Gravity anomalies, crustal structure, and seismicity at subduction zones: 2. Interrelationships between fore-arc structure and seismogenic behavior

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Abstract
An ensemble-averaging technique is used to remove the long-wavelength topography and gravity field associated with subduction zones. Short-wavelength residual anomalies are attributed to the tectonic structure of subducting and overthrusting plates. A paired (positive-negative) fore-arc anomaly is observed consisting of a long (>1000 km), linear, trench-parallel ridge landward of the deep-sea-terrace basin. Ridges have amplitudes of 1500–3000 m and 160–240 mGal, wavelengths of 150–200 km, and high gravity anomaly to topography ratios (50–75 mGal km⁻¹). The ridge crests correlate with the downdip limit of coseismic slip and strong interplate coupling and in Cascadia, the updp limit of tremor epicenters. The ridge crest may be interpreted as defining the boundary between the velocity-weakening and seismogenic region of the subduction interface and the downdip frictional transition zone. In Tonga-Kermadec, the Kuril Islands and Chile landward ridges are associated with extinct volcanic arcs. Paired anomalies are attributed to the preferential subduction erosion of the outer fore arc and a spatially varying combination of (a) lower crustal underplating beneath the inner fore arc, (b) the transformation of interseismic strain into permanent geologic strain via faulting, folding, or buckling of the inner fore arc, and (c) the relative trenchward migration of extinct volcanic arcs in regions operating with a net crustal deficit. Along-strike transitions in fore-arc morphology and seismogenic behavior are related to preexisting crustal structure of subducting and overthrusting plates. Fore arcs have the added potential of recording the time-integrated response of the upper plate to subduction processes, and fore-arc structure should be considered in tandem with seismological observations.

1. Introduction

The largest earthquakes on Earth take place on the megathrusts of subduction zones where the shallow trajectory of subducting plates and the depression of isotherms result in wide zones over which seismic processes can occur [Lay et al., 2011; Simons et al., 2011]. The fault slip behavior of megathrusts is variable and although paleoseismological, historical, and instrumental evidence clearly demonstrates the potential of some margins to rupture in great earthquakes, in many other regions such evidence is lacking. It is difficult to know whether these observations reflect the physical inability of some active margins to generate great or giant earthquakes, or simply the short duration of historical and instrumental records with respect to the multicentury recurrence intervals associated with the largest earthquakes [McCaffrey, 2008]. In this study, we consider other lines of evidence that may be diagnostic of seismogenic behavior.

Intermargin variability is further compounded by intramargin along-strike segmentation and early work in the 1970s showed that many seismic gaps and quiet zones within regionally active margins may be modulated by the tectonic structure of subducting plates [Sykes, 1971; Kelleher et al., 1973; Kelleher et al., 1974; Kelleher and McCann, 1976; Spence, 1977]. In the following decades, a growing body of evidence has documented seamounts [Kodaira et al., 2002; Mochizuki et al., 2008], aseismic ridges [Spence et al., 1999; Bilek, 2010; Sparkes et al., 2010], and fracture zones [Spence, 1977; Barrientos and Ward, 1990; Schoffel and Das, 1999; Robinson et al., 2006] impeding, arresting, or resulting in complex coseismic rupture [Bilek et al., 2003; Wang and Bilek, 2011], and geodetic observations suggest creeping is the predominant mode of subduction in regions of rough seafloor [Wang and Bilek, 2014]. But how does the structure of the overthrusting plate influence and/or reflect seismogenic behavior in subduction zones?
One of the key difficulties in considering this question is identifying short-wavelength structure within the high topographic and gravimetric gradients that characterize subduction fore arcs. Where the Louisville ridge seamount chain subducts beneath the Tonga-Kermadec trench, spectral-averaging techniques designed to isolate and remove the average trench-normal topographic profile (supporting information Figure 1) have been shown to be an effective means of suppressing these gradients, and here enable the bathymetric expression of the subducting ridge to be identified up to 60 km landward of the trench axis [Bassett, 2014]. This study extends spectral averaging to regional data sets of bathymetry and free-air gravity anomalies for all subduction zones on Earth (Figure 1), the results of which are presented in two parts. Part 1 [Bassett and Watts, 2015] considers the structure of subducting plates and is focused on resolving the bathymetric expression of subducting topographic relief. Global constraints are placed on where large bathymetric anomalies may be present within seismogenic zones, enabling relationships with seismicity and mechanical models of relief subduction to be reconsidered. Part 2 (this paper) is focused on overthrusting plate structure and interrelationships with the seismogenic behavior of the underlying subduction interface. In contrast to part 1, where a clear direction of causality exists between subducting plate structure and slip behavior, it is known that the physical properties of the subduction interface strongly influence the structure of the overthrusting wedge [Davis et al., 1983; Dahlen et al., 1984; Wang and Hu, 2006]. We thus follow earlier studies [Ruff and Tichelaar, 1996; Song and Simons, 2003; Wells et al., 2003] in placing particular emphasis on considering the degree to which fore-arc structure, as revealed by residual bathymetry and residual free-air gravity anomalies, may reflect the mechanical and seismogenic behavior of the subduction interface.

2. The Dimensions of Seismogenic Zones

Constraining the dimensions and segmentation of the strongly coupled, velocity-weakening, and seismogenic region of the subduction interface is key to understanding both seismic hazard and the underlying physical controls on megathrust slip behavior (Figure 2). Up and downdip transitions to regions of stable
(aseismic) sliding are often, in relatively warm subduction zones [Hyndman and Peacock, 2003], considered to be thermally modulated and associated with the dehydration of smectite clays to illite and chlorite at temperatures of 100–150°C [Wang, 1980], and the transition from friction to intracrystalline plasticity as the dominant deformation mechanism in quartzofeldspathic rocks at temperatures of 325–350°C [Hyndman and Wang, 1993, 1995; Hyndman et al., 1997; Oleskevich et al., 1999]. In relatively cold subduction zones, this latter isotherm may be depressed to a depth below the intersection between the subducting slab and fore-arc Moho and it has been suggested that the seismogenic zone may be limited by the juxtaposition of stable sliding serpentinites against the subduction interface [Hyndman and Peacock, 2003]. A similar model has been employed by Ruff and Tichelaar [1996] to explain a broad correlation between coastlines and the downdip limit of seismogenic zones, but in Sumatra distributions of coseismic slip [Dessa et al., 2009; Klingelhofer et al., 2010; Collings et al., 2012] and observations of interseismic deformation [Simoes et al., 2004] are not consistent with the seismogenic zone being limited by the fore-arc Moho.

Direct estimation of the location of these transitions, which are important for tsunami generation and the proximity of strong ground shaking to densely populated coastal regions, is challenging. Marine measurements of heat flow are sparse, difficult to make, associated with large uncertainties, and hydrothermal fluid circulation has a large effect on the results of thermal models [Hyndman and Wang, 1993, 1995; Currie et al., 2002; Cozzens and Spinelli, 2012]. Geodetic observations of surface deformation improve the resolution with which fault locking can be estimated [Wallace et al., 2004; Chlieh et al., 2008; Chlieh et al., 2011; Metois et al., 2012]; however, the majority of these observations are made onshore, often at large distances from the locked region of the fault and resolution near the trench is typically poor. Geodetic observations also typically represent only a small fraction (tens of years) of an interseismic great earthquake cycle (hundreds of years) and viscoelastic relaxation causes the pattern of deformation observed from surface motions to evolve during the interseismic period [Wang et al., 2003; Wang et al., 2012]. Seismological and geodetic observations of stick slip, slow slip, or fault creep offer the best contemporary constraints on the dimensions of the seismogenic zone and we make use of these constraints when considering interrelationships between fore-arc structure and megathrust slip behavior.

As demonstrated by the Sumatran earthquakes of December 2004 (Mw 9.1–9.3) and March 2005 (Mw 8.6), subduction megathrusts are also highly segmented along-strike and structural and geometrical heterogeneities can act as barriers to along-strike rupture propagation. The degree of segmentation can limit the area affected by strong ground motions, the amplitude and length scale of associated tsunami, and the amount of displacement, which is often proportional to rupture length [Wells and Coppersmith, 1994]. Along-strike
transitions in slip behavior, however, are less likely to be thermally modulated or even to have a common causality. Some are clearly related to subducting plate morphology [Sykes, 1971; Kelleher et al., 1973; Kelleher et al., 1974; Kelleher and McCann, 1976], but others are not and occur in regions of comparatively smooth seafloor. What other physical, geometrical, and structural properties change along strike; where, why, and how could these explain or reflect differences in slip style?

In this study, we investigate the working hypothesis that trench-parallel fore-arc ridges (TPFRs) may provide insights into the dimensions, seismogenic behavior, and segmentation of subduction thrust faults. In many regions, these ridges are associated with constructional features associated with extinct volcanic arcs or thick accumulations of accreted materials (Sumatra), but it is also possible some reflect, in part: the accommodation of small fractions of interseismic elastic strain via permanent deformation of the overthrusting plate over many earthquake cycles; or surface uplift driven by lower crustal underplating. We suggest TPFRs are important for three primary reasons:

1. Contrasts in the presence or absence of TPFRs between adjacent margin segments constrain first-order margin segmentation.

2. Where present, TPFRs provide coherent laterally contiguous features enabling the identification of structural variations along strike, which may reflect second-order margin segmentation.

3. The contrasting forms of TPFRs across fore arcs allow the geological processes influencing their formation and/or preservation to be considered.

We interpret clear interrelationships between the location and character of TPFRs and seismogenic behavior. The objective of this manuscript is to demonstrate these relationships, establish the direction of causality where possible, and consider the potential significance of TPFRs in evaluating the long-term seismogenic behavior of subduction margins.

3. Summary of Methodology

3.1. Construction of Residual Free-Air Gravity and Topography Grids

In this study we analyze the 1° × 1° min satellite-derived free-air gravity anomaly grid (Version 23.1) of Sandwell et al. [2014] and the GEBCO 1° × 1° min shipboard bathymetry grid (Version 2.0) [IOC, 2003]. The 1° × 1° min GEBCO grid does not incorporate satellite-derived bathymetry data and its use ensures the data sets we analyze are independent.

