Aleutian Basin Oceanic Crust

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Abstract. We present two-dimensional P-wave velocity structure along two wide-angle ocean bottom seismometer profiles from the Aleutian basin in the Bering Sea. The basement here is commonly considered to be trapped oceanic crust, yet there is a change in orientation of magnetic lineations and gravity features within the basin that might reflect later processes. Line 1 extends ~225 km from southwest to northeast, while Line 2 extends ~225 km from northwest to southeast and crosses the observed change in magnetic lineation orientation. Velocities of the sediment layer increase from 2.0 km/s at the seafloor to 3.0-3.4 km/s just above basement, crustal velocities increase from 5.1-5.6 km/s at the top of basement to 7.0-7.1 km/s at the base of the crust, and upper mantle velocities are 8.1-8.2 km/s. Average sediment thickness is 3.8-3.9 km for both profiles. Crustal thickness varies from 6.2-9.6 km, with average thickness of 7.2 km on Line 1 and 8.8 km on Line 2. There is no clear change in crustal structure associated with a change in orientation of magnetic lineations and gravity features. The velocity structure is consistent with that of normal or thickened oceanic crust. The observed increase in crustal thickness from west to east is interpreted as reflecting an increase in melt supply during crustal formation.

Keywords: Aleutian basin; Bering Sea; oceanic crust; wide-angle seismic; ocean bottom seismometer

1. Introduction

The northern and northwest Pacific is rimmed by marginal basins, most of which formed by backarc spreading behind island arcs [e.g., *Karig, 1970; Packham and Falvey, 1971; Xu et al., 2014*]. In contrast, the Aleutian basin (Figure 1) is commonly modeled as oceanic crust, trapped ~55-50 Ma when subduction moved from the Beringian margin to its current location beneath the Aleutian ridge [*Scholl et al., 1975; Cooper et al., 1976b; Cooper et al., 1987; Worrall, 1991; Cooper et al., 1992*]. Little is known about the deep structure of the Aleutian basin, except that early seismic refraction profiles suggest a velocity structure consistent with oceanic crust [*Shor, 1964; Ludwig et al., 1971; Cooper et al., 1979*].

Magnetic studies initially identified north-south trending magnetic lineations in the Aleutian basin that supported the hypothesis that the basin crust was trapped oceanic crust formerly part of the Mesozoic Pacific basin [Cooper *et al.*, 1976a]. However, a later magnetic compilation mapped more complex anomalies in the basin that follow trends observed in gravity data (Figure 1); these anomalies cannot be reliably correlated with the geomagnetic reversal time scale and thus spreading rates at which this crust accreted are unknown [Cooper *et al.*, 1992]. Cooper *et al.* [1992] suggest that northeast-southwest oriented magnetic lineations are associated with early Cenozoic back-arc extension that took place after entrapment of Pacific basin oceanic crust. The proposed region of extension is located on the 100-200 km wide Vitus arch (Figure 1), which is characterized by northeast-southwest trending basement horsts and grabens [Cooper *et al.*, 1992]. An alternate model has the Vitus arch associated with a push-up structure along a strike-slip fault zone active in the middle Eocene [Chekhovich *et al.*, 2012; Chekhovich and Sheremet, 2013].

Recent seismic reflection profiles over the Vitus arch image thick, flat-lying sediments overlying a rough basement, but do not constrain crustal thickness because Moho is not clearly imaged (Figure 2). In this study we present new seismic refraction profiles that characterize the crust in the region of complex magnetic lineations on the Vitus arch (Figure 1). These new data are consistent with thickened oceanic crust, with no clear change in structure associated with the prominent change in magnetic anomaly trend on the northwestern flank of Vitus arch.

