the overriding crust accommodated seamount subduction over a broad region of the forearc (e.g. Scholz & Small 1997), more recent studies find that seamounts deform the overriding forearc in a series of ring and radial faults proximal to the interplate boundary (Dominguez et al. 1998; Wang & Bilek 2011).
The ability to determine the exact location of currently subducting seamounts along the Tonga-Kermadec forearc is limited by the morphological complexity of the lower-trench slope (Fig. 9a), and the quality and quantity of data acquired in the region. We find, with the acquisition geometry of Profile B, that the seismic velocity model is only sensitive to changes of >0.5 km s\(^{-1}\) in the lower-trench slope region (Fig. 6; Table 2), and that only features larger than 30 km by 3 km may be reliably resolved through inversion modelling (Fig. 7). Such sensitivities are comparable to the perturbation expected by the presence of an intact subducting seamount (Lonsdale 1988; Contreras-Reyes et al. 2010), suggesting current WA datasets and associated modelling techniques may be unable to discriminate between the presence and absence of seamounts in the lower-trench slope of the Tonga-Kermadec subduction system.

Despite the current limitations in locating the subducting LRSC beneath the lower-trench slope, the broad-scale and historical influence of seamount subduction on Tonga-Kermadec forearc structure can be more clearly determined. The lower-trench slope crust exhibits low seismic velocities (3-6 km s\(^{-1}\)) along the subduction system (Fig. 11c; Crawford et al. 2003), suggesting that the region comprises crustal material that has been highly fractured by erosive subduction processes (e.g. von Huene & Ranero 2003; Sallarès & Ranero 2005), although the poor resolution here precludes further analysis. Along Profile B, the ~25 km-wide and relatively flat-topped mid-slope basin is underlain by typical forearc velocities (4-7 km s\(^{-1}\)). Similar basins are observed to the south (e.g. Ballance et al. 1999; Funnell et al. 2014), and north of ~22°S (Wright et al. 2000), but are significantly reduced in the vicinity of recent LRSC subduction (Contreras-Reyes et al. 2011; Stratford et al. 2015). Coincident with this mid-slope basin deformation is a change in the distance between the trench and onset of the Tonga Ridge, or upper-trench slope, which increases from >80 km-trench offset along Profile B to <70 km along Profile A and profile P02 (Fig. 12). Together, these observations support the hypothesis that subduction of the LRSC increases the basal and frontal erosion of the lower-trench slope crustal material (Ballance et al. 1989; Contreras-Reyes et al. 2011; Stratford et al. 2015). The data also suggest that the mid-slope crust and basins, which are typically preserved by steady-state subduction processes (Wright et al. 2000), are deformed and removed by stresses either directly as seamounts subduct (Wang & Bilek 2011) or as the inner trench slope re-equilibrates gravitationally through extension and collapse (Clift & Vannucchi 2004; Funnell et al. 2014).

These erosional and structural recovery processes that define the subduction of seamounts can thus be used, in the context of the characterised variations in forearc structure (Fig. 12g), to track the effects of LRSC-subduction along the Tonga-Kermadec subduction system. Immediately north of the present-day LRSC-trench intersection point, the Tonga trench reaches its deepest point (10.8 km; Ballance et al. 1989), and the trench slopes step westward by up to 80 km (Contreras-Reyes et al. 2011). The trench slopes appear to return to a more typical structure around ~23°S (Stratford et al. 2015), although the bathymetric ridge and high FAA representative of the extinct Tonga arc exhibits a significant reduction in depth (0.5 km) and amplitude (50 mGal), whilst the trenchward edge of the ridge steps ~20 km westward, at 22°S (Fig. 12). At 22°S the Tonga forearc also displays parallel normal faults that delimit the northern extent of the region of significant faulting and doming observed by Herzer & Exon (1985), which they attribute to extensional stresses active from ~5-1 Ma based on well-tied
seismic data. This suggests that although the Tonga forearc crust thins gradually to the south along the margin (Fig. 12), the pronounced step in forearc structure at this latitude formed as a direct result of seamount subduction-related erosion since ~5 Ma (Ruellan et al. 2003; Timm et al. 2013), and the subsequent extensional collapse of the forearc to re-establish trench-slope stability. The apparent lack of pervasive deformation north of 22°S (Herzer & Exon 1985; Fig. 12) supports the proposal of Ballance et al. (1989) that LRSC-subduction initiated near that latitude and has subsequently migrated southward. The presence of a more westerly trend of the subducted segment of the LRSC than is bathymetrically expressed is thus required to enable sufficient time to pass for the migration, from the onset of subduction at 22°S ~5-4 Ma to the present-day collision zone (Ballance et al. 1989; Ruellan et al. 2003). Despite the apparent documentation of seamount subduction at ODP site 841 ~1.5 Ma (MacLeod 1994; Clift & MacLeod 1999), there is significant evidence that LRSC seamounts deform a wider region of the complex and faulted lower-trench slopes than expected by their bathymetric expression. This result makes determining the history of seamount subduction more challenging and only with further drill sites along the forearc for comparison would the exact history of seamount subduction at the Tonga-Kermadec margin be reliably determinable.