The gravity and topography grids were sampled along trench-normal profiles, which were analyzed using spectral-averaging techniques designed to isolate and remove the average trench-normal topographic and gravimetric profile in subduction zones. This procedure reduces the dynamic range, removes the steep gradients associated with the seaward and landward trench-slopes, and significantly increases the resolution of short-wavelength structure. Our processing methodology is illustrated Figure 1 of supporting information. For each subduction margin, the average trench-normal profile is calculated by sampling regional grids of bathymetry and free-air gravity anomaly along trench-normal profiles extending >600 km up and down-dip of trench axis and spaced ~25 km along strike. The trench axis is defined as the deepest point using the GEBCO2008 bathymetric grid [IOC, 2003]. A fast Fourier transform is used to calculate the frequency spectrum of each profile. The average frequency spectrum is then calculated from the Fourier coefficients at each frequency before an inverse transform is applied to calculate the ensemble average trench-normal profile for each data set and margin. The linearity of the Fourier transform and its inverse means spectral averages should be identical to arithmetic or stacked averages; however, small differences resulting from finite precision of the Fourier transform and associated downsampling are present for small ensembles (<30 profiles) (supporting information Figure 3). Maintaining the geometry of the trench, grids of each average profile are created and subtracted from the original data sets to produce grids of residual bathymetry and residual free-air gravity anomaly. As opposed to methodologies involving the regridding of residuals calculated along profiles, the calculation and subtraction of an ensemble average grid preserves the full (1° × 1° min) resolution of the original data sets (supporting information Figure 4). This analysis is applied to all subduction zones on Earth and the grids generated are made available as an electronic supplement and online for future use [Bassett and Watts, 2015].
It is important to note that the absolute values of the residual anomalies simply reflect the degree of structural heterogeneity within the margin segments analyzed, are strongly influenced by its along-strike extent, and have no physical meaning. In the Alaska-Aleutian margin, for example, residual grid values in the Atka fore-arc basin differ by up to 50 mGal and 1 km depending on whether average profiles and residuals are calculated across the whole margin, or individually for the Alaska and Aleutian segments (supporting information Figure S). The short-wavelength structure of residual grids, however, does not change, which is why the focus of this study is on interpreting the detailed fore-arc structure revealed as opposed to absolute anomaly values.

3.2. Seismicity

Structural observations from residual gravity and topography anomalies are compared with seismological and geodetic inferences of megathrust slip behavior. This analysis is conducted using a catalog we have compiled containing the epicenters, seismic asperities, and slip areas (either aftershock defined or from distributed slip modeling) for over 150 of the largest (Mw 7–9.5) subduction zone earthquakes (margins with bold labeling in Figure 1). Many asperities are defined from seismic source time functions [Beck and Ruff, 1987; Schwartz and Ruff, 1987; Beck and Ruff, 1989], but where distributed slip models are available we follow the asperity definition of Yamanaka and Kikuchi [2004] in selecting region with slip ≥ half the maximum observed slip in an earthquake. We also analyze the Global Centroid Moment Tensor (GCMT) [Dziewonski et al., 1981; Ekström et al., 2012], Engdahl, van der Hilst, and Buland (EHB) [Engdahl et al., 1998], and reviewed International Seismological Center (ISC) (http://www.isc.ac.uk/, accessed January 2015) earthquake bulletins; and a catalog of seismic tremor occurring in Cascadia since August 2009 [Wech and Creager, 2008; Wech, 2010]. These data sets are integrated with geodetic inferences of plate boundary locking [Chlieh et al., 2008; Chlieh et al., 2011; Metois et al., 2012].

We are interested in earthquakes occurring on the megathrust and extract events of all magnitudes from the GCMT catalog (period 1976 to July 2013) with thrust faulting mechanisms (T axis plunge ≥45°, P axis plunge ≤45°), centroid depth ≤65 km [Heuret et al., 2011; Hayes et al., 2012], and with strike and dip within 20° and 15°, respectively, of the subducting SLAB1.0 geometry [Hayes et al., 2012]. The full duration (1960–2009) and magnitude range (mb ≥ 3) of the EHB catalog is downloaded and reported event times are used to relocate GCMT centroids to their EHB hypocenters. The GCMT catalog is also used to reassess proposed links between regions of negative (≤−40 mGal) trench-parallel residual free-air gravity anomalies and regions of high seismic-moment release [Song and Simons, 2003].

4. Fore-Arc Structure: Trench-Parallel Fore-Arc Ridges

Trench-Parallel Fore-arc Ridges (TPFRs) are defined as long, linear, fore-arc ridges contiguous over distances of at least 500 km of strike. Their preservation and detection within residual free-air gravity and topography grids requires that they are not universally present within the trench segment analyzed or that they change in geometry and/or position along strike. TPFRs are most prominent in Tonga, Kuril Islands/Kamchatka, Chile, Aleutian Islands, Cascadia, Mexico, and Sumatra. Since 1700 AD, the maximum earthquake magnitude observed in each of these regions is ≥8.0 [Stein and Okal, 2007]. TPFRs are not observed along the Marianas, Izu-Bonin, New Hebrides, South Sandwich, Scotia, Lesser Antilles, and southern Kermadec arcs. Many of these margins are considered either aseismic or associated with a lesser potential of generating great (Mw ≥ 8.0) or giant (Mw ≥ 9.0) subduction thrust earthquakes. The largest earthquake observed along each of these margins since 1700 AD are all ≤8.0, with the majority ≤7.5 [Stein and Okal, 2007]. Possible links between TPFRs and megathrust slip behavior are explored below by describing the structural characteristics of TPFRs observed along a suite of circum-Pacific subduction margins and the relationships shown with seismicity. Many of the maps used to demonstrate these relationships are shown only in their interpreted form; however, uninterpreted maps and all grid files are provided as an electronic data supplement.

4.1. Tonga-Kermadec

The Tonga fore arc is occupied by the Eocene-Oligocene Tonga Ridge. The bathymetric expression of this extinct island arc is ~800 km long (26°S–18.5°S) and located entirely within the Tonga fore arc; however, Geosat gravity data reveal the southern extension of this TPFR within the northern Kermadec fore arc increasing its length to >1400 km [Collot and Davy, 1998].
Figure 3. (a) Residual free-air gravity anomalies for the Tonga and northern Kermadec subduction zone. Light gray, dark gray, and black-dashed lines mark the trench axis, and the perimeter and crest of the Tonga Ridge, respectively. Red triangles show locations of Holocene active volcanoes [Simkin and Siebert, 2002]. Solid lines show the geometry of active source seismic profiles. (b) Residual bathymetry. (c) Engdahl van der Hilst and Buland (EHB) earthquake density and Global Centroid Moment Tensor (GCMT) thrust earthquakes with Mw ≥ 6. (d) Wide-angle forward velocity models traversing the Tonga Ridge [Bassett, 2014]. Dashed gray lines mark the gravimetric Trench Parallel Fore-arc Ridge (TPFR) margins. Profile L1 is part of a longer profile and the model of Stratford et al., (2015) also resolves the high-velocity Tonga ridge in the inner fore-arc. (e) Profiles of residual bathymetry (green) and free-air gravity anomalies (black) across the TPFR. Dashed gray, blue, and red lines mark the trench axis, outer-arc high, and TPFR crest, respectively. Subducting SLAB1.0 geometry is from Hayes et al., [2012]. Earthquakes from the GCMT catalog (all event classes) are projected from EHB hypocenters along strike with the maximum projection distance shown on each plot. For each mechanism, directions of maximum (P) and minimum (T) compressive stress are plotted in black and red, respectively.
In Figure 3a, we show residual free-air gravity anomalies, calculated by removing an ensemble average profile calculated across the southern Kermadec arc. The gravimetric expression of the Tonga Ridge is clearly resolved with amplitude $\geq 50$ mGal and width $110 \pm 20$ km as far as $30.5^\circ$S. In contrast to the margin-parallel northern segment, south of $25^\circ$S the ridge strike is approximately north-south ($\sim 0^\circ$) resulting in a progressive narrowing of the trench-ridge crest distance. This geometry is consistent with seismic velocity constraints on the location of the Tonga Ridge, which is seismically defined by higher $V_p$ (locally up to 0.8 km s$^{-1}$) at intermediate depth (3–8 km below seabed), a shallow westward dip of isovelocity contours beneath the ridge platform and a sharp reduction in $V_p$ across the eastern ridge flank [Contreras-Reyes et al., 2011; Bassett, 2014; Stratford et al., 2015]. Wide-angle forward velocity models L1 and M4 showing the correlation between the seismic and gravimetric expression of the Tonga Ridge are displayed in Figure 3d, with gray-dashed lines marking the gravimetrically defined ridge margins from Figure 3a.

Residual bathymetry is shown in Figure 3b. The gravimetrically defined margins and crest of the TPFR are dashed in black and reveal a contrasting expression of the TPFR across the ridge crest. In contrast to the landward ridge flank, which is relatively smooth along strike and has shallow gradients of 0.5–2°; the seaward margin appears heavily incised, is structurally heterogeneous along strike and has steeper gradients that average 3–5°, and locally exceed 15°. Some incised canyons are up to 50 km long, 30 km wide, and 1.3 km deep. South of $\sim 25^\circ$S, the residual bathymetric expression of the ridge is largely absent.

In Figure 3c, EHB earthquake density is calculated as the number of earthquakes with depth $\leq 65$ km within a 10 km search radius. Where the TPFR is present, the gravimetrically defined ridge crest defines the landward extent of almost all shallow earthquakes, the majority of which likely occur on the subduction interface (Figure 3c). Between 20°S and 25°S, a second correlation is interpreted between the downdip limit of shallow earthquakes and the steepest seaward dipping bathymetric gradients of the TPFR. Thrust earthquakes with $M_w \geq 6$ are scaled for magnitude and colored for depth and where the TPFR is strongly expressed, these events are located entirely seaward of the gravimetric TPFR crest and either on, or trenchward of, steep seaward dipping residual bathymetric and gravimetric gradients. Cross sections further demonstrating the relationship between the gravimetric ridge crest and the downdip limit of thrust earthquakes are shown in Figure 3e, with the ridge crest and trench-axis dashed in black and gray, respectively.

Large residual gravimetric anomalies in the northern Tonga arc are not a clear extension of the TPFR and thrust earthquakes are widely distributed. The Louisville seismic gap is also clearly expressed and is considered analogous to other regions of aseismic seamount subduction [Bassett and Watts, 2015]. South of $\sim 25^\circ$S, thrust earthquakes remain seaward of the gravimetric TPFR crest, except at its southern extremity where both the gravimetric and bathymetric expressions of the Tonga Ridge are further reduced or absent.

### 4.2. Kuril-Kamchatka

The Vityaz Ridge occupies the Kuril Islands fore arc (Figure 4). This ridge was a site of sedimentation and active volcanism throughout the Eocene and early Oligocene, subsidence throughout the Late Oligocene and Neogene, and Pliocene-Pleistocene volcanic rocks associated with transverse faults crossing the Vityaz Ridge are analogous to the tholeitic and calc-alkali volcanic rocks of the Kuril arc [Laverov et al., 2006; Kulinich et al., 2007; Lelikov and Emelyanova, 2011]. This TPFR extends over $> 1000$ km and is expressed in the residual free-air gravity field as a seaward verging triangular prism with an amplitude and half-wavelength of $\sim 200$ mGal and $\sim 100$ km, respectively. The residual bathymetric character of the ridge is asymmetric with the rough and steep (up to $20^\circ$) seaward ridge flank contrasting with the smooth and shallowly dipping (1–3°) morphology of the landward slope. In contrast to the broad correlation observed in Tonga, the sharp transition in the residual bathymetric character of the TPFR is located up to $\sim 50$ km seaward of the gravimetric ridge crest. The TPFR is detached from the volcanic arc by a dextral strike-slip fault that accommodates south-westward fore-arc sliver translation [Kimura, 1986; Wang, 1996].

