Title: Subduction Zone Interface Structure within the Southern M_W9.2 1964 Great Alaska Earthquake Asperity: Constraints from Receiver Functions Across a Spatially Dense Node Array

Authors: Evans A. Onyango[1*], Lindsay L. Worthington[1], Brandon Schmandt[1], Geoffrey Abers[2]

***Corresponding Author:**

Evans A. Onyango, eaonyango@unm.edu

Department of Earth and Planetary Sciences, Northrop Hall, 221 Yale Blvd NE, University of New Mexico, Albuquerque, New Mexico 87131, USA.

Affiliations:

[1]Department of Earth and Planetary Sciences, Northrop Hall, 221 Yale Blvd NE, University of New Mexico, Albuquerque, New Mexico 87131, USA.

[2]Department of Earth and Atmospheric Sciences, Cornell University, 112 Hollister Drive, Ithaca, NY, 14853-1504, USA.

Key Points:

- We present receiver function imaging from a dense three-component nodal array deployment on Kodiak Island above the subducting Pacific Plate.
- A clear slab Moho conversion is found but, in contrast to the Kenai Peninsula, there is no coherent low-velocity layer atop the slab.
- The 1964 Great Alaska Earthquake ruptured across structural segments with different plate interface properties.

Abstract

We conduct a high-resolution teleseismic receiver function investigation of the subducting plate interface within the Alaskan forearc beneath Kodiak Island using data collected as part of the Alaska Amphibious Community Seismic Experiment in 2019. The Kodiak node array consisted of 398 nodal geophones deployed at ~200-300 m spacing on northeastern Kodiak Island within the southern asperity of the 1964 Mw9.2 Great Alaska earthquake. Receiver function images at frequencies of 1.2 and 2.4 Hz show a coherent, slightly dipping velocity increase at ~30-40 km depth consistent with the expected slab Moho. In contrast to studies within the northern asperity of the 1964 rupture, we find no evidence for a prominent low-velocity layer above the slab Moho thick enough to be resolved by upgoing P-to-S conversions. These results support evidence from seismicity and geodetic strain suggesting that the 1964 rupture connected northern (Kenai) and southern (Kodiak) asperities with different plate interface properties.

Plain Language Summary

We use 398 portable seismometers that were deployed as part of the Alaska Amphibious Community Seismic Experiment to image the boundary between the subducting Pacific plate and the base of the North American plate. The seismometers, spaced ~200-300 m apart, were stationed on Kodiak Island in 2019 within the southern rupture area of the 1964 Mw9.2 Great Alaska earthquake. We analyze conversions from compressional to shear waves from distant earthquakes to understand the conditions of the plate interface. Our results show a dipping velocity increase at ~30-40 km depth at the expected location of the Pacific slab crust-mantle boundary. In contrast to prior results from the northern 1964 rupture zone, we do not find a low-velocity layer on the subducting plate. Our results indicate that the 1964 rupture connected segments of the Alaskan subduction zone with different plate interface properties.

1 Introduction

Understanding plate interface structure and subduction geometries can illuminate slip mechanisms, earthquake rupture behavior and shallow subduction zone processes. Because most global forearc regions are submerged, they are commonly studied via marine seismic methods, which, thus far, precludes dense-array natural source seismic imaging. Therefore, well-exposed forearcs such as Kodiak Island provide rare opportunities to study subduction zone and plate interface structure within the shallow forearc using a dense seismic array. Here, we use three-component node array data acquired in 2019 across northeastern Kodiak Island as part of the Alaska Amphibious Community Seismic Experiment (AACSE) to compute Ps teleseismic receiver functions (RFs) to better understand the nature of the plate interface in the rupture area of the 1964 Mw9.2 Great Alaska earthquake.