7 CONCLUSIONS

New WA seismic and density models are synthesised with previously published profiles and satellite-derived gravity maps to highlight along-trench structural variability in the Tonga-Kermadec subducting plate and overriding forearc. We use these observations to characterise inherited forearc structure and determine the spatial and temporal extent of LRSC subduction-related deformation of the overriding plate. We find that:

1. Sampling of the flanks of the LRSC-seamounts fails to image high internal velocities, possibly indicating a laterally limited intrusive mafic core. High seismic velocities, >7.2 km s⁻¹, observed at the base of the crust over 50 km from Canopus seamount suggests their structural support extends over a wide region.

2. Subducting oceanic crust and mantle, which is typically faulted and hydrated as it bends towards the trench resulting in rapid structural weakening, deforms less in the vicinity of the LRSC. Typical bending of the subducting plate generates full graben with up to 2 km vertical throw and a plate dip of <5°, whereas within the vicinity of the LRSC swell, the near-horizontal subducting plate is offset by <200 m dominantly trenchward-dipping normal faults. Shoaling of the trench apex by up to 4 km at the LRSC-trench intersection shows that, despite the limited bathymetric expression of the seamounts, the chain sits atop an ~250 km-wide ridge.

3. The Kermadec forearc comprises lower- and mid-trench slopes that persistently undergo extension due to frontal and basal erosion as well as associated gravitational collapse. Crustal seismic velocities (3.5-6.0 km s⁻¹) are unusually low in this region. Modelling of more typical densities for oceanic crust suggests that they are highly fractured, faulted, and possibly hydrated.

4. The upper forearc slope is dominated by a basin and ridge structure that predates significant sedimentation at the margin. This bathymetric ridge is underlain by high velocities, 7.0-7.4 km s⁻¹, that are best represented by
high density lower-crust (3.2 g cm⁻³), that most likely formed during subduction initiation along the Tonga-
Kermadec margin (~51 Ma).

5. The nature and extent of this Eocene arc can be constrained using the broad-scale high FAA that corresponds
to the high density lower crustal anomaly. The extinct arc extends from 32°S to the Fonualei Discontinuity at
18°S, with the overriding crustal thickness increasing to the north.

6. Subduction of the LRSC along the Tonga-Kermadec trench increases the erosion of material from the
overriding crust. The lower- and mid-trench slopes are shortened, and the trenchward edge of the Eocene arc
most likely undergoes extensional collapse to re-equilibrate trench-slope stability.

7. Although the trench slopes are heavily deformed and eroded as far north as 23°S, extensional faulting of the
Tonga forearc crust is observed to 22°S. LRSC-subduction is thus proposed to have begun at 22°S, ~5 Myr
ago, although the seamount-associated swell may have enabled a broader footprint of deformation.