In the southern Kuril Islands, the TPFR and a narrow ($\sim 40$ km) and lower amplitude ($< 1.5$ km and $\sim 50$ mGal) outer-arc high (blue dash) are truncated by the Bussol Graben (labeled 3 in Figure 4a). This graben marks the western boundary of a $\sim 500$ km wide zone of extensional deformation where NW-SE trending normal faults form an asymmetric, westward tilting graben within fore-arc basement [Kulinich et al., 2007]. The bathymetric
expression of the TPFR is absent north of the Bussol Graben, but residual free-air gravity anomalies and seismic reflection observations [Kulinich et al., 2007] suggest extension at least as far as Onekotan Island.

Extensional deformation is most intense between Urap and Onekotan Islands [Kulinich et al., 2007] and appears spatially correlated with the region of the back arc occupied by the Kuril Basin. Southeast of Onekotan Island, a bathymetric lineation traversing the fore arc (labeled 1 in Figure 4a) appears contiguous with the northeastern Kuril Basin margin. A second lineation (labeled 2) within the zone of extensional deformation is parallel to the northeastern Kuril Basin margin and marks the western boundary of the zone in which the TPFR is bathymetrically absent. We hypothesize that lineations 1 and 2 may represent the original margins of the Kuril Basin, with extensional deformation, particularly focused between lineations 2 and 3, associated with westward fore-arc sliver translation resulting in a progressive widening of the basin between the arc and trench [Kimura, 1986; Wang, 1996]. This hypothesis is consistent with the northeast-southwest gradient in the intensity of extensional deformation, the geometry of faults mapped within the fore arc (dashed red in Figure 4a) [Kulinich et al., 2007] and explains the continuity between the Bussol Graben and the southwestern Kuril Basin margin. This hypothesis is not consistent with existing models of Kuril Basin

**Figure 4.** Kuril/Kamchatka subduction zone. Symbols as in Figure 3 or as described in text. (a) Residual bathymetry. Yellow stars mark large (Mw ≥ 7, 1875–present) earthquake epicenters. Possible subducting relief and the TPFR are dashed in fine gray and black, respectively. Outer arc high and Kuril Basin margins are thick and dashed in blue and black, respectively. Thin gray dashed lines show fracture zones [Matthews et al., 2011]. Solid and dashed red lines mark strike-parallel [Kimura, 1986] and extensional [Kulinich et al., 2007] faults accommodating fore-arc sliver translation. (b) Residual free-air gravity anomaly showing a correlation between the landward limit of thrust earthquakes and the TPFR crest (thick black dash). (c) Published asperities (shaded) and slip-distributions/aftershock areas for large magnitude (Mw ≥ 7.5) earthquakes. (d) Cross sections showing residual bathymetry (green), residual free-air gravity anomaly (black), and the geometry of the seismogenic zone [Hayes et al., 2012].
evolution, which suggest opening about a pivot located near Onekotan Island during the Middle-Miocene [Maeda, 1990], but is more consistent with the parallel nature of symmetrical Kuril Basin margins, fore-arc kinematics, and may explain abrupt changes in fore-arc morphology and seismogenic behavior.

The TPFR is interpreted as far as the north-eastern Kuril Basin margin and possibly ~200 km southeast of Kamchatka, but this latter extension is uncertain. The majority of the Kamchatka fore arc does not contain a TPFR, and fore-arc morphology and megathrust seismicity in this region are discussed in part 1 of this study [Bassett and Watts, 2015].

Two unnamed fracture zones (FZ) near the NE Hokkaido Rise define the sharp south-western boundary of a region of elevated near trench morphology and leave a ~1 km deep reentrant scar in the TPFR crest (Figure 4a). The Nosappu FZ is similarly associated with a saddle in the TPFR and a transition in outer fore-arc morphology, with the southern region rough and characterized by positive residuals. Subducting seamounts have been identified at the junction of the Kuril and Japan trenches [Lallemand and Le Pichon, 1987] and the similarity in near trench morphology may reflect the presence of subducting and/or accreted relief beneath the northern Japan and Kamchatka fore arcs. Southwest of the Nosappu Fracture Zone, the gravitational expression of the TPFR is narrower, rougher, and of lower amplitude, and earthquake distributions support this being the site of a second-order segmentation of the southern Kuril Islands.

Small and intermediate magnitude thrust earthquakes (Mw < 6.5) along the southern Kuril Islands are focused between the outer-arc high and the gravitational TPFR crest (Figures 4b and 4d). Larger magnitude thrust earthquakes are similarly distributed and we note that most epicenters in our catalog of large earthquakes (Mw ≥ 7.0; yellow stars, Figure 4a) are near the bathymetric TPFR scarp. Published slip and aftershock distributions for these events show predominantly updip and along-strike rupture propagation with few earthquakes showing any slip landward of the TPFR crest (Figure 4c). Where seismic asperities are identified, the majority of which in this region come from studies of seismic source time functions [Schwartz and

Figure 4. (continued)
Ruff, 1985; Beck and Ruff, 1987; Schwartz and Ruff, 1987], these also cluster near the gravimetric ridge crest along the seaward TPFR slope.

The resolution of slip near the trench for older events is likely uncertain, but in contrast to large earthquakes northeast of the Nosappu Fracture Zone, slip distributions and aftershocks for the 1952 and 1973 events further southwest do not extend to the trench, with rupture arresting within the morphologically rough and bathymetrically high near-trench region (Figure 4a). Near-trench morphology of this nature southeast of Kamchatka and along the northern Japan trench is similarly characterized by low densities of recent thrust earthquakes (Figures 4a and 4b). The along-strike rupture widths of large and great earthquakes do not appear limited by subducting fracture zones, but the sharp reduction in the number of earthquakes with Mw > 6.5 southwest of the Nosappu Fracture Zone supports morphological evidence of a second-order segmentation of the Southern Kuril Islands.

In contrast to the southern region, the frequency of large earthquakes along the Northern Kuril Islands is lower with the only significant earthquakes occurring in 2006 (Mw 8.3) [Lay et al., 2009] and 1915 (Ms 8.0) [Geller and Kanamori, 1977; Pacheco and Sykes, 1992]. These events are both located where the gravimetric expression of the TPFR is strong and have epicenters, coseismic slip/aftershock distributions, and asperities (2006 event) focused on steep seaward dipping gravity gradients (Figure 4c).

Along the Northern Kuril Islands, only two earthquakes have been recorded with 6.5 ≤ Mw ≤ 8 and are located between the Bussol Graben and the interpreted position of the original southwestern Kuril Basin margin (labeled 3 and 2, respectively). These earthquakes and the majority of small magnitude thrust events located further east, cluster on or seaward of the TPFR crest thereby extending the correlation between the TPFR crest and the downdip limit of seismic slip observed further south.

Abrupt changes in the seismological character of the plate boundary appear correlated with morphological transitions in the fore arc and the lineations interpreted as the Kuril Basin margins. Between these lineations, the Pacific Plate is potentially subducting beneath thinner Kuril Basin crust, which may explain the apparently lower frequency and magnitude of earthquakes occurring along the Northern Kuril Islands segment.

From observations made thus far in Tonga-Kermadec and the Kuril Islands, we suggest the key relationships observed between fore-arc morphology and the character and distribution of seismicity are as follows:

1. Intermediate magnitude thrust earthquakes are focused seaward of the gravimetric TPFR crest with larger magnitude events focused on steep, seaward dipping bathymetric and gravimetric gradients near the ridge crest.

2. The epicenters and seismic asperities of large and great earthquakes are also focused on steep gravity anomaly gradients seaward of and proximal to the TPFR crest.

3. Rupture areas of large and great earthquakes show predominantly updip and along-strike rupture propagation with minimal coseismic slip realized landward of the TPFR crest.

4. A close correlation is observed between the morphological, gravimetric, and seismological segmentation of subduction margins, and segment boundaries are predominantly controlled by preexisting upper and lower plate structure, such as the Kuril Basin margins, subducting fracture zones, and the Louisville and Hawaii-Emperor seamount chains.

4.3. Chile

The longest TPFR observed in the circum-Pacific region parallels the coastline of Chile for >2000 km between 42°S and 20°S and is coincident with the Chilean coastal cordillera (Figure 5). The irregular morphology of this TPFR likely reflects the structural complexity of the subducting Nazca plate, upper plate tectonics, and the effects of subaerial erosion. The highest amplitude anomalies (~250 mGal and >1.5 km) are observed in northern Chile and the triangular prismatic form and seaward asymmetry in topographic gradients are consistent with TPFR observed in Tonga and seaward of the Kuril Islands.

The tectonic fabric of the Nazca plate is well expressed in grids of residual free-air gravity anomaly and the TPFR is truncated and segmented along strike by subducting seamount chains, fracture zones, and upper plate structures contiguous between the fore arc and back arc. Two prominent lineations truncating the TPFR, labeled 1 and 2 in Figures 5a and 5b, can be identified over 300 km beyond the TPFR.
Figure 5. Chile subduction zone. Symbols as in Figure 3. (a) Uninterpreted residual free-air gravity anomaly. Arrows denote TPFR offsets 1 and 2 discussed in text. (b) Interpreted residual free-air gravity anomaly. SCR = South Chile Rise; LOFZ = Liquiñe-Ofqui Fault Zone; ETF = El Tígre Fault; AFZ = Atacama Fault Zone; PFS = Precordillera Fault System. The TPFR crest is marked as a thick black dash. (c) Published asperities and slip-distributions/aftershock areas for large magnitude earthquakes. Insets show the correlation between the TPFR crest and the downdip limit of strong geodetic locking [Chlieh et al., 2011; Metois et al., 2012]. (d) Cross sections showing residual bathymetry (green), residual free-air gravity anomaly (black), and the geometry of the seismogenic zone. Deep seismic reflection data from the ANCORP profile (data location shown in Figure 5c [Buske et al., 2002]) are projected onto cross-section A-A'. Interseismic coupling is shown on cross-sections C-C' and D-D' [Metois et al., 2012].
Figure 5. (continued)
Lineation 1 separates two well-defined blocks in which the TPFR displays the highest amplitude residual free-air gravity anomalies (Figure 5a). Inland, this lineation is correlated with steps in both the Atacama (AFZ) and Pre-Cordilleran (PFZ) fault systems, marks the northern boundary of a prominent fore-arc basin and landward bight in the volcanic arc, and is correlated with a sharp increase in the density of arc volcanoes. Offshore, this lineation is contiguous with the bathymetric and gravitational trough marking the southern boundary of the Iquique Ridge.