2. Seismic Data, Analysis, and Results

2.1. Data Acquisition and Processing

Cruise MGL1111 acquired seismic refraction and reflection data in the Aleutian basin during August 2011 using the R/V *Marcus G. Langseth*. The source for seismic data acquisition was a 36-airgun 4-string array towed at a depth of 9 m, with a total volume of 6600 cubic inches and a pressure of 2000 psi. Seismic refraction profiles had a shot interval of 150 m, and were recorded by ocean bottom seismometers (OBSs) positioned at a seafloor spacing 20 km on Line 1 and 5-

20 km on Line 2 (Figure 3). The OBSs included a vertical, hydrophone, and two horizontal channels, and used a sample interval of 4 ms. Coincident multichannel seismic reflection profiles were acquired with the same source array, with a shot interval of 50 m, and were recorded by an 8-km-long solid streamer with 636 channels at 12.5 m spacing towed at a depth of 9 m. Seismic reflection record length is 16 s with an original sample interval of 2 ms.

Initial processing of the OBS data included applying a correction for clock drift during deployment and inverting the direct water-wave arrivals for instrument location. Instrument location errors are small in the along-track direction but much larger in the cross-track direction; maximum instrument location errors are estimated to be 200 m. Further processing consisted of applying a Butterworth filter with a low cut of 3 Hz, a high cut of 15 Hz, and a 48 dB/octave rolloff. Seismic reflection data processing included trace regularization, bandpass filtering, velocity analysis, normal move-out correction, muting, multiple suppression, stacking, and f-k migration.

2.2. OBS Data Analysis

Representative OBS record sections are displayed in Figure 4, and show phases typical of sedimented oceanic crust. At offsets of ~7-12 km, refracted arrivals through the sediment (*Psed*) are observed as first arrivals with apparent velocities of ~2.5-3.0 km/s. The *Pg* phase (crustal refraction) exhibits a change in slope, with apparent velocities increasing from ~5-5.5 km/s to ~6.5-7 km/s, while the *Pn* phase (mantle refraction) has apparent velocities of ~8-8.5 km/s. A basement reflection (*Pbase*) is observed at near offsets, while a prominent mantle reflection (*PmP*) is observed as a triplication behind the *Pg* and *Pn* phases. The *Pn* arrival is observed to longer offsets on the Line 1 (Figure 4a-4b) than on Line 2 (Figure 4c-4d), while the opposite is true of the *PmP* phase.

Travel time picks of the *Psed*, *Pg*, *Pn*, *Pbase*, and *PmP* phases were made by hand at 150-m spacing. Both vertical and hydrophone channels were inspected, and picks were made on the channel with higher signal-to-noise ratios. For inversion, pick uncertainties of the *PmP* picks,

which are observed as secondary arrivals, were set to 100 ms. Pick uncertainties for the other phases were set to fixed values based on source-receiver offset: 20 ms for offsets <25 km, 30 ms for offsets 25-50 km, 40 ms for offsets 50-75 km, and 50 ms for offsets >75 km. There were a total of 3177 picks from 7 instruments on Line 1, and 4369 picks from 11 instruments on Line 2.

We created a starting model for the two profiles using forward modeling and the method of *Zelt and Smith* [1992], with each model consisting of water, sediment, basement, and mantle layers. We then carried out tomographic inversions using the method of *Van Avendonk et al.* [2004] which simultaneously constrains seismic velocities and layer boundaries. For the shortest path method we used a grid spacing of 0.2 km. During the inversions, horizontal grid spacing was set to 1.0 km, and vertical grid spacing increased from 0.4 km at 0 km depth (sealevel) to 0.6 km at 30 km depth. The inversions were carried out with a layer-stripping approach, where we first inverted for the sediment layer velocities, then for velocities in the sediment and basement layers and depth to the sediment-basement interface, and finally for all velocities and interface depths. The models have best resolution where raypaths are reversed: model distances 50-175 km for Line 1 and 50-200 km on Line 2; ray coverage is good from seafloor to upper mantle for the reversed regions (Figure 5).

2.3. Resolution

The final velocity models (Figure 6) fit the picked travel times (Figure 5) with a chi-square value of 1.2 (Line 1) and 1.1 (Line 2). A chi-square value of 1.0 means that the data are fit within the uncertainties; the values of 1.1 and 1.2 suggest that the data uncertainties might be slightly overestimated. We carried out a series of checkerboard tests to obtain an objective measure of model resolution using the method described by *Zelt* [1998]. Each checkerboard consisted of alternating positive and negative velocity anomalies of 5% superimposed on the final velocity model at depths below the seafloor. There were 4 independent checkerboard patterns in order to reduce the nonlinearity associated with the grid polarity and registration [Zelt, 1998]. Travel times were calculated through each checkerboard model using the real shot-receiver geometry.