ACKNOWLEDGEMENTS

This research project was funded by the Natural Environmental Research Council (NERC) grant
NE/F004273/1 (Peirce & Watts 2011). We are grateful to the two anonymous reviewers for their thoughtful and
helpful comments on this paper. We would like to thank all involved in the planning and acquisition of data during
research cruise S0215, including the officers and crew of the R/V Sonne, the scientific party, and all seagoing
technicians. The NERC Ocean-Bottom Instrumentation Facility (Minshull et al. 2005) and GEOMAR provided
the OBSs used in this project, together with their technical support at sea. The MCS data were processed using
Claritas™, and manipulated for plotting using Seismic Unix. All figures were prepared using the Generic Mapping
Tools (GMT) package (Wessel & Smith 1998). All data from cruise SO215 are archived at the NERC’s British
Oceanographic Data Centre, and the final accepted version of this manuscript is available through Durham
Research Online (dro.dur.ac.uk).
REFERENCES


Sallarès, V. & Ranero C.R., 2005. Structure and tectonics of the erosional convergent margin off Antofagasta,
Tonga-Kermadec forearc structural variability


Table 1. Summary of picked and modelled phases. The final model compares over 45,000 modelled arrivals to observed phases, with a combined $\chi^2 < 2$. Phases through different model sections, as well as the final model as a whole, are listed separately to highlight the disparity in data quality and modelling capabilities along the profile.
Table 2. Summary of results from independently testing the sensitivity of different final forward model sections.

Control tests, in which rays from a subset of instruments were traced through the unedited respective model sections, indicated the contribution of each set to the overall model fit. Crustal layer sensitivities are presented in this table as the average of the three layers, rather than as individual values, for brevity and thus may vary from values quoted in the text. ΔZ and V_p signify change in depth and P-wave seismic velocity, respectively. Note that layer thicknesses were not tested in the trench section because of model complexity.
Fig. 1. Regional bathymetric map of the Tonga-Kermadec subduction system and the northwestern end of the LRSC (IOC et al. 2003). All major tectonic features are labelled, and black arrows highlight the extent of the bathymetrically expressed extinct Tonga arc. FD, LRSC, OSC, and HDB (green circle) represent the Fonualei Discontinuity, Louisville Ridge Seamount Chain, Osbourn spreading centre, and Horizon Deep Bight respectively. Labelled black and red lines represent existing geophysical profiles and Profile B (PB), the focus of this study, respectively. White circles indicate locations of numbered ODP and DSDP drill sites (Burns & Andrews 1973; Clift et al. 1994).
FIGURE 2