Lineation 2 marks a sharp reduction in the width of the TPFR and is associated with a trench-normal clustering of arc volcanoes that mark the southern termination of the volcanic arc in Northern Chile. This gap in the arc is associated with a region of flat slab subduction, which relocated earthquakes attribute to the buoyancy of the Juan Fernandez Ridge subducting further south [Anderson et al., 2007]. Lineation 2 and the northern extent of this gap are also aligned with a chain of small seamounts on the Nazca plate that are parallel with the Juan Fernandez Ridge (Figure 5b).

This TPFR is further truncated by a possible unnamed fracture zone at 31°S (labeled FZ1) and by the Juan Fernandez Ridge, which are coincident with scallops in the coastline at La Serena (LS) and Valparaiso (V), respectively (Figure 5a). South of 35°S, the TPFR appears to widen toward the trench before bifurcating into coastline and near-trench components. The near-trench component likely reflects the shallower water depths associated with subduction of young (<25 Myr) buoyant oceanic crust and lithosphere of the Chile Rise and although segmented along strike by a gap between the Valdivia and Mocha fracture zones, the TPFR can be traced as far as 47°S.

In northern and southern Chile, the TPFR is separated from the volcanic arc by arc-parallel gravity anomaly lows marking the longitudinal valley and central valley, respectively (Figure 5b). The landward margins of these basins are defined by the Pre-Cordilleran and Liquine-Ofqui dextral strike-slip fault zones [Hoffmann-Rothe et al., 2006], which are considered analogous to faults accommodating westward sliver translation of the southern Kuril Islands TPFR. The absence of these faults in central Chile is coincident with a gap in the volcanic arc and may reflect the importance of arc-related heat flux in facilitating the detachment of fore-arc sliver plates in margins where convergence is not strongly oblique [McCaffrey, 1992; Wang, 1996].

The maximum residual free-air gravity anomalies are interpolated along strike to define the TPFR crest (dashed black in Figure 5b). Thrust earthquakes of all magnitudes are focused predominantly seaward of this lineation on steep seaward dipping gravity gradients (Figures 5b and 5d). Great earthquake epicenters are located proximal to and seaward of the TPFR crest, and coseismic slip models, aftershock areas, and earthquake asperities are similarly distributed (Figure 5c). Particularly strong relationships between the TPFR crest and the downdip limit of coseismic slip are displayed by recent and well-constrained earthquakes in 1985, 1995, 2007, and 2010 [Mendoza et al., 1994; Pritchard et al., 2006; Delouis et al., 2010; Loveless et al., 2010].

Observations of surface displacements from campaign GPS measurements and Synthetic Aperture Radar Interferometry (InSAR) have been used to calculate interseismic strain accumulation and the distribution of geodetic locking along the Central Andean (19°S–25.5°S) [Chlieh et al., 2011] and Central Chilean (30°S–42.5°S) [Metois et al., 2012] segments of the South American plate boundary. These models are displayed as insets in Figure 5c with the Central Chilean model [Metois et al., 2012] also displayed in cross-sections C–C’ and D–D’ in Figure 5d. In both models, a strong correlation is observed between the downdip limit of strong geodetic locking and the TPFR crest. This correlation has been reported for Northern Chile by Béjar-Pizarro et al. [2013] and is shown here to be similarly present in central Chile. This correlation is consistent with and further strengthens the apparent relationship between the TPFR crest and a change in slip-behavior of the subduction interface. We note that the segment of fore arc between the Mocha and Valdivia fracture zones, where the TPFR is not observed, is a zone of weak interseismic coupling (Figure 5d).

At least five discrete morphological segments may be interpreted on the basis of TPFR offsets, but only two of these potential segment boundaries show any correlation with changes in the seismogenic behavior of the megathrust. These segment boundaries are lineation 2 and the TPFR bifurcation marking the onset of Chile Rise subduction. Seismogenically, the former marks the southern termination of a particularly active segment of the megathrust extending north to 21°S, although the southward propagation of the 1922 earthquake was not limited by this boundary. The latter boundary marks the cessation of intermediate magnitude (5 ≤ Mw < 7) thrust seismicity along the Chile trench.
In northern Chile and near cross-section A-A’, the TPFR crest is coincident with a transition in the reflection character of the subduction interface at ~50 km depth from a narrow (~3 km thick) band of reflectors updip to the deeper and broader (5–10 km thick) Nazca reflection zone [Figure 5d] [ANCORP-Working-Group, 1999; Buske et al., 2002]. This transition is coincident with the downdip extent of locally recorded aftershocks following the 1995 Mw 8.0 Antofagasta earthquake [Buske et al., 2002] and may support hypotheses linking both the TPFR crest and a broadening of the zone of plate interface reflections with a transition from unstable to stable megathrust slip regimes [Nedimović et al., 2003; Kimura et al., 2010].

4.4. Aleutian Islands

The segmentation of the Alaska-Aleutian fore arc is modulated by discontinuities in the thickness and age of the overriding and subducting plates, and the east-west increase in the obliquity of Pacific Plate convergence (Figure 6) [Geist and Scholl, 1992; von Huene et al., 2012]. The boundary between the Alaska and Aleutian segments and the transition from ocean-ocean to ocean-continent subduction is coincident with the Beringian margin, the trenchward continuation of the Black-Hills ridge, and a major domain boundary fault that can be traced >1000 km along the Alaskan shelf (dashed in Figure 6b). This fault likely accommodated pre-Paleocene oblique convergence or strike-slip motion between the North American and Kula plates [Marlow and Cooper, 1980; Lewis et al., 1988]; and transverse faulting at the base of the inner trench slope and the 1948 earthquake that appears to have propagated along the fault trace suggest reactivation within the contemporary stress regime.

East of the Beringian margin, the transition to ocean-continent subduction is characterized in the landward half of the fore arc by a region of positive residual free-air gravity anomalies that is well correlated with the extensively studied Shumagin Islands seismic gap [Sykes, 1971; Kelleher et al., 1973; McCann et al., 1980; Davies et al., 1981; Boyd and Jacob, 1986]. This gap is likely associated with the transition in overthrusting plate structure and is defined by the absence of slip in the 1938 Mw 8.2 earthquake [Johnson et al., 1994] and 1957 Mw 8.7 Andreanof Islands earthquake [Johnson et al., 1994]. This gap has, however, ruptured in numerous small and intermediate magnitude (Mw ≤ 7) thrust events, in addition to a Ms 7.5 earthquake in 1948. The region around the residual gravimetric high is also suggested to have ruptured in 1917 (Mw 7.5–8.0) providing the first evidence of large-scale seismic-moment release in the Schumagin Islands region [Boyd and Lerner-Lam, 1988].

The Western, Central, and Eastern Aleutians are comprised of discrete clockwise rotating and westward translating fore-arc blocks, the number of which within each segment increases from east-west in conjunction with the increasing obliquity of Pacific plate convergence (Figure 6b) [Geist et al., 1988; McCaffrey, 1992; Ruppert et al., 2012]. Fore-arc blocks in the Western Aleutians are observed as discrete and irregular bathymetric and gravimetric highs that encompass the entire fore-arc. Fore-arc highs in the Central and Eastern Aleutians, in contrast, are observed as coherent trench-parallel linear ridges and are interpreted as TPFRs. The Western Aleutians contains few thrust epicenters with Mw ≥ 6.5, and the Central and Eastern Aleutians, where TPFRs are interpreted, have nucleated most thrust earthquakes occurring along the Alaska-Aleutian margin in the last century [Butler, 2012; Ryan et al., 2012b].

The TPFR within the Andreanof block (the Aleutian Ridge of Scholl et al. [1983] and Scholl et al. [1987]) is flanked by a zone of outer-arc highs that mark the geomorphic break between the lower slope of the Aleutian trench and the flat Atka fore-arc basin [Ryan and Scholl, 1989]. Negative anomaly values within this basin are amplified due to inclusion of the Alaska margin segment, which has significantly shallower near-trench topography, in the ensemble profile calculation (supporting information Figure 5); and the low residual anomalies (<−4 km and <−200 mGal) in the fore-arc basin contribute to steep seaward dipping topographic (5–10°) and gravimetric gradients (5 mGal km⁻¹) seaward of the TPFR crest. The crest of the Aleutian Ridge is bounded by a right-lateral intra-arc strike-slip fault and the eastern termination of this fault, and the Aleutian ridge, marks the morphological segment boundary between the Central and Eastern Aleutian Islands (Figure 6a).

The Eastern Aleutians is morphologically similar to the Central segment; however, the TPFR is broader, the fore-arc basin is shallower, and the zone of outer-arc bathymetric highs is not observed in the residual free-air gravity field.

In both the Central and Eastern Aleutian Islands the distribution of small magnitude thrust earthquakes is restricted to the region between the outer-arc high and the crest of the TPFR (Figure 6c). Large and
Figure 6. Aleutian subduction zone. Symbols as in Figure 3. (a) Residual free-air gravity anomaly and seismicity. The outer-arc high, trench-parallel fore-arc ridge and block-bound faults are dashed in blue, black, and red, respectively. Annotations are AP = Amchitka Pass; BHR = Black-Hills Ridge; SS = Sunday Summit Basin; PD = Pratt Depression. (b) Published asperities and slip-distributions/aftershock areas for large magnitude earthquakes. (c) Cross sections showing residual bathymetry (green), residual free-air gravity anomaly (black), and the geometry of the seismogenic zone [Hayes et al., 2012]. Uninterpreted residual grids for this region are shown in supporting information Figures 4 and 5.
intermediate magnitude earthquakes (6.5 < Mw < 8) are focused landward of the fore-arc basin depocenter and track the steep topographic and gravimetric gradients of the TPFR. Large segments of the Andreanof block, which contains the deep Atka fore-arc basin and displays the steepest seaward dipping gravimetric and topographic gradients, ruptured in 1986 (Mw 8.0), 1996 (Mw 7.9), and this block also nucleated and contained three of four asperities in the 1957 (Mw 8.7) Andreanof Islands earthquake (Figure 6b) [Johnson and Satake, 1994]. Seismicity within the Eastern Aleutian Islands is predominantly of lower magnitude and apart from large events in 1946 (Mw 8.2) and 1948 (Ms 7.5), only eight earthquakes since 1976 have had Mw ≥ 6.5 and all had Mw ≤ 7. All large and great earthquakes epicenters within this segment are located near the Alaska/Aleutian transition from ocean-ocean to ocean-continent subduction.

The Rat block is morphologically distinct and nucleated the 2003 Mw 7.8 earthquake, which ruptured the majority of the Rat Block megathrust [Balakina and Moskvina, 2009], and the 1965 Mw 8.7 Rat Islands earthquake, which propagated almost entirely to the west rupturing the Rat, Buldir, and Near fore-arc blocks (Figure 6b) [Beck and Christensen, 1991]. The Rat block has also acted as a barrier to rupture propagation and Mw 8.7 earthquakes in 1957 and 1965 both arrested near the eastern boundary of Rat Island [Beck and Christensen, 1991; Johnson et al., 1994].