Random noise was added to the calculated travel times, which were then inverted using the final velocity model as a starting model. A local semblance measure was then computed between the actual and recovered velocity perturbations; a node with a semblance of >0.7 is considered to be well resolved [Zelt, 1998]. The semblances for all 4 checkerboards for each cell size were then averaged. The semblance results for Line 1 and Line 2 for an anomaly width of 25 km and anomaly thickness of 2.5 km are shown in Figure 7. An average horizontal resolution <25 km (semblance >0.7) is obtained throughout most of the sediment and crust layers, and in the uppermost mantle, beneath the region of reversed OBS coverage on each profile (e.g., distances 50-175 km on Line 1 and 50-190 km on Line 2).

There are some tradeoffs between seismic velocities and interface depth in the inversion process; the checkerboard models have identical interface depths with the starting models, but these are perturbed in the recovered models. Differences between actual and recovered interface depths provide one estimate for interface depth resolution. In the region of reversed OBS coverage, the basement interface has depth variability as great as 0.8 km for Line 1 and 0.3 km for Line 2, while the Moho interface has depth variability as great as 1.3 km for Line 1 and 0.9 km for Line 2.

2.4. Velocity Models

Average sediment thickness is 3.8-3.9 km for both profiles (Figure 6), with some variability related to the basement topography imaged on the coincident seismic reflection profiles (Figure 2). In contrast, crustal thickness varies along the profiles, with a minimum thickness of 6.2 km near the southwest end of Line 1 (model distance 70 km, Figure 6), and a maximum thickness of 9.6 km near the southeast end of Line 2 (model distance 193 km, Figure 6). The average velocity-depth function is plotted for each profile in Figure 8. Velocities of the sediment layer increase from 2.0 km/s at the seafloor to 3.0-3.4 km/s just above basement, crustal velocities increase from 5.1-5.6 km/s at the top of basement to 7.0-7.1 km/s at the base of the crust, and upper mantle velocities are 8.1-8.2 km/s. The largest difference between the two profiles is

crustal thickness, which averages 7.2 km for Line 1 and 8.8 km for Line 2 (Figure 6 and Figure 8).

2.5. MCS Moho Reflections

Two events located at 10.2-10.3 s two-way travel time (2.5-2.9 s below basement) are tentatively interpreted as Moho on the coincident seismic reflection profiles (Figure 2). We used our velocity models to convert basement and Moho picks to depth, and calculate a crustal thickness of 8.0 km and 9.0 km for Line 1 and Line 2, respectively, at the locations of the labeled Moho events on Figure 2. These thicknesses differ by <0.15 km from those of the velocity models displayed in Figure 6, and are consistent with a greater crustal thickness for Line 2 compared to Line 1.

3. Discussion

3.1. Nature of Aleutian Basin Crust

The original hypothesis for the nature of Aleutian basin crust was Mesozoic oceanic crust, trapped when subduction moved ~55-50 Ma from the Beringian margin to its current location at the Aleutian ridge [Scholl *et al.*, 1975; Cooper *et al.*, 1976b; Cooper *et al.*, 1987; Worrall, 1991; Cooper *et al.*, 1992]. However, later magnetic analyses suggested a change in lineation orientation at the Vitus arch (Figure 1), leading to speculation that the crust in the study area could be back-arc crust formed during early Cenozoic back-arc extension [Cooper *et al.*, 1992], or a push-up structure in oceanic crust associated with a strike-slip fault zone active in the middle Eocene [Chekhovich *et al.*, 2012; Chekhovich and Sheremet, 2013]. The change in magnetic fabric orientation occurs along Line 2 between OBSs 7 and 10 (Figure 3), model distances 54-105 km.