CONTINUED ON NEXT PAGE
FIGURE 2 CONTINUED

Fig. 2. Overview of geophysical data acquired along Profile B. a) Swath bathymetry map with LC- (purple) and KUM-type (orange) ocean-bottom seismographs marked on Profile B (black line). b) Fully processed MCS section with major region classifications above, and the interpreted base-sediment reflector shown by the red line. c) First-pass velocity model-stacked MCS data over the Kermadec forearc ridge. d) Final WA velocity model-restacked data over the same region as c). Dark blue dots indicate the forward WA model sediment-crust interface. e) First-pass velocity model stack. f) Final WA velocity model-restacked MCS data highlighting the improved clarity of the Moho reflector and its fit with the Moho of the forward velocity model (light blue dots).
Fig. 3. Kermadec forearc. a) Hydrophone data from OBS 02 plotted with a reduction velocity of 8 km s⁻¹, with major phases labelled. b) Picked phases (yellow – Pₙ; orange – Pₕ; green – Pₘ; brown – Pₚ). Colour bar height indicates pick uncertainty, and black lines represent modelled arrival traveltimes. c) Calculated ray-paths for each picked phase through the final forward model. The location of OBS 02 is indicated by the red inverted triangle.
Fig. 4. Kermadec trench. a) Hydrophone data from OBS 12 plotted with a reduction velocity of 8 km s\(^{-1}\), with major phases labelled. b) Picked phases (yellow – \(P_n\); orange – \(P_g\); green – \(P_{mP}\); brown – \(P_n\)). Colour bar height indicates pick uncertainty, and black lines represent modelled arrival travel times. c) Calculated ray-paths for each picked phase through the final forward model. The location of OBS 12 is indicated by the red inverted triangle.
Fig. 5. Pacific plate a) Hydrophone data from OBS 19 plotted with a reduction velocity of 8 km s$^{-1}$, with major phases labelled. b) Picked phases (yellow – $P_n$; orange – $P_g$; green – $P_{nP}$; brown – $P_d$). Colour bar height indicates pick uncertainty, and black lines represent modelled arrival traveltimes. c) Calculated ray-paths for each picked phase through the final forward model. The location of OBS 19 is indicated by the red inverted triangle.
Fig. 6. a) Location of OBSs along Profile B. b) Final forward velocity model for Profile B, with layer boundaries marked in black. c) Traced-ray density through the forward model space. Model misfits are indicated in Table 1, and the sensitivity of the model fit to boundary perturbations is highlighted in Table 2.
Fig. 7. a) Initial inversion model with velocities increasing to 8 km s\(^{-1}\) at 15 km-depth. Velocity contours for a-d are annotated in km s\(^{-1}\). b) Result from the first run of five non-linear inversions on the initial model, using a 10
km by 1 km inversion cell size. c) Final result, following a further five non-linear inversions of the result of the first run, using a 5 km by 0.5 km inversion cell size. The box in the bottom-right hand corner of b) and c) indicates inversion cell sizes for each model space. d) Final forward WA velocity model from Fig. 6, with the same velocities contoured for comparison with parts a-c. e) Final velocity model inversion ray density. f) 30 km by 3 km checkerboard test result, showing good anomaly recovery to ~16 km depth and within 20 km of the model edges. The black grid lines indicate the input checkerboard dimensions. g) 20 km by 2 km checkerboard test result, indicating good recoverability is limited to within the subducting plate and between 50-150 km offset in the overriding crust.
Fig. 8. a) Observed satellite- and shipboard-derived free air anomalies (FAA) compared to the calculated anomalies from different density models. b) Layered block model, with lower-crustal high-density structure. Values denote assigned block density (expressed in g cm$^{-3}$ for brevity and clarity of annotation). c) Combined geophysical model and structural interpretation along Profile B. The green line in parts b) and c) depicts the slab1.0 model projected along Profile B (Hayes et al. 2012).
Fig. 9 a) Bathymetric map of the Pacific plate in the vicinity of the LRSC-Tonga trench collision zone. WA profiles, and the locations of the 1D seismic velocity profiles, presented in b), are indicated by lines and circles, respectively. Profile names are annotated (Profile A is PA, etc.) b) Velocity profiles through the crust and mantle. The grey-shaded region represents average seismic velocities for mature Pacific oceanic crust (>25 Ma) from White et al. (1992).
Fig. 10. a) and b) MCS data from the trench-proximal Pacific crustal regions of Profiles B and D respectively (Fig. 9a), highlighting the varying nature of normal faulting along the margin. Green dots on the Profile B MCS data indicate the TWTT of the final forward velocity model crust-mantle layer boundary.
Fig. 11. a), b) & c) 1D velocity profiles through the crust and upper mantle of the active arc, frontal inactive arc, and lower trench slopes respectively. 1D profiles presented are, S-15, from Stratford et al. (2015), CR-11, from Contreras-Reyes et al. (2011), C-03, from Crawford et al. (2003), and H-99, from Holbrook et al. (1999). The Izu-Bonin-Mariana range and average 1D-seismic velocity profiles are derived from a compilation of published results from across the margin, including Takahashi et al. (2007), Calvert et al. (2008), Takahashi et al. (2009), Kodaira et al. (2010).
FIGURE 12

Tonga-Kermadec forearc structural variability

FIGURE 12
Fig. 12. Maps and WA seismic velocity models highlighting variable forearc structure along the Tonga-Kermadec subduction system. a) Regional bathymetry map, with relevant WA profiles labelled. b), c), and d) Trench-oriented and perpendicular distance offset (arcwards negative) WA seismic profiles presented at the same scale for comparison. Profile P02 (P02) data are from Contreras-Reyes et al. (2011), Profile A (PA) from Stratford et al. (2015), and Profile B (PB) from this study. Forearc crust generally thickens from Profile B to P02. e) Regional FAA. f) Vertical gravity gradient (VGG), which highlights the extent of the extinct Tonga Ridge beneath the forearc between 18 and 32°S as a highly positive gravity gradient. g) Summary of along-arc changes in forearc and arc structure, highlighting major tectonic boundaries and regions of LRSC deformation. Annotated features are: the Central and Eastern Lau Spreading Centres (CLSC and ELSC, respectively), the Fonualei Rift (FR), and the Fonualei Discontinuity (FD).