Amchitka Pass (AP in Figure 6a) is a bathymetric canyon marking the eastern boundary of the Rat block and appears wider than other fault controlled canyons along the Aleutian arc. Anticlockwise rotation of the Rat block and closure of the Sunday summit basin and Prat depression (labeled SS and PD in Figure 6b) associates Rat Island with the extension of Bowers Ridge [Ryan and Scholl, 1993]. This ridge is believed to be an extinct island arc [Kienle, 1971; Kawabata et al., 2010; Wanke et al., 2012] and multichannel seismic reflection data reveal a deep sediment-filled trench like structure on its northeastern side [Cooper et al., 1981; Scholl, 2007]. We associate Rat Island and Amchitka Pass with the southerly extension of the island-arc and trench axis of this Cenozoic subduction zone and suggest upper plate structures associated with this paleo-plate boundary may have acted to limit along-strike rupture propagation of earthquakes in 1957, 1965, and 2003.

In the eastern Andreanof block, the eastern limit of small magnitude (Mw < 6.5) thrust seismicity and the eastern limit of the Atka fore-arc basin is coincident with the intersection of the Amlia fracture zone with the Aleutian trench [Ryan et al., 2012a]. This fracture zone offsets Tertiary Pacific Plate isochrons by left-lateral offsets of 220 km [Grim and Erickson, 1969], corresponding to a west-east age step of +6 Myr and separates the rough and heavily bend-faulted crust with a prominent outer-rise to the west from the smoother and bathymetrically shallower subducting crust to the east.

We suggest TPFRs in the central and western Aleutians are analogous to TPFRs observed in Tonga-Kermadec, the Kuril Islands, and Chile. In reiterating the key relationships interpreted thus far between TPFR morphology and the character and distribution of megathrust seismicity, we suggest these relations are upheld by observations made along the central and eastern segments of the Aleutian Islands arc.

1. Intermediate magnitude thrust earthquakes are focused seaward of the gravimetric TPFR crest with larger magnitude events focused on steep, seaward dipping bathymetric and gravimetric gradients near the ridge crest.
2. The epicenters and seismic asperities of large and great earthquakes are also focused on steep gravity anomaly gradients seaward of and proximal to the TPFR crest.
3. Rupture areas of large and great earthquakes show predominantly updip and along-strike rupture propagation with minimal coseismic slip realized landward of the TPFR crest.
4. A close correlation is observed between the morphological and seismological segmentation of the subduction margin, and segment boundaries are predominantly controlled by preexisting upper (e.g., Bowers Ridge and the Beringian margin) and lower (e.g., Amlia, Adak, and Rat Island fracture zones) plate structure.

4.5. Cascadia

Oblique subduction of the Juan de Fuca plate beneath North America has segmented the Cascadia fore-arc into large rotating and northward migrating blocks [Wells et al., 1998; Wells and Simpson, 2001]. Fore-arc kinematics are dominated by clockwise rotation of the Oregon block resulting in uplift and transpression in western Washington as it is compressed against the Canadian Coast Mountains restraining bend (Figure 7a). Residual free-air gravity anomalies reveal a TPFR along the full length of the Cascadia fore-arc, associated with the uplifting Cascadia coastal ranges (Figure 7b) [Kelsey et al., 1994]. This TPFR, and much of the Cascadia fore-arc basement, is comprised of Paleocene and Early-Eocene basalts, which are up to 35 km
thick and collectively termed the Siletzia terrane [Trehu et al., 1994]. The origin of this terrane is uncertain and suggested possibilities include formation above a hot spot and buoyancy-driven accretion [Duncan, 1982; Trehu et al., 1994], or in situ formation during oblique rifting along the leading edge of the continental margin [Snively and Wells, 1996].

The TPFR is ~80 km wide within the Oregon block and near 45°N, a sharp increase in TPFR amplitude (~100 mGal) is correlated with a transition in morphology of the outer fore-arc (Figures 7a and 7b). A second transition in TPFR structure is coincident with thrust faults marking the boundary between the Oregon and Washington blocks [Wells et al., 1998; Wells and Simpson, 2001], north of which the TPFR increases in width (~90–100 km) to encompass the full width of Vancouver Island (Figure 7b). The TPFR crest is defined as the seaward limit of a broad zone of high amplitude, positive and irregular gravity anomalies and is dashed black in Figures 7b and 7d.

Despite a history of great earthquakes [Goldfinger et al., 2003; Leonard et al., 2004; Leonard et al., 2010], the Cascadia megathrust has generated no earthquakes with Mw $\geq 6.5$ in the last 50 years. This prevents us from analyzing the relationship between the TPFR and seismicity in the way we have for other margins, however, the Cascadia fore-arc hosts a dense network of onshore seismometers with the deployment coupled to an automatic detection and location algorithm for recording nonvolcanic tremor [Wech and Creager, 2008].

In contrast to normal earthquakes of similar magnitude, the lack of high-frequency content in radiated seismic waves suggests nonvolcanic tremor results from a slow, low stress-drop process [Obara, 2002], possibly associated with high pore fluid pressure [Kodaia et al., 2004; Kao et al., 2005; Shelly et al., 2006]. Spatiotemporal correlations between tremor and slow slip events (SSEs) in Japan [Obara et al., 2004] and Cascadia [Rogers and Dragert, 2003] contribute to a growing body of evidence suggesting tremor and slow slip are different manifestations of the same shear process at depth [Shelly et al., 2006; Shelly et al., 2007; Wech and Creager, 2007, 2011]. Tremor epicenters can thus be used to supplement geodetic spatial constraints on the slow-slip region of the plate interface.

Most SSEs occur on subduction interfaces at 30–45 km depth [Dragert et al., 2001; Larson et al., 2004; Obara et al., 2004; Ohta et al., 2004, 2006; Peng and Gomberg, 2010] and are commonly interpreted as occurring within a transition zone between the updip, strongly coupled, velocity-weakening, and seismogenic region of the subduction interface, and the downdip, steadily creeping, velocity-strengthening, and aseismic region beneath [Schwartz and Rokosky, 2007]. In the margins discussed thus far, we have demonstrated that megathrust earthquakes are predominantly focused on the seaward dipping flank of the TPFR and suggest a correlation between the TPFR crest and the downdip limit of the seismogenic zone. One clear prediction of this hypothesis is that slow-slip events, and hence tremor epicenters, should be focused landward of the TPFR crest.

Episodic Tremor and Slip (ETS) describes the simultaneous observation of tremor and slow-slip, and in northern Cascadia, Wech et al. [2009] use tremor epicenters from four ETS events to place detailed mapview constraints on the slow-slip region. Tremor density is shown to correlate with cumulative quantities of simultaneous slow-slip and overlaying tremor density on grids of residual free-air gravity anomalies show that ETS tremor epicenters are indeed focused on the landward flank of the TPFR (see inset in Figure 7b). The sharp updip limit of tremor epicenters, which is correlated with the TPFR crest, is interpreted by Wech et al. [2009] as marking a change in the coupling properties of the plate interface and likely reflects the updip limit of the frictional transition zone.

We expand this analysis along the Cascadia margin by calculating a regional grid of tremor density using all reported tremor epicenters since August 2009 [Wech and Creager, 2008; Wech, 2010]. Each epicenter represents 5 min of tremor and tremor density is calculated using a 5 km search radius (Figure 7c). This analysis reveals a strong regional correlation between the updip limit of tremor and the TPFR crest which is dashed in black.

The density and distribution of tremor appears correlated with fore-arc morphology and may also be useful in constraining along-strike segmentation. Within the Oregon block, tremor density is low and focused within a narrow band along the landward TPFR flank (Figure 7c). Formation of a contiguous tremor band near 44.5°N is correlated with the transition in near-trench morphology and the south-to-north increase in
BASSETT AND WATTS  
CRUSTAL STRUCTURE AND SEISMICITY: 2. FORE-ARC STRUCTURE  
18
TPFR amplitude. A second along-strike transition is coincident with thrust faults defining the Oregon and Washington fore-arc block boundary [Wells et al., 1998; Wells and Simpson, 2001], north of which TPFR width and both the density and lateral spread of tremor epicenters sharply increase. Assuming all tremor takes place on the plate interface, the geometry of the Juan de Fuca slab from Audet et al. [2010] suggests that tremor occurs at 25–40 km depth, with the updip limit of tremor particularly responsive to the 40 km slab isobath (Figure 7c).

In Figure 7d, we show the locked and effective transition zones (ETZ) calculated from a 3-D dislocation model of interseismic deformation rates [Wang et al., 2003]. The fully locked zone is defined by the position of the 350°C isotherm, which is modeled from heat flow observations and located entirely offshore at interface depths shallower than ~15 km [Hyndman and Wang, 1993, 1995; Hyndman and Rogers, 2010]. The ETZ is modeled from horizontal deformation data including strain rates and surface velocities, has an exponentially decreasing slip distribution updip, and only the updip half of the ETZ (coupling >0.25) is included in the calculation of coseismic deformation [Wang et al., 2003].

This region of coseismic deformation is located seaward of both the updip limit of tremor and the crest of the TPFR (thick black dash in Figure 7d). Assuming the close correlation between these boundaries suggests links with the megathrust source zone are similarly applicable in Cascadia, the seismogenic zone of this subduction interface may extend >100 km further landward in some regions than as predicted in the model of Wang et al. [2003]. This suggestion has significant implications for potential hazards associated with the Cascadia megathrust, but we stress that this is only one of several possible interpretations for the correlation between the TPFR crest and tremor; and we have not considered the many other data sets required to qualify this study to make strong conclusions concerning the width of the seismogenic zone in Cascadia [Hyndman, 2013]. Other possible interpretations are described in our discussion, but, assuming links with the megathrust source zone are valid, below we consider one possible explanation for the contrast between the greater width possibly inferred by this study and the results of thermal/geodetic models of plate coupling and coseismic slip.

Relative to earlier thermal models, including the effects of fluid circulation within the oceanic crust and constraining thermal models by the location of the seismically observed basalt-to-eclogite transition [Bostock et al., 2002; Nicholson et al., 2005; Rondenay et al., 2008; Abers et al., 2009] is suggested to widen the thermally defined seismogenic zone in Cascadia by up to 55 km [Cozzens and Spinelli, 2012]. The 350 and 450°C isotherms from Cozzens and Spinelli [2012] are dashed in black in Figure 7d and these bound the updip limit of tremor and the TPFR crest. Associating the updip limit of ETS tremor with the base of the seismogenic zone, Chapman and Melbourne [2009] show that low degrees of plate coupling onshore are not required to fit GPS surface velocities in Washington or coseismic coastal subsidence in Oregon and British Columbia.