The Alaska-Aleutian subduction zone has hosted more M>8 earthquakes than any other system globally and offers opportunities to explore relationships between megathrust slip phenomena, seismicity, deformation and forearc structure. The Kodiak node array (Fig. 1a-c) lies within the southern rupture area of the 1964 Mw9.2 Great Alaska earthquake, the second largest earthquake ever recorded (Kanamori, 1977, Fig. 1a). Coseismic slip and ground shaking from this event created damage across a 600-800 km section of the Alaskan margin and triggered local and far-field tsunami. Previous work investigating static deformation, seismic waves, and tsunami propagation from this event revealed two major coseismic slip asperities: the Kenai asperity in the north and the Kodiak asperity in the south (Christensen & Beck, 1994; Ichinose et al., 2007; Johnson et al., 1996; Suito & Freymueller, 2009; Fig. 1a). Differences in coseismic slip

(Johnson et al., 1996), major earthquake recurrence interval (Wesson et al., 1999; Nishenko and Jacob, 1990), locking (Zweck et al., 2002; Li and Freymueller, 2018), subduction geometries (Christeson et al., 2010) and sediment input (Worthington et al., 2012; Reece et al., 2011) between these two regions suggest major differences in subduction and interface properties within south-central Alaska.

2 Geologic Background and Previous Geophysical Studies of the 1964 Rupture Area

Kodiak Island (*Qikertaq* in Alutiiq) is part of an archipelago that represents an exposed section of the Mesozoic-Tertiary Alaska-Aleutian accretionary complex uplifted either via duplex accretion and underplating (Sample & Fisher, 1986), out-of-sequence splay faulting (e.g., Rowe et al., 2009), or a combination of these processes. The surface exposures consist of Jurassic to Eocene formations bounded by NW-dipping and NE-striking thrusts (Wilson et al., 2015; Fig. 1b). The thrust-bounded units get progressively younger towards the southeast, approaching the current subduction trench offshore (Fig. 1b). Potentially active Quaternary fault systems include the Albatross Bank, Kodiak Shelf and Narrow Cape fault zones (Figs. 1b and 1c). Paleocene granitic intrusions (~58-50 Ma) from ridge subduction (Ayuso et al., 2009; Farris et al., 2006; Fig. 1b) form the mountainous spine of the island interior. In the duplex accretion and underplating scenario for Kodiak Island formation and deformation, a stacked section of marine sediments builds up near the subduction decollement, forming a series of flat-ramp-flat geometries of imbricated material at depth within the overriding plate (Sample & Fisher, 1986; Fig. 1d (i)). The build-up of the underthrust material causes the accretionary prism to grow vertically, with minimal fault penetration or deformation within the overlying sediments. In the splay fault model (Fig. 1d (ii)), the island was uplifted due to deformation on one or several seaward-vergent thrusts possibly rooted at the megathrust.

Prior to our study, the 2007-2008 Multidisciplinary Observations of Onshore Subduction (MOOS; J. Li et al., 2013; Fig. 1a) measured structure and seismicity beneath the Kenai Peninsula in the northern 1964 rupture zone. The MOOS experiment included 34 broadband seismometers deployed at 10-15 km station spacing. Major results include RF imaging showing a 3-5 km-thick low velocity zone (LVZ) sandwiched between the overriding North American plate and the subducting Yakutat microplate (Y. Kim et al., 2014). This low-velocity zone suggests the presence of subducting sediments and/or the presence of fluids within or below the plate interface. Imaging via autocorrelation of P-wave coda from local earthquakes replicates these results and further suggests that S-wave velocity within this zone decreases with depth (D. Kim et al., 2019).

A more recent study of the subducting crust beneath southcentral Alaska suggests that the LVZ extends far beyond the location of the MOOS array. In their scattered-wave imaging of the subduction zone beneath southcentral Alaska, Mann et al. (2022) analyzed seismic data recorded by 218 broadband seismometers across southcentral Alaska. Using data from the Wrangell Volcanism and Lithospheric Fate (WVLF; Fig. 1a) array, the Broadband Experiment Across the Alaska Range (BEAAR; Fig 1a) array, the Transportable Array (TA) and the MOOS array, they found that the LVZ covers > 450 km of the subducting Yakutat terrane (Mann et al., 2022). Our study tests whether these features extend southward, controlling structure beneath northeast Kodiak Island.