Cooper *et al.* [1992] propose initial extension, uplift, and faulting of the Vitus arch at ~55 Ma, with extension and back-arc spreading continuing to 42 Ma and possibly as recent as 29 Ma. If this model is correct we might expect to find thinned oceanic crust near the change in magnetic orientation associated with the initial extension, and back-arc crust in the region of

northeast-southwest lineations. Back-arc crust is observed to have structure varying from thick crust (8-10 km) with lower crustal velocities of 7.2-7.4 km/s, to crust with velocities and thicknesses indistinguishable from normal oceanic crust [e.g., *Takahashi et al.*, 2009; *Arai and Dunn*, 2014]; these velocity and thickness differences are attributed to the influence of slab water enhancing melt production near the active arc [*Arai and Dunn*, 2014]. Since back-arc crust can have velocities and thicknesses similar to normal oceanic crust it is impossible to use velocity structure alone to distinguish crustal type. However, we do note that the Line 1 and Line 2 models (Figure 6) show smoothly varying velocity structure along each profile, with no clear signal associated with the change in magnetic lineation orientation at 54-105 km on Line 2. There is a single major basement step at ~60-65 km imaged in both the coincident seismic reflection profile (Figure 2b) and velocity model (Figure 6b), but it is not associated with any systematic change in crustal thickness as predicted by the *Cooper et al.* [1992] model. The basement step is associated with a zone of slightly lower velocities in the mid to lower crust, which could be consistent with a southeastward-dipping normal fault zone, although there are no reflections within the crust to identify such a through-going fault system. Within the sediment column, there is evidence of episodic gentle vertical adjustment associated with this basement step, down to the southeast (Figure 2b). It is possible that this is entirely due to differential compaction over the basement topography, with episodicity supplied by diagenetic changes at depth. We conclude that although we cannot exclude the back-arc model for crust in our study area, there are no definitive observations to support this model.

The strike-slip model suggests that the Vitus arch is a push-up structure along a Bering Sea strike-slip system [*Chekhovich et al.*, 2012; *Chekhovich and Sheremet*, 2013]. However, the reflection data do not image any near-vertical faults or flower structures along either Line 1 or Line 2 (Figure 2), nor is there an abrupt change in crustal structure (Figure 6) that might be associated with crossing a strike-slip fault. We do not think it likely that the strike-slip hypothesis for the nature of crust in our study region is correct.

The primary hypothesis for Aleutian basin formation is that it is underlain by Mesozoic oceanic crust [Scholl *et al.*, 1975; Cooper *et al.*, 1976b; Cooper *et al.*, 1987; Worrall, 1991; Cooper *et al.*, 1992]. Figure 8b compares the average velocity structure for this study with that of older Pacific crust as compiled by White *et al.* [1992]. Ocean crust typically has a high-velocity gradient upper crust ('Layer 2', velocities <6.6 km/s) overlying a low-velocity gradient lower crust ('Layer 3', velocities >6.6 km/s), a characteristic that is also observed for the Aleutian Basin (Figures 6 and 7). Velocities at the top of basement for Lines 1 and 2 are higher (5-5.5 km/s compared to 4-4.5 km/s) than that of the oceanic crust plotted in Figure 8b, but we expect this is a consequence of sediment thickness differences. The Pacific ocean crust compiled by White *et al.* [1992] has sediment thicknesses of <350 m, significantly less than the 3.8-3.9 km observed in the Aleutian basin. Increased lithostatic pressure related to a thick sediment layer will close cracks and increase seismic velocities [e.g., Birch, 1960]; for example, basement velocities of 5 km/s are observed beneath 2 km of sediment for Pacific oceanic crust in the Gulf of Alaska [Christeson *et al.*, 2010]. Hence, we argue that the basement velocities of 5-5.5 km/s are consistent with oceanic crust. A change in velocity gradient at 3.0-3.5 km below the top of basement is observed in Figure 8b, and is identified as the Layer 2/3 boundary. Line 1 and Line 2 Layer 3 velocities of 6.6-7.1 km/s and upper mantle velocities of 8.1-8.2 km/s fall within the range observed for oceanic crust [White *et al.*, 1992] (Figure 8b). The Layer 2/3 boundary on our profiles is deeper than the 2.1 ± 0.5 km depth observed globally by White *et al.* [1992], and deeper than the 1-1.5 km depth of Figure 8b for old Pacific crust, but similar observations have been made for some regions of thickened oceanic crust [Mutter and Mutter, 1993].