In Japan [Mochizuki et al., 2005; Kimura et al., 2010] and Chile [Buske et al., 2002], the reflective character of the subduction interface has been linked with slip behavior and the reflective domains of the Cascadia megathrust are displayed in Figures 7a and 7d [Nedimović et al., 2003]. The transitional domain R2 separates regions of thin (~<2 km = R1) and thick (~5–7 km = R3) interface reflection bands. Thin reflection bands (R1) are associated with the shallow locked section of the fault, the updip limit of which is well correlated with the 350°C isotherm of Cozzens and Spinelli [2012]. The thicker E reflection band (R3) is associated with stable sliding and slow slip. Receiver function analyses indicate low velocities at the depth of the E reflection band [Cassidy and Ellis, 1993], magnetotelluric data suggest high porosities above the subducting slab [Kurtz et al., 1986; Hyndman, 1988], and the E reflection band is interpreted as a zone of lower crustal underplating where crustal rocks are imbricated and accreted within a wide deformation zone above the subducting plate [Green et al., 1986; Calvert and Clowes, 1990; Calvert, 1996; Calvert et al., 2003; Calvert, 2004]. The updip limit of E-band reflections is ~20–30 km updip of the shallowest tremor epicenters near Vancouver Island.
and offset updip by ~50 km further south. The downdip limit of E-band reflections is near the 40 km slab isobath [Audet et al., 2010], bounds the downdip limit of large tremor densities (Figure 7d), and is coincident with the intersection of the megathrust with the fore-arc Moho [Calvert, 2004].

Southwest of Oregon, a second zone of high reflectivity above the plate interface (green box in Figure 7d) [Keach et al., 1989] is suggested to coincide with a low-resistivity (<10 Ω m) layer resolved beneath the Oregon coast range [Wannamaker et al., 1989]. Low resistivity is associated with pore water or sediments along the plate interface, high reflectivity is associated with magmatically or tectonically underplated material [Trehu et al., 1994], and the correlation with tremor may suggest the E-reflection band extends southward along the Cascadia margin following a geometry similar to that of the TPFR [Kurtz et al., 1986; Hyndman, 1988]. We note that this zone is similarly located near the 40 km depth contour of the subducting slab and is likely near the subducting slab–fore-arc Moho intersection [Trehu et al., 1994].

The Cascadia TPFR appears analogous to those interpreted along the other margins discussed thus far, but the absence of recent thrust earthquakes make it difficult to know if links with the dimensions of the seismogenic zone are similarly valid. The observed correlation with the updip limit of tremor may suggest that they are, but this correlation (and the TPFR) may also reflect the influence of lower crustal underplating on surface morphology, or be simply coincidental. These possibilities are each considered in greater detail in our discussion.

5. Residual Free-Air Gravity Anomalies As a Proxy for Seismic Moment Release

The interrelationships we propose between fore-arc structure and seismicity have important implications for the spatial distributions of both present and future earthquakes, but are often inconsistent with the suggestion by Song and Simons [2003] of a strong correlation between regions of high seismic-moment release and negative (<−40 mGal) trench-parallel gravity anomalies (TPGA). TPGA are analogous to the residual free-air gravity anomalies we have calculated from spectral averaging and below, we reconsider links between residual free-air gravity anomalies and seismic-moment release.

Using all thrust earthquakes with 6.0 ≤ Mw < 8.0 from the GCMT catalog (1976 to July 2013), empirically derived relations [Kanamori, 1983] are applied to calculate the slip area from seismic moments assuming a constant 1 MPa stress drop [Kanamori, 1977]. A square rupture geometry is assumed and centered on the earthquake centroid with sides parallel and perpendicular to dip, which remains a square unless either the trench or downdip limit of thrust seismicity is surpassed, in which case the square becomes rectangular extending along strike until the required rupture area is realized. These rupture areas are supplemented by published rupture characteristics for >150 of the largest (Mw 7–9.5) earthquakes occurring along subduction megathrusts since 1875 and compared with the distribution of residual topography and free-air gravity anomalies.

The distribution of residual topography and free-air gravity anomalies within seismogenic zones are shown by the gray-shaded histograms in Figure 8. With respect to residual grid values, the black lines reveal similar and approximately normal distributions of global seismic-moment release for GCMT thrust earthquakes with 6.0 ≤ Mw < 8.0. Including and replacing associated GCMT events with published rupture areas for large (Mw ≥ 7) earthquakes since 1875 does not significantly alter this distribution (blue line in Figure 8). Large earthquake epicenters and seismic asperities are normally distributed with respect to residual free-air gravity anomalies but skewed to positive values of residual bathymetry. This is consistent with the general observation of both epicenters and seismic-asperities being located downdip and near TPFR crests where present. Individual seismic-moment distributions for each margin are provided in supporting information and although some margins display concentrations of seismic-moment release in regions of either positive (e.g., Japan, Kuril/Kamchatka) or negative (e.g., Tonga-Kermadec) residual free-air gravity and/or topography, no systematic relationship is observed and most moment distributions closely match the distribution of residual grid values within the seismogenic zone.

Our analysis does not reveal a significant global concentration of seismic-moment release in regions of positive or negative residual free-air gravity anomalies and below we consider methodological differences between our study and Song and Simons [2003]. We analyze similar data [Sandwell and Smith, 1997, 2009; Sandwell et al., 2014] and although our processed grids are of higher resolution, which demonstrates the advantages of constructing and removing an ensemble average grid, the first-order structure and amplitudes of the residual free-air gravity fields derived are comparable (supporting information Figure 4).
The slip areas calculated for events in the GCMT catalog are restricted to the region between the trench and the downdip limit of seismogenic slip. This lower limit is defined by Song and Simons [2003] using the 50 km slab contour of Gudmundsson and Sambridge [1998]. In our study, this limit is defined using the distribution of earthquakes with focal depths (<65 km) and mechanisms consistent with interplate slip, and the landward extent of slip in published coseismic slip distributions and aftershock locations. A comparison of the 50 km slab contour of Gudmundsson and Sambridge [1998] with subducting SLAB1.0 geometries [Hayes et al. 2012] shows that this contour is predominantly associated with shallower SLAB1.0 depths, which are as low as 20 km in Japan and the Kuril Islands (supporting information Figure 6). These slab geometries appear closest along the Alaska-Aleutian and South American margins, but in most other regions (e.g., Japan, Kuril Islands, and Tonga-Kermadec) this 50 km depth contour significantly underestimates the width of the seismogenic zone. Supporting information Figure 7 shows that in the Kuril Islands, almost all thrust earthquakes occur on the seaward dipping TPFR flank in regions where residual free-air gravity anomalies are not strongly negative. These events, however, are also located landward of the 50 km slab contour of Gudmundsson and Sambridge [1998] and would thus be excluded from the analyses under the definition of Song and Simons [2003].

We do not include events from the ISC catalog, nor do we limit our use of the GCMT catalog to events with a quality factor of A or B. The results of Song and Simons [2003] are not dependent on the earthquake catalog used and the GCMT catalog is shown to be sufficient to resolve the global correlation proposed. We
also do not limit our analyses to offshore regions. This constraint has the largest influence in Mexico and along the South American subduction zone where most thrust earthquakes occur onshore and in regions of generally positive residual free-air gravity anomalies (Figures 5a and 5b).

We suggest the key differences between our analysis and that of Song and Simons [2003] are related to our contrasting definitions of the downdip limit of thrust seismicity and the exclusion of onshore earthquakes. In Song and Simons [2003], both restrictions appear to exclude earthquakes occurring in regions of nonnegative residual free-air gravity anomalies.

Finally, we reiterate that the absolute values of residual anomalies, which simply reflect the degree of structural heterogeneity within the margin segments analyzed, are strongly influenced by its along-strike extent, and have no direct physical meaning. As demonstrated in Figure 5 of supporting information, residual grid values along the Alaska-Aleutian margin differ by up to 50 mGal and 1 km depending on whether average profiles and residuals are calculated across the whole margin, or individually for the Alaska and Aleutian segments. The short-wavelength structure of residual grids, however, does not change, which validates our focus on interpreting the detailed fore-arc structure revealed as opposed to absolute anomaly values.

6. Discussion

6.1. Structural and Seismological Characteristics of Trench-Parallel Fore-Arc Ridges

The relationships between structural, seismological, and geodetic observations in regions where TPFRs are observed are summarized in Figure 9.

TPFRs occupy the inboard half of at least five circum-Pacific subduction fore-arcs and represent the landward (positive) half of a paired anomaly containing a broader negative low nearer the trench (Figure 9a). This low is often coincident with the deep-sea terrace basin [Wells et al., 2003]. The positive half of this anomaly is expressed in the residual free-air gravity field as a seaward verging triangular anomaly with typical amplitudes and half wavelengths of 160–240 mGal and 75–100 km, respectively. The residual bathymetric expression of TPFRs, is highly asymmetric with the steep (up to 20°), rough and faulted seaward ridge flank contrasting with the smooth and shallowly dipping (1–3°) morphology of the landward slope. Trench-normal profiles traversing paired anomalies in the Kuril Islands (Figure 4d) and Chile (Figure 5d) show a good correlation between the residual gravity anomaly and bathymetry; and gradients of ~75 mGal/km (R = 0.62) and ~50 mGal/km (R = 0.8), respectively, suggest these ridges are isostatically uncompensated at wavelengths of ~200 km (supporting information Figure 2).

Seismic reflection data acquired across fore-arcs in Cascadia [Keach et al., 1989; Trehu et al., 1994; Nedimović et al., 2003; Calvert, 2004], Northern Chile [ANCORP-Working-Group, 1999; Buske et al., 2002], and Southern Hikurangi [Henrys et al., 2013] show a transition in the reflective properties of the subduction interface that occurs near or slightly seaward of the TPFR crest (Figures 5d and 7d). The plate interface updip is characterized by a narrow (~2–3 km thick) reflector band and this reflection character is typically associated with the interseismically locked, velocity-weakening, and seismogenic region of the subduction interface. Broader (~5–10 km thick) reflector bands observed at greater depth are typically associated with the weakly coupled, velocity strengthening, and aseismic region of the megathrust and have been linked with lower crustal underplating and the imbrication of subducted and tectonically eroded strata within a wide deformation zone above the subducting plate [Green et al., 1986; Calvert and Clowes, 1990; Trehu et al., 1994; Calvert, 1996; Calvert et al., 2003; Calvert, 2004; Henrys et al., 2013]. Receiver function analysis and seismic tomography indicate low seismic-velocities and low (~400) Qp (inverse of attenuation) at the depth of thick reflection bands [Cassidy and Ellis, 1993; Eberhart-Phillips et al., 2005; Reynolds and Eberhart-Phillips, 2009] and magnetotelluric data show low resistivities consistent with the presence of pore water and sediments along the plate interface [Kurtz et al., 1986; Hyndman, 1988; Wannamaker et al., 1989].