3 Data and Methods

3.1 The AACSE

The AACSE took place in 2018-2019 between Kodiak Island and Sanak Island (Abers et al., 2019; Barcheck et al., 2020; Fig. 1a). All experiment data is publicly available and was open immediately upon completion of quality assurance, control and archiving. The AACSE included 75 broadband ocean-bottom seismometers (OBS), 30 broadband land seismometers, several dozen additional nearby permanent and EarthScope Transportable Array seismometers, complementary strong motion sensors and absolute- and differential-pressure gauges, and >3,000 km of active source wide-angle refraction profiles collected by the R/V *Marcus G Langseth* (Barcheck et al., 2020). The Kodiak node array was deployed in 2019 as a supplement to the larger AACSE. The array consisted of 398 Fairfield autonomous node sensors (from PASSCAL and University of Utah) with 3-component 5-Hz geophones deployed along a ~50 km road network centered on the city of Kodiak (Figs. 1b and 1c). Sensors were deployed at ~200-300 m station spacing over the course of 6 days (May 18-24) and recovered over 3 days (June 19-21). The full nodal array was operational for 25 days (May 25 – June 18). All continuous waveform data from the node array are available in PH5 format via IRIS Data Services (network code 8J from 2019).

3.2 Receiver Function Processing

Previous work shows that the autonomous three-component 5-Hz geophones used in this study can yield high quality RFs comparable with co-located broadband seismometers (Liu et al., 2018; Ward et al., 2018; Ward & Lin, 2017). Like those earlier studies, our short deployment period limited the number of teleseismic events for RF calculation. Out of 52 teleseismic events >Mw 5.0 occurring within the 30°- 90° search radius, we retained 7 events (Table S1; Fig. S1(a) and S1(b)) that met the selection criteria: (1) a magnitude >5.5, (2) a 30° – 90° epicentral distance from the center of the array, and (3) a signal-to-noise ratio (SNR)>3 and an identifiable incident P wave across the array (Figure S1c).

Prior to calculating RFs, we windowed the seismograms from 15 s before to 75 s after the theoretical P arrival. Next, we decimated the waveforms to 50 samples per second using a finite impulse response filter to prevent aliasing. We then removed the mean and the trend and applied a Hanning taper. Finally, we removed the instrument response from the nodal geophones (5 Hz corner frequency). We followed the above steps as outlined by Ward et al. (2018). We then filtered the resulting time series using a bandpass of 0.2 – 2.0 Hz. To groundtruth our waveform processing workflow, we retrieved waveforms for the selected 7 events recorded by AACSE broadband stations deployed within the node array footprint (Z. Li et al., 2020), performed the same pre-processing procedure, and compared the resultant broadband waveforms with the pre-processed nodal time series (Fig. S2).

After preprocessing, we culled additional noisy signals by applying a SNR-based noise reduction procedure which eliminated traces with SNR< 2.0 on the vertical component or SNR < 1.25 on the north component. Then we rotated from the station ZNE (vertical, north, east) coordinate system to the earthquake ZRT (vertical, radial, transverse) system. To compute the RFs for each event, we deconvolved the radial component seismograms with vertical component seismograms at each station using the time-domain iterative deconvolution method (Ligorria & Ammon, 1999) with a Gaussian filter parameter of 2.5 (~1.2 Hz) and 5.0 (~2.4 Hz). All analysis was performed via Python using the open-source rf software package (Eulenfeld, 2020).

Before stacking the RFs, we applied a Ps phase moveout correction using the iasp91 (Kennett & Engdahl, 1991) model and calculated piercing points. We set the piercing point depth at 20 km based on estimates of slab depth (20 – 27 km) beneath the study area from the Slab2.0 model (Hayes et al., 2018), created equal profile boxes along the array (Fig. S3), and then stacked the receiver functions by common conversion points (Fig. 2). Both the stacked 1.2 Hz and 2.4 Hz RFs were converted to depth (Fig. 2b and 2c) using the rf software and the iasp91 velocity model (Kennett & Engdahl, 1991).

3.3 1-D Synthetic Modeling

To aid our interpretation, we produced synthetic RFs (assuming a ray parameter of 0.05 s/km) that tested three simple velocity-density models of the structure below Kodiak Island. Our primary goal was to evaluate resolution of hypothetical structures near the top of the subducting oceanic crust and compare with previous results from the northern 1964 rupture area. To better account for the RF variability across the Kodiak profile, we selected groups of RFs from three different sections (6-km bins, centered at 10, 22 and 32 km distance along the profile) which showed good signal-to-noise ratios (Fig. 2c) and calculate uncertainties by bootstrap resampling the RFs in each bin before producing the bins' unweighted stacks. We then used the position of the slab Moho Ps arrival on the resultant stacked traces to define the slab Moho depth of the models (Figs. 3a-c).