Globally, oceanic crust has a mean crustal thickness of 6-7 km for crust formed at full spreading rates >20 mm/yr [Chen, 1992; White *et al.*, 1992; White *et al.*, 2001]. White *et al.* [1992] estimate a mean thickness of 7.1 ± 0.8 km for oceanic crust away from fracture zones and hot spots, whereas Chen [1992] estimates a mean thickness of 6 km with greater variability observed for slow-spreading crust (20-40 mm/yr full rates). Crustal thickness on our study profiles varies from 6.2-9.6 km, with average thickness of 7.2 km on Line 1 and 8.8 km on Line

2. Overall, the Aleutian Basin crustal structure in our study region is similar to normal or thickened oceanic crust (Figure 8b).

Other seismic refraction constraints on deep crustal structure in the Aleutian Basin are from early studies in the southeast portion of the basin (red lines, Figure 1) [Shor, 1964; Ludwig *et al.*, 1971]. These data use the slope-intercept method [e.g., Raitt, 1963] to estimate a mean thicknesses of 1.5 km and 5.4 km for layer 2 and layer 3, respectively, and a total crustal thickness of 6.9 km [Shor, 1964; Ludwig *et al.*, 1971]. White *et al.* [1992] determine that the slope-intercept method typically underestimates thicknesses by 20%, which suggests that crustal thickness in the southeast Aleutian Basin is 8.3 km. These data suggest that thickened oceanic crust might be found throughout the Aleutian Basin.

3.2. Crustal Structure Variability

There are lateral changes in crustal thickness along the two profiles, with thinnest crust of ~6.2 km in the southwest portion of Line 1, and thickest crust of ~9.6 km on the southeast end of Line 2 (Figure 6). Thickness changes of 2-4 km have been observed in gravity and seismic data for slow-spreading crust formed at the Mid-Atlantic Ridge, with thicker crust at segment centers interpreted as being associated with an enhanced supply of melt [Lin *et al.*, 1990; Tolstoy *et al.*, 1993; Detrick *et al.*, 1995; Hooft *et al.*, 2000]. However, distance between thinnest and thickest crust on the Mid-Atlantic Ridge is typically 20-30 km [Lin *et al.*, 1990; Detrick *et al.*, 1995; Hooft *et al.*, 2000], whereas crustal thickness changes gradually over distances of 100-150 km on Line 1 and Line 2 (Figure 6). We therefore do not think that segment variability at slow-spreading crust is a good analog for the thickness observations along the Aleutian Basin profiles.

Alternatively, thick oceanic crust has been observed adjacent to magmatically enhanced regions. At the intermediate-spreading Galápagos Spreading Center oceanic crust thickens by 2.3 km as the Galápagos plume is approached, increasing from 5.6 km to 7.9 km; most of the thickness change occurs over a horizontal distance of ~300 km [Canales *et al.*, 2002]. Crustal thicknesses of 8-14 km are measured near the Agulhas Plateau [Parsieglia *et al.*, 2008], and of

~8-9 km (estimated from gravity data) at the Azores Platform [Detrick *et al.*, 1995]. Based on these observations, we associate the crustal thickness changes on Line 1 and Line 2 with variability in magmatic input, with melt supply increasing from west to east during crustal formation.

We note that the character of the *PmP* and *Pn* arrivals differ on Line 1 and Line 2: on Line 1 the *PmP* arrivals are observed to distances of 60-70 km, with *Pn* arrivals to offsets of 90-100 km, while for Line 2 *PmP* arrivals are observed to distances of 80-100 km, with only faint *Pn* arrivals (Figure 4). One explanation for these observations is a difference in thickness of the Moho transition zone. An abrupt change from crustal to upper mantle velocities will result in strong *PmP* arrivals over long offsets (Line 2), while a more gradual change will result in *PmP* arrivals over a shorter distance range (Line 1). Finite difference modeling also indicates that a thicker Moho transition zone can result in higher amplitude *Pn* arrivals (Line 1), depending on the exact nature of internal layering within the Moho transition [Carbonell *et al.*, 2002]. We speculate that the change in Moho character is associated with variability in melt supply, with melt extraction more efficient with increased magmatic input resulting in a sharper Moho [Lizarralde *et al.*, 2004; Conley and Dunn, 2011; Aghaei *et al.*, 2014].