Our observations suggest a relationship may exist between the crest of TPFRs and the downdip limit of the seismogenic zone. Small and intermediate magnitude earthquakes (Mw < 7) occur between the outer-arc high and the TPFR crest, with the epicenters and asperities of larger earthquakes (Mw ≥ 7) located toward the rear of the seismogenic zone in regions of steep seaward dipping gravity gradients. Coseismic rupture and aftershock areas show predominantly updip and along-strike rupture propagation with few earthquakes showing any slip landward of the TPFR crest.
In Cascadia, slow-slip events and nonvolcanic tremor are recorded beneath the landward slope of the TPFR (Figure 7c); and correlations between the TPFR crest with the updip limit of tremor in Cascadia and the downdip limit of strong geodetic locking in Chile (Figures 5c and 5d) [Chlieh et al., 2011; Metois et al., 2012] are consistent with suggestions linking the TPFR crest to a change in the physical and mechanical properties of the megathrust.

6.2. Interpretation of TPFR
In Tonga-Kermadec, the Kuril Islands and Chile, the TPFRs we interpret within the fore-arc are related to extinct volcanic arcs, specifically the Tonga Ridge, the Chilean coastal cordillera, and the Vityaz Ridge. In the
Aleutian arc, the Aleutian Ridge is also composed of arc basement that likely formed in Early Eocene time [Scholl et al., 1987] and active arc volcanoes are located along the Northern (landward) flank of the Aleutian ridge [Ryan and Scholl, 1989, 1993]. In Cascadia, the TPFR is likely comprised of Paleocene and Early-Eocene basalts of the Siletzia terrane [Duncan, 1982; Trehu et al., 1994; Snively and Wells, 1996].

We suggest three interrelated mechanisms may contribute to the correlation that we have observed between fore-arc morphology, gravity anomalies, and seismicity. We first consider the processes influencing the growth, preservation, or destruction of these ridges, as these processes are most likely to be related to the mechanical behavior of the subduction interface. We then describe a possible tectonic model, which provides one mechanism by which extinct volcanic arcs may later occupy subduction fore-arcs. We note that each of these mechanisms appear to generate steep topographic gradients near the downwarp limit of coseismic slip and may thus offer an alternate interpretation for the correlation observed in some regions between the coastline and the landward extent of thrust earthquakes [Ruff, 1989].

### 6.2.1. Subduction Erosion and Lower Crustal Underplating

Seaward of the TPFR, inferences of sustained subsidence of the deep-sea terrace basin are confirmed by ocean drilling and seismic profiling. von Huene and Scholl [1991] associate this subsidence with basal tectonic erosion of the fore-arc crust by the subducting slab. Rates of basal erosion estimated in Peru, Northeast Japan, and Northern Chile are 25–50 km³ Myr⁻¹ and erosion is documented from similar observations off Costa Rica, Tonga, Ecuador, Alaska, and the Aleutian Islands [Scholl et al., 1980; von Huene and Lallemand, 1990; von Huene and Scholl, 1991; von Huene et al., 2000; Collot et al., 2002; Scholl et al., 2002; Wells et al., 2003; Vannucchi et al., 2004]. Recognizing a link between the asperities of megathrust earthquakes and free-air gravity constraints on the geometry of fore-arc basins, Wells et al. [2003] propose a long-term relationship between seismic slip and fore-arc/basinal structure. The key ingredient in this model is subduction erosion, which provides a means of achieving sustained permanent subsidence of the outer-fore-arc, as predicted if elastic deformation is not recovered, within the space constraints imposed by the subducting slab.

Considering the paired nature of the residual bathymetry and gravity anomalies, we extend this model landward and interpret the reflection character of the subduction interface as constraining the sign of local fluxes of fore-arc material (Figure 10b). While thin interface reflector bands typically underlie the outer-forearc, where inferences of subduction erosion are made, the wide interface reflector bands observed proximal to the TPFR are interpreted as documenting lower crustal underplating [Green et al., 1986; Calvert and Clowes, 1990; Trehu et al., 1994; Calvert, 1996; Calvert et al., 2003; Calvert, 2004; Henrys et al., 2013]. This interpretation is consistent with observations of tectonic erosion beneath the outer fore-arc being paralleled by underplating further landward in Chile and Peru [Clift and Hartley, 2007], interpretations of long-term uplift of coastal ranges in Mexico [Ramirez-Herrera et al., 2004; Ramirez-Herrera et al., 2011] and Cascadia [Kelsey et al., 1994], and inferences of underplating from seismic reflection and refraction observations in Alaska [Moore et al., 1991] and along the Hikurangi margin [Sutherland et al., 2009; Bassett et al., 2010; Scherwath et al., 2010; Henrys et al., 2013]. Hence, we adopt the underplating mechanism proposed by Sutherland et al. [2009] in suggesting that preferential (seaward) tectonic erosion, oversteepening of the trench-slope, and some degree of (not necessarily complete) lower crustal underplating may contribute in creating the paired anomaly we observe within fore-arcs.

In this model (Figure 10b), sediments on the downgoing plate are incorporated with material tectonically eroded from the toe and base of the trench slope into a subduction channel located above the subducting slab (thin reflector band). Accompanying subduction, this material is transported landward and downward to a position near the fore-arc Moho. In the lower crust, the density contrast between subducted sedimentary and crustal material (~2300–2700 kg/m³) and the mantle wedge (~3200 kg/m³) prevents further subduction. The material in the subduction channel is weakened by temperature and fluid overpressure and accreted across the subduction thrust to the upper plate, driving local rock uplift and forming a thick interface reflector band. Inboard uplift synchronous with outboard erosion causes the trench slope to steepen above its critical angle of stability, which may explain the rough, steep, and faulted morphology of the seaward dipping flank of the TPFR. Gravitational collapse takes place both arcward and trenchward of the locus of underplating, with trenchward collapse creating the highly faulted trench slope and delivering material to the deep-sea terrace basin or subduction front. In Cascadia, truncation of the E-reflection band at the intersection of the megathrust with the fore-arc Moho [Calvert, 2004] supports this model of underplating.
Figure 10. Schematic diagram illustrating the spatial relationships between the structure and seismogenic behavior of subduction zones and observations of plate interface reflectivity.
and is similar to analogous observations made in Hikurangi [Sutherland et al., 2009; Bassett et al., 2010; Scherwath et al., 2010; Henrys et al., 2013].

Inferences of underplating from seismic reflection observations often extend seaward of the TPFR crest. For underplating to occur, the subduction interface must step down from the roof of the subduction channel and Henrys et al. [2013] suggest the required increase in fault dip may be a stronger control on the transition from unstable to stable slip regimes than any relative weakness associated with underplated material. Assuming interface reflection band thickness is a valid constraint on the region undergoing basal tectonic erosion, the seaward offset of thick interface reflection bands with respect to the TPFR crest may explain the seaward offset occasionally observed between the topographic and gravimetric crests of TPFRs (Figure 10b).

### 6.2.2. Interseismic Elastic and Inelastic Deformation

Simulating plate coupling through the application of back slip at the plate convergence rate over the locked section of the fault, the broad pattern of interseismic deformation can be modeled by placing an edge dislocation at the downdip end of the locked fault with its Burgers vector parallel to fault dip [Savage, 1983]. This model predicts a broad interseismic subsidence over the majority of the locked fault. Maximum uplift occurs near the downdip limit of locking (Figure 11a), which appears to mirror the correlation we observe between the TPFR crest and the downdip limit of coseismic slip.

In Figure 11b, profiles of the residual free-air gravity anomaly across TPFR observed in the Kuril Islands and Chile are plotted against red profiles showing the vertical deformation predicted by the dislocation model assuming linear faults, average dips of the subducting slab [Hayes et al., 2012] and constraining the downdip limit of locking by the distribution of thrust earthquakes near the plate interface.

Recognizing the landward TPFR high and seaward flanking low as a paired anomaly, and noting the striking similarity in the form of this anomaly pair with the predicted distribution of interseismic uplift and subsidence, it is possible both the high and low may represent some fraction of the elastic deformation accumulated in the interseismic period that is not recovered in earthquakes or postseismic slip [Wells et al., 2003].

In Cascadia, Kelsey et al. [1994] compare geodetic uplift rates from resurveyed benchmarks (1941–1988) with uplift rates calculated from the elevation of marine terraces uplifted since the Quaternary. Geodetic uplift rates are up to an order of magnitude larger than vertical crustal velocities from uplifted shore platforms, suggesting contemporary strain accumulation on the plate interface. Latitudinal correlation between high rates of interseismic uplift with high rates of long-term uplift led Kelsey et al. [1994] to interpret long-term uplift as representing interseismic strain not recovered in earthquakes. Assuming observations of interseismic and long-term uplift were made in similar locations with respect to the locked fault, the fraction of interseismic strain accommodated inelastically may be ~5–10%.

### 6.2.3. Fore-Arc Tectonics

We suggest that paired fore-arc anomalies are sculpted from arc crust by the mechanical behavior of the underlying megathrust. Preferential subduction erosion beneath the outer fore-arc and above much of the seismogenic zone results in permanent subsidence and the formation of deep-sea terrace basins [Sugiyama, 1994; Wells et al., 2003]. The presence of a long linear trench-parallel ridge landward of these basins may reflect: (a) uplift associated with lower crustal underplating (Figure 10); (b) the transformation of some proportion of interseismic strain into permanent geologic strain via faulting, folding, or buckling of the inner-fore-arc over many earthquake cycles (Figure 11), and/or (c) the operation of a fore-arc net crustal deficit whereby the rate of erosion is greater than rates of frontal accretion and lower crustal underplating.

The latter is schematically illustrated in Figure 12. Assuming consistency in the trench-arc distance with time, operation of a net crustal deficit through tectonic erosion, crustal thinning and subsidence will result in trench retreat and the relative trenchward migration of the fore-arc and the oldest arc-volcanoes [Lallemand, 1995; Scholl and von Huene, 2010]. In Chile, radiometric ages of igneous rocks demonstrate the eastward migration of the locus of arc magmatism and this migration has long been interpreted as evidence of subduction erosion [Rutland, 1971; Karig, 1974; Andriessen and Reutter, 1994]. In Tonga, the Kuril Islands, and Chile, TPFRs are clearly associated with extinct volcanic arcs, each of these margins are classified as erosional plate boundaries with trench-retreat rates >3.0 km my⁻¹, and this model is likely the dominant contributing factor forming the TPFR in these regions [Clift and Vannucchi, 2004].
Figure 11. (a) Schematic diagram illustrating the elastic back slip model of strain accumulation in subduction zones [Savage, 1983]. (b) Comparison between the geometry of paired anomalies observed in the Kuril Islands and (c) North Chile, with predictions of interseismic elastic deformation (thick red line). Uplift rates are proportional to convergence rates and are stretched vertically to achieve the best fit with the form of the paired anomaly.
When the extinct arc has been translated to a location near the trench, its leading edge will begin to collapse, both gravitationally and as a result of basal tectonic erosion. Its landward margin may be relatively well preserved until it also overlies the seismogenic zone. This may explain the contrasting expression of the extinct Tonga Ridge north and south of 25.5°S (Figures 3a and 3b). In the north, the gravimetric TPFR crest is located 50–60 km from the trench, separates the contrasting morphologies of the seaward and landward ridge flanks, and marks the landward extent of shallow megathrust earthquakes. South of 25.5°S, the approximately north-south geometry (25°005°S) of the ridge places almost the entire ridge volume, as defined by residual gravimetric anomalies >30 mGal, within 60 km of the trench. Here residual gravimetric amplitudes are a factor of \( \frac{24}{24} \) lower than observed in the north. Forward-velocity models show that the region of high lower-crustal velocities (\( V_p > 6.0 \) km s\(^{-1}\)) is only 6 km thick, which is also a factor of \( \frac{24}{24} \) lower than the 11–12 km thicknesses modeled further north (Figure 3d). It thus seems likely that the lower amplitude and/or absence of residual gravimetric and bathymetric anomalies south of 25.5°S may reflect the near trench location of the extinct arc and the influence of erosive processes on the entire ridge volume.