Model 1 (Table S2; Fig. 3a) is a four-layer model based on the Kim et al. (2014) Kenai Peninsula model beneath the Kenai asperity. The model consists of a featureless upper crust, a 3 km-thick LVZ at the plate interface and an 8 km-thick oceanic crust. To construct model 2 (Table S2; Fig. 3b), we removed the 3-km-thick LVZ from model 1 and calculated synthetics using just the featureless upper crust and the 8 km-thick oceanic crust. For Model 3 (Table S2; Fig. 3c), we eliminated the 3-km LVZ and the top of the oceanic crust resulting in a simple two-layer model with one increase in velocity at the slab Moho depth.

4 Results

4.1 Receiver Function Imaging

Our final common conversion point stack produces a NW-SE-trending, approximately trench-perpendicular profile that samples a ~50 km segment of the Alaska subduction forearc up to 80 km deep (Fig. 2). Both the stacked 1.2 Hz (Fig. 2a) and the stacked 2.4 Hz images (Fig. 2c) show a coherent, SE to NW dipping positive conversion at ~ 30-40 km depth consistent with the expected slab Moho depth from previous studies. For reference, we plotted earthquakes from the AACSE catalog (Ruppert et al., 2021a; Ruppert et al., 2021b) beneath the study area (57.40-58.0 N, 152.083-152.75 W) which are within one standard deviation of the mean hypocentral depth of 24.96 km on our CCP images (black dots in Fig. 2b and 2d). We also plotted the top of the slab depth from Hayes et al. (2018) and inferred the slab Moho depth assuming an 8-km thick oceanic crust (blue and red dashed lines in Fig. 2b and 2d). We do not observe a negative top-of-slab conversion above the positive slab Moho conversion.

We observe intermittent segments of shallow (above ~10 km depth) positive conversions across the length of the profile in our high frequency (2.4 Hz) stacked image (Fig. 2d). One such horizon at ~ 5 km depth extends from about ~8-12 km along the profile, and another beneath Kalsin Bay at ~7 km depth extends from 28-35 km along the profile. Since the depths of these early arrivals vary along the line, the features generating them are likely laterally discontinuous. A mixture of the resultant reverberations and other possible primary arrivals could explain the chaotic character of the traces between ~ 5 km and 35 km depths. Increasing the Gaussian value to 10 (~4.8 Hz) sharpened the amplitudes of coherent arrivals and introduced noise that degraded prominent features such as the slab Moho Ps (Fig. S4(b)).

4.2 Synthetic Modeling Results

Since we were only modeling the features at slab depth and only considering the upgoing Ps conversion, we calculated correlation coefficients of the predicted and the observed waveforms from 2 s after the P arrival to 10 seconds after the P arrival. Model 1 (Fig. 3a) produced the worst fitting synthetics of all three models (average correlation coefficient of 0.003). Model 2 (Fig. 3b) is a better fit compared to the first model (average correlation coefficient of 0.54). Model 3 (Fig. 3c), the simple two-layer model with an increase in velocity at the slab Moho depth, is the best fitting model with an average correlation coefficient of 0.59. The results suggest that the Vp, Vs and density above the slab Moho must be similar to obtain an optimal fit to the observed data. In other words, introducing additional features in the model above the Moho, even an oceanic crust, creates synthetics that poorly match the observational data.

5 Discussion

5.1 Absence of Oceanic Crust Arrival

In subduction zone environments, RFs are commonly used to investigate plate interface structure since the method exploits the conversion of incident P waves from a teleseismic event to S waves at significant seismic-velocity discontinuities. RFs have identified LVZs along the plate interfaces in subduction zones globally as negative amplitude pulses atop positive amplitude pulses at slab depth (Bostock, 2013; Audet & Kim, 2016). This dipole character has been observed in the Japan (Kawakatsu & Watada, 2007; Akuhara et al., 2017), Cascadia (Janiszewski & Abers, 2015; Ward et al., 2018), Costa Rica (Audet & Schwartz, 2013), Mariana (Tibi et al., 2008), Alaska (Ferris et al., 2003), and the central Mexico (Pérez-Campos et al., 2008; Y. Kim et al., 2012) subduction zones. Depending on how far down dip the study area is located, the negative pulse is typically interpreted as hydrated oceanic crust or mantle hydrated by fluid expelled from the subducting slab due to the low S-wave velocities observed, while the positive amplitude pulse is generally the slab Moho. In Cascadia, Janiszewski and Abers (2015) interpreted the LVZ as metamorphosed sediments, while Bangs et al. (2009) interpreted the LVZ in Nankai as high porosity underthrust sediment. In the northern 1964 segment, Y. Kim et al. (2014) also observed this typical negative-to-positive character, attributing the negative arrivals to an LVZ of subducted marine sediments along the plate interface. Neither our observed nor the preferred synthetic RFs (Figs 2 and 3) feature the negative-positive dipole character observed within the northern 1964 asperity, highlighting a significant difference in RF character within the 1964 rupture area. The lack of major arrivals before the positive slab Moho phase suggests three possibilities for subsurface structure: (1) The presence of metasediments at the plate interface with seismic properties similar to the base of the upper plate and top of the subducting slab; (2) there may be a sedimentary layer too thin to be resolved by 1.2 – 2.4 Hz RFs; and (3) there may be no sediments at the plate interface after having been scraped off at the trench during subduction.

We note some negative arrivals above the slab Moho at both ends of our profiles (Fig. 2) that may suggest limited areas of low velocity at the interface, perhaps sediments. Plate interface material is commonly inferred from trench sediment input to the subduction zone (Morgan, 2004; Underwood, 2007). Approximately 2 km of pelagic and Surveyor Fan sediment (von Huene et al., 2012; Reece et al., 2011; Fig. 1a) comprise the subduction input near Kodiak. It is therefore unlikely that the plate interface beneath Kodiak is devoid of sediments. We suggest that the subduction zone environment has altered the properties of most of the subducted sediment at the interface, thus suppressing the velocity and density contrast between the sediment and the surrounding rock across most of the interface. There is ample evidence from magnetotelluric (Heise et al., 2012), laboratory (Miller et al., 2021) and field studies of exhumed metasedimentary rocks from subduction zone forearcs (Rowe et al., 2009; Rowe et al., 2013) pointing to instances of hundreds of meters of metamorphosed sediments lining the plate interface. It is likely that the metasedimentary rocks exhumed on Kodiak Island are close enough in seismic properties (e.g., Miller et al., 2021) to the Pacific crust that there is no significant discontinuity at the interface to resolve with Ps RFs. Therefore, the absence of a well-defined LVZ channel at the plate interface beneath our study area does not necessarily mean an absence of subducted sediment. In their study of P- and S-wave velocities of exhumed Kodiak metasediments, Miller et al., (2021) reported anisotropy of ~8-28% in Vp and ~6.5-8% in Vs, with lower wave speeds perpendicular to the rocks' dominant fabric. This suggests an absence of foliation or obliquely foliated rocks conducive for higher wavespeeds beneath our study area.

While the Ps RFs presented here use relatively high frequencies for teleseismic imaging (1.2 – 2.4 Hz), there may be coherent structural layers that are too thin to be resolved. For example, using controlled source seismic reflection data, J. Li et al. (2018) estimated a thin 600-900 m low-velocity channel at shallower (~8-10 km) depths along the plate interface south of Kodiak Island inside the 1938 Mw 8.2 Semidi rupture zone. Our synthetic test of 2.4 Hz Ps RFs showed that although we can detect a 750 m thick LVZ, it is very close to the limit of our resolution (Fig. S4(a)). RFs recovered from a 500 m thick LVZ fall within 2 standard deviations (2σ) of the field data (Fig. S4(a)) suggesting that, if an LVZ exists beneath our study area, it is less than 500 m thick. We also tested using higher frequency observations, 4.8 Hz, but the signal-to-noise ratio of teleseismic sources decreases and the prominent velocity increase interpreted as the slab Moho is only resolved sporadically across the array (Fig. S4(b)). In areas where potential slab Moho arrivals are observed in the 4.8 Hz RF image, we still do not find evidence for an overlying LVZ (Fig. S4(b)). Thus, we cannot rule out a thin LVZ (<500 m) but we can be confident that a thicker LVZ (~3-5 km) like that imaged by Kim et al. (2014) in the Kenai asperity would be resolvable if it existed beneath our study area. Mann et al. (2022) used scattered P and S coda of teleseismic P waves to successfully image a continuous ~7-km thick low-velocity layer lining the top of the subducted Yakutat crust. While we see reverberations in sections of our profile, their quality is too low to allow for interpretation. The short deployment window (~25 days) and the limited back-azimuth distribution of the events used in this study limits the usefulness of later arrivals.

5.2 Evidence of Rupture Across a Heterogeneous Plate Interface

The simple plate interface structure beneath Kodiak compared to the more complicated plate interface structure beneath the Kenai Peninsula supports other evidence that the 1964 earthquake ruptured multiple segments across distinctive asperities. During the 1964 event, the northern Kenai asperity slipped an average of 18 m, while Kodiak slipped an average of ~10 m (Johnson et al., 1996). Major earthquakes in the Kenai area have a recurrence interval of 700-800 years (Wesson et al., 1999) and the plate interface is strongly locked (Zweck et al., 2002). In Kodiak, the major earthquake recurrence interval is 60 years (Nishenko & Jacob, 1990) and, while the southern end of the Kodiak interface appears strongly locked (S. Li & Freymueller, 2018), locking decreases to the north. Subduction geometry in the Kenai segment is controlled by subduction of the Yakutat microplate, a thick, buoyant oceanic plateau (Christeson et al., 2010) and a thick, subducting sediment package (Y. Kim et al., 2014; Worthington et al., 2012). Beneath Kenai, the plate interface dips shallowly at ~3-4 degrees. In Kodiak, the Pacific plate subducts beneath North America at ~8 degrees, and incoming plate structure includes ~2.5 km-thick sediments from the distal Surveyor Fan (Reece et al., 2011) and the Kodiak-Bowie seamount chain (Fig. 1a).

Large megathrust earthquakes at other subduction zones, such as the 1700 M 9.0 Cascadia (Wang et al., 2013), 2011 M 9.0 Tohoku-Oki (Wei et al., 2012), 2004 M 9.2 Sumatra (Chlieh et al., 2007), and the 2011 M 8.8 in Chile (Lorito et al., 2011) events encompassed patches of slip rates different from the ambient slip rates within their rupture extents. The ubiquity of heterogeneous coseismic slip during large earthquakes further illustrates that the Great Alaska earthquake entraining multiple major segments during rupture is not unique to the Alaska subduction zone.

5.3 Implications for Rupture Dynamics

Since Ruff (1989) observed that large earthquakes occurred in subduction segments with large sediment inputs, a growing number of studies have linked the occurrence of great megathrust earthquakes with subducted sediment thickness $\geq$ 1.2 km (e.g., Scholl et al., 2015; Seno, 2017). Many of these studies argue that, depending on the quantity and mineralogical properties of the subducted sediments, a sedimentary layer can level inter-plate relief facilitating rupture propagation over long distances (Ruff, 1989). Numerical modeling (e.g., Brizzi et al., 2020) suggests that a total absence of sediments at the plate interface would yield significantly smaller earthquakes (M<8.5) compared to interfaces with just a 1.5 km thick sediment layer. The 2011 M 9 Tohoku-Oki provides an example of a great earthquake that occurred with < 1 km thick sediment layer at the interface (Heuret et al., 2012). We did not find any recorded great megathrust earthquakes occurring at subduction zones with no trench sediment input.

In their study of Kodiak region seismicity between 1964 and 2001, Doser et al. (2002) found that, while most earthquakes occur within the downgoing plate, several events beneath southern Kodiak Island have depths and thrust faulting mechanisms consistent with seismicity on the interface, suggesting the existence of subducted topographic features such as seamounts from the Kodiak-Bowie chain (Fig. 1a) beneath Kodiak that have not been smoothed with a thick sediment padding. Detailed seismicity studies on the Kenai Peninsula using the MOOS array show a well-defined seismic zone concentrated in the down-going plate, just below the plate boundary, that parallels the megathrust zone and is dominated by normal faulting mechanisms (J. Li et al., 2013). In contrast to observations in the Kodiak region, active thrusting and seismicity on the plate interface itself was absent (J. Li et al., 2013), possibly related to thick sediment subduction between the North American and Yakutat plates smoothening localized asperities and favoring uniform rupture in great earthquakes but not small heterogenous ruptures.

6 Conclusions

We analyzed teleseismic P waves from 398 autonomous three-component 5-Hz nodal geophones on Kodiak Island as part of the Alaska Amphibious Community Seismic Experiment. We calculated RFs with a Gaussian value of 2.5 (~1.2 Hz) and a Gaussian value of 5.0 (~2.4 Hz). The lower frequency (1.2 Hz) RFs were comparable to RFs from near-collocated broadband seismometers, and the higher frequency (2.4 Hz) RFs image produced more details. In both low and high frequency images, there is a coherent, SE to NW dipping positive phase at the expected slab Moho depth but no observable negative arrival to indicate phase conversions at the oceanic crust. To help explain the observed RFs, we calculated synthetic RFs from 1-D models. These synthetic tests suggest that the overriding forearc material and Pacific oceanic crust have nearly identical seismic velocities and densities. We conclude that the 1964 Great Alaska Earthquake ruptured beyond the extent of the low-velocity shear zone observed in the Kenai asperity into a structural setting beneath Kodiak Island with little seismic contrast across the plate boundary interface.

Data and Resources:

The nodal seismic data used in this study are available from the IRIS DMC (dmc.iris.edu) under the network code 8J (doi: 10.7914/SN/8J_2019). The IRIS DMC is supported by the National Science Foundation under Cooperative Support Agreement EAR-1851048. We obtained digital elevation data for Figures 1a and 1c from the GEBCO Compilation Group (doi:10.5285/a29c5465-b138-234d-e053-6c86abc040b9, last accessed August 2021). Geologic map data for Figure 1b and 1c was obtained from the USGS Scientific Investigations map 3340 (https://pubs.er.usgs.gov/publication/sim3340, last accessed August 2021).

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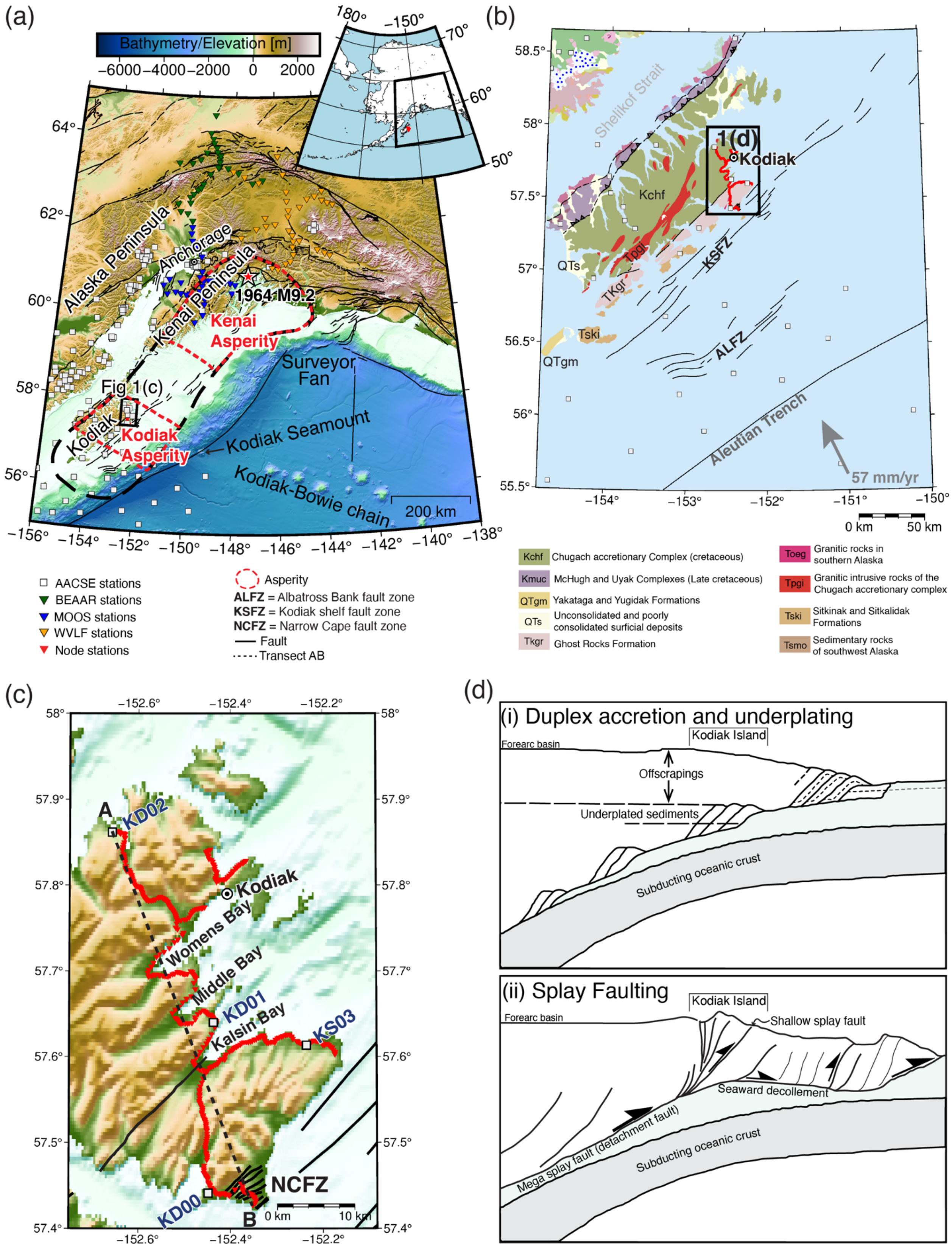
(a)
Bathymetry/Elevation [m]
−6000 −4000 −2000 0 2000
Alaska Peninsula
Anchorage
Kenai Peninsula
1964 M9.2
Kenai Asperity
Surveyor Fan
Fig 1(c)
Kodiak
Kodiak Asperity
Kodiak Seamount
Kodiak-Bowie chain
200 km
AACSE stations
BEAAR stations
MOOS stations
WVLF stations
Node stations
Asperity
ALFZ = Albatross Bank fault zone
KSFZ = Kodiak shelf fault zone
NCFZ = Narrow Cape fault zone
Fault
Transect AB
(b)
Shelikof Strait
1(d)
Kodiak
Kchf
Tpgi
QTs
TKgr
KSFZ
Tski
QTgm
ALFZ
Aleutian Trench
57 mm/yr
0 km 50 km
Kchf Chugach accretionary Complex (cretaceous)
Kmuc McHugh and Uyak Complexes (Late cretaceous)
QTgm Yakataga and Yugidak Formations
QTs Unconsolidated and poorly consolidated surficial deposits
Tkgr Ghost Rocks Formation
Toeg Granitic rocks in southern Alaska
Tpgi Granitic intrusive rocks of the Chugach accretionary complex
Tski Sitkinak and Sitkalidak Formations
Tsmo Sedimentary rocks of southwest Alaska
(c)
A
KD02
Kodiak
Womens Bay
Middle Bay
KD01
Kalsin Bay
KS03
NCFZ
KD00
B
0 km 10 km
(d)
(i) Duplex accretion and underplating
Kodiak Island
Forearc basin
Offscrapings
Underplated sediments
Subducting oceanic crust
(ii) Splay Faulting
Kodiak Island
Forearc basin
Shallow splay fault
Seaward decollement
Mega splay fault (detachment fault)
Subducting oceanic crust

Figure 1. (a) Shaded topographic map and faults of southern Alaska and the Kodiak Islands region. MOOS array (blue triangles), BEAAR array (green triangles), WVLF array (Orange triangles). **(b)** Geology map of the Kodiak Islands region. Refer to Wilson et al., (2015) for geologic unit explanation. **(c)** Shaded topographic map of the study area. **(d)** Schematic diagrams depicting scenarios for Kodiak Island formation and deformation. (i) Modified from Paterson and Sample (1988) illustrates the duplex accretion and underplating scenario. (ii) Modified from Tsuji et al., (2014) illustrates the splay faulting scenario.