4. Conclusions

1. Aleutian Basin wide-angle seismic profiles constrain a crustal structure consistent with oceanic crust. Crustal velocities increase from 5.1-5.6 km/s at the top of basement to 7.0-7.1 km/s at the base of the crust, and upper mantle velocities are 8.1-8.2 km/s. A change in velocity gradient at 3-3.5 km below the top of basement is interpreted as the Layer 2/3 boundary.

2. There is no clear change in crustal structure associated with a change in orientation of magnetic lineations and gravity features. Instead, crustal structure varies gradually along both profiles.

3. Crustal thickness varies from 6.2-9.6 km, with average thickness of 7.2 km on Line 1 and 8.8 km on Line 2. The crustal thickness values for Line 2 are greater than expected for normal

oceanic crust; we suggest that these reflect melt supply increasing from west to east during crustal formation.

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Figure Captions

Figure 1. Location maps of the study region displaying a) Seafloor bathymetry [Smith and Sandwell, 1997] and onshore topography (GTOPO30 from USGS); b) Gravity anomaly [Sandwell and Smith, 2009]. Black lines show wide-angle profiles presented in this paper, with OBS positions displayed by white circles. Red lines and black circles show seismic refraction profiles and stations of Shor [1964] and Ludwig *et al.* [1971]. White lines and orange dashed line mark magnetic lineations and Vitus arch, respectively, as defined by Cooper *et al.* [1992]. Inset displays the larger setting.

Figure 2. Two-dimensional multichannel seismic reflection profiles coincident with a) Line 1; b) Line 2. Data is plotted with a 1.0 second automatic gain control. Vertical exaggeration is ~8:1 at the basement interface. White circles show OBS locations. TWTT: two-way travel time.

Figure 3. Location map of the experiment region, with colored basemap displaying multibeam bathymetry acquired during cruise MGL1111. Black lines show wide-angle profiles presented in this paper. OBS positions are displayed by circles, with solid black circles indicating location of OBSs corresponding to the record sections of Figure 4. White lines mark magnetic lineations of Cooper *et al.* [1992]; note the west to east change in orientation.

Figure 4. Representative record sections from Line 1 and Line 2 for a) OBS 1; b) OBS 7; c) OBS 9; d) OBS 15. Data are plotted with a reduction velocity of 7 km/s, with a 0.5-s automatic gain control applied, and with travel time picks overlain. Labels identify different seismic phases used in travel time inversion. OBS locations are shown by black circles in Figure 3.

Figure 5. Observed and calculated travel time picks and calculated ray paths for a) Line 1; b) Line 2. Observed picks are shown in gray, with height equal to the pick uncertainty. Triangles indicate OBS locations. Every sixth ray path is plotted.

404 **Figure 6.** Velocity models for a) Line 1; b) Line 2. Lighter areas mark regions along profiles
405 with ray coverage. Contour interval is 0.5 km/s. White lines mark layer boundaries. Circles plot
406 OBS locations, with black circles marking instruments with record sections plotted in Figure 3.

407 **Figure 7.** Average semblance between actual and recovered velocity perturbation for four
408 checkerboard patterns with an anomaly width of 25 km and thickness of 2.5 km for a) Line 1; b)
409 Line 2. A location with a semblance of >0.7 (red contour) is considered to be well resolved.
410 Circles plot OBS locations, and blue lines mark layer boundaries.

411 **Figure 8.** a) Average velocity-depth profile beneath seafloor for Line 1 and Line 2. b) Average
412 velocity-depth profiles beneath top of basement for Line 1 and Line 2 compared to 29-140 Ma
413 Pacific oceanic crust presented by *White et al.* [1992]. Averages are computed for model
414 distances from Figure 6 of 50-175 km for Line 1 and 50-200 km for Line 2.

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8