This model may also offer an alternate explanation for the landward offset observed in some margins between the topographic and gravimetric crests of TPFR. In Tonga, these crests are approximately correlated; but in margins where the extinct arc has not yet reached the seismogenic zone, the gravimetric crest, which will also reflect extinct arc crustal structure, would be located landward of the topographic contrast (e.g., Kuril Islands, Figures 4a and 4b). As the extinct arc is drawn nearer the trench, progressive fracturing and erosion at its leading edge will reduce both its residual bathymetric and gravimetric expression. The highest residual gravimetric anomalies would be expected where the extinct arc is best preserved, which will be landward of, or approximately correlated with, the topographic crest marking the lateral extent of subduction erosion.

In the Aleutian Islands, arc volcanism has been progressive since the Early Eocene and is expressed as lines of northward migrating volcanic centers. This migration is consistent with the removal of material by subduction erosion and has resulted in the formation of a contiguous arc ridge undergoing subduction erosion at the same time contemporary volcanism is occurring on the inner (landward) ridge flank. In Cascadia, landward migration of the volcanic arc is evidenced by the progressive east-to-west age progression of

![Figure 12. Schematic diagram illustrating how trench retreat and a landward jump in the location of the volcanic arc results in the preservation of extinct volcanic arcs within fore-arc areas.](image-url)
volcanic belts [Guffanti and Weaver, 1988; Smith, 1989]. This migration is consistent with a net removal of fore-arc material, but has also been associated with clockwise rotation and northward migration of fore-arc blocks in response to the obliquity of relative plate motion [Wells, 1990; Wells et al., 1998; Wells and Simpson, 2001]. If rates of frontal accretion and/or lower crustal underplating in Cascadia exceed rates of subduction erosion such that relative trenchward migration of the arc and fore-arc is not occurring at all, the TPFR we interpret within the fore-arc may be a sufficient distance from the trench and seismogenic zone that it is no longer, or has never been, influenced by subduction related erosion and links between this TPFR crest and the megathrust source-zone may be invalid.

6.3. Fore-Arc Structure and Trench-Parallel Fore-Arc Ridges As a Constraint on Seismogenic Behavior

The residual bathymetric and gravimetric grids we have constructed are significant in illuminating interrelationships between the structure and seismogenic behavior of subduction zones. In contrast to part 1 of this study [Bassett and Watts, 2015], which focused on subducting plate structure, the observations presented in this study are significant in further demonstrating the ability of the subduction interface to shape the crustal structure of the overthrusting fore-arc [Wells et al., 2003].

In Tonga-Kermadec, the Kuril Islands, Aleutian Islands, and Chile, we have shown that a paired anomaly occupies the fore-arc and that the gravimetric crest of the TPFR is correlated with the downdip limit of thrust-earthquake epicenters, coseismic slip, and strong geodetic locking. TPFR crests may thus be interpreted as providing a viable means of estimating the downdip limit of the seismogenic zone and hence the proximity of strong shaking to densely populated coastal regions. Our observations, however, are made in margin segments where the downdip extent of coseismic slip is already well defined by recent earthquakes, and the real utility of these relations may come in regions where large-thrust earthquakes are predicted or evidenced from geodetic and paleoearthquake observations, but have not yet occurred since instrumental earthquake records began. Cascadia and the southern Hikurangi margin are two such regions, but are the links between the TPFR crest and the seismogenic zone observed elsewhere likely to be valid here?

In Cascadia for example, the ubiquitous occurrence of nonvolcanic tremor on the landward TPFR flank and correlation between the updip limit of tremor and the TPFR crest is exactly what would be predicted if the TPFR crest marks the boundary between the seismogenic zone and downdip frictional transition zone (Figure 7c). Latitudinal correlation between strike-parallel transitions in the width and amplitude of the TPFR, and the width and density of the tremor band suggests some association between fore-arc structure and megathrust properties, but it is possible both observations may reflect and be linked by lower crustal underplating. We note that the 350°C isotherm from the thermal models of Hyndman and Wang [1993, 1995] is approximately correlated with the topographic transition between the rough, steep outer fore-arc and the smooth continental shelf (Figure 7a), which may support existing and detailed studies suggesting an entirely offshore seismogenic zone [e.g., Hyndman, 2013]. This scenario suggests the TPFR is significantly offset and unaffected by subduction erosion, but also requires a separate explanation for the strong and compelling relation between the updip limit of tremor and the TPFR crest. It is clearly not possible in this study to differentiate between these scenarios, but consistency between the predicted and observed locations of tremor, and the potential implications for the width of the seismogenic zone, make further consideration of whether the TPFR crest hypothesis is applicable in Cascadia worthy of future research.

Potential constraints on coseismic slip areas are not limited to two-dimensions and in most cases, sharp along-strike changes in the seismological or geodetic character of the subduction interface can be correlated with structural transitions in either the subducting or overthrusting plates. As demonstrated by the 2007 Mw 8.1 Solomon Islands earthquake [Taylor et al., 2008], for some structures, predictive power is limited. Not all spreading centers or fracture zones will influence the seismogenic zone in the same manner and the behavior of a particular structural feature may vary from one earthquake cycle to the next. On the subducting Pacific plate the Amlia fracture zone, for example, is correlated with the eastern limit of slip in the 1986 Mw 8.0 earthquake [Houston and Engdahl, 1989; Das and Kostrov, 1990] and the eastern limit of large slip (>3 m) in the 1957 (Mw 8.7) Andreanof Islands earthquake. In this latter event, although the full rupture extent is uncertain, tsunami waveforms and a 1200 km long zone of aftershocks have been interpreted as suggesting that rupture did not arrest at the Amlia fracture zone and may have contained with low-slip (<1 m) up to ~700 km further northeast [Sykes, 1971; House et al., 1981; Johnson and Satake, 1994].
Several more prominent morphological and seismological boundaries are likely to be temporally stable, however, and we suggest the southwestern Kuril Basin margin (Bussol Graben: Figures 4a and 4b) and the Cenozoic subduction boundary associated with Rat Island and Amchitka Pass (Figures 6a and 6b) are likely examples of permanent barriers to along-strike rupture propagation. We note that both these examples, and many of the most prominent observations of along-strike segmentation, are associated with structures in the upper plate and we submit that the tectonic structure of the fore-arc is certainly as influential as subducting plate structure in modulating the segmentation and slip-behavior of subduction zones.

Since the original recognition of seismic gaps and the segmentation of subduction boundaries in the 1970s [Sykes, 1971; Kelleher et al., 1973; Kelleher et al., 1974; Kelleher and McCann, 1976; Spence, 1977], the overwhelming majority of work considering interrelationships between crustal structure and slip behavior has focused on subducting plate structure and the influence of aseismic ridges, seamounts and fracture zones on megathrust properties. In this study, we have shown that fore-arc structure is just as important in modulating the segmentation and slip-behavior of megathrusts, and has the added potential of recording the time-integrated response of the upper plate to subduction processes. Paired anomalies are one example of this response and if such structures can be robustly correlated with the mechanical behavior and physical conditions of the megathrust, they may provide a means of overcoming the limitations of short instrumental earthquake records and improving hazard assessment in regions where large subduction zone earthquakes are possible but have not occurred since instrumental records began. This twofold significance makes close consideration of overthrusting plate structure an essential component of any study aiming to fully understand the physical controls on, the manifestations of, and hazards associated with, the seismogenic behavior of subduction zones.

7. Conclusions

An ensemble averaging technique is developed to isolate and remove the long-wavelength, large-amplitude (~7.5 km and ~200 mGal), trench-normal topography and gravity fields associated with subduction zones. This procedure lowers the dynamic range, removes the steep gradients associated with the seaward and landward trench-slopes and significantly increases the resolvability of short-wavelength/lower-amplitude structure. This technique has been applied to all subduction zones on Earth and the grids of residual bathymetry and residual free-air gravity anomaly developed are available as an electronic supplement and online.

A paired anomaly consisting of a seaward trough and landward ridge is observed in residual topography and gravity anomalies of five circum-Pacific subduction zones (Figure 13). The gravimetric ridge crest is strongly correlated with the down-dip limit of coseismic slip and strong interplate coupling (Chile, Hikurangi), the up-dip limit of tremor epicenters (Cascadia), and may be interpreted as defining the boundary between the up-dip velocity-weakening and seismogenic region of the subduction interface and the down-dip frictional transition zone. Paired anomalies are interpreted as reflecting: the preferential subduction erosion of the outer-forearc and lower-crustal underplating beneath the inner forearc; the transformation of some proportion of interseismic strain into permanent geologic strain via faulting, folding or buckling of the inner-forearc; and the relative trenchward migration of arc crust and extinct volcanic arcs in regions operating with a net crustal deficit.

The pre-existing crustal structure of the overthrusting and subducting plates control almost all along-strike variations in forearc morphology and the seismogenic segmentation of megathrusts. Upper-crustal features such as faults, paleo and active plate boundaries, and transitions in crustal structure associated with ocean-continent margins often coincide with first order structural and seismological segment boundaries. Forearcs are just as important as subducting plate structure in modulating the segmentation and slip-behavior of megathrusts, and have the added potential of recording the time-integrated response of the upper-plate to subduction processes. Forearc structures, if robustly correlated with the mechanical behavior and physical conditions of the megathrust, may provide a means of overcoming the limitations of short instrumental earthquake records and improving hazard assessment in regions where large subduction zone earthquakes are possible but have not occurred since instrumental records began.
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CRUSTAL STRUCTURE AND SEISMICITY: 2. FORE-ARC STRUCTURE

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Figure 13. Schematic diagram summarizing the key spatial associations interpreted between the morphology of the fore-arc and variations in the seismogenic behavior of subduction megathrusts.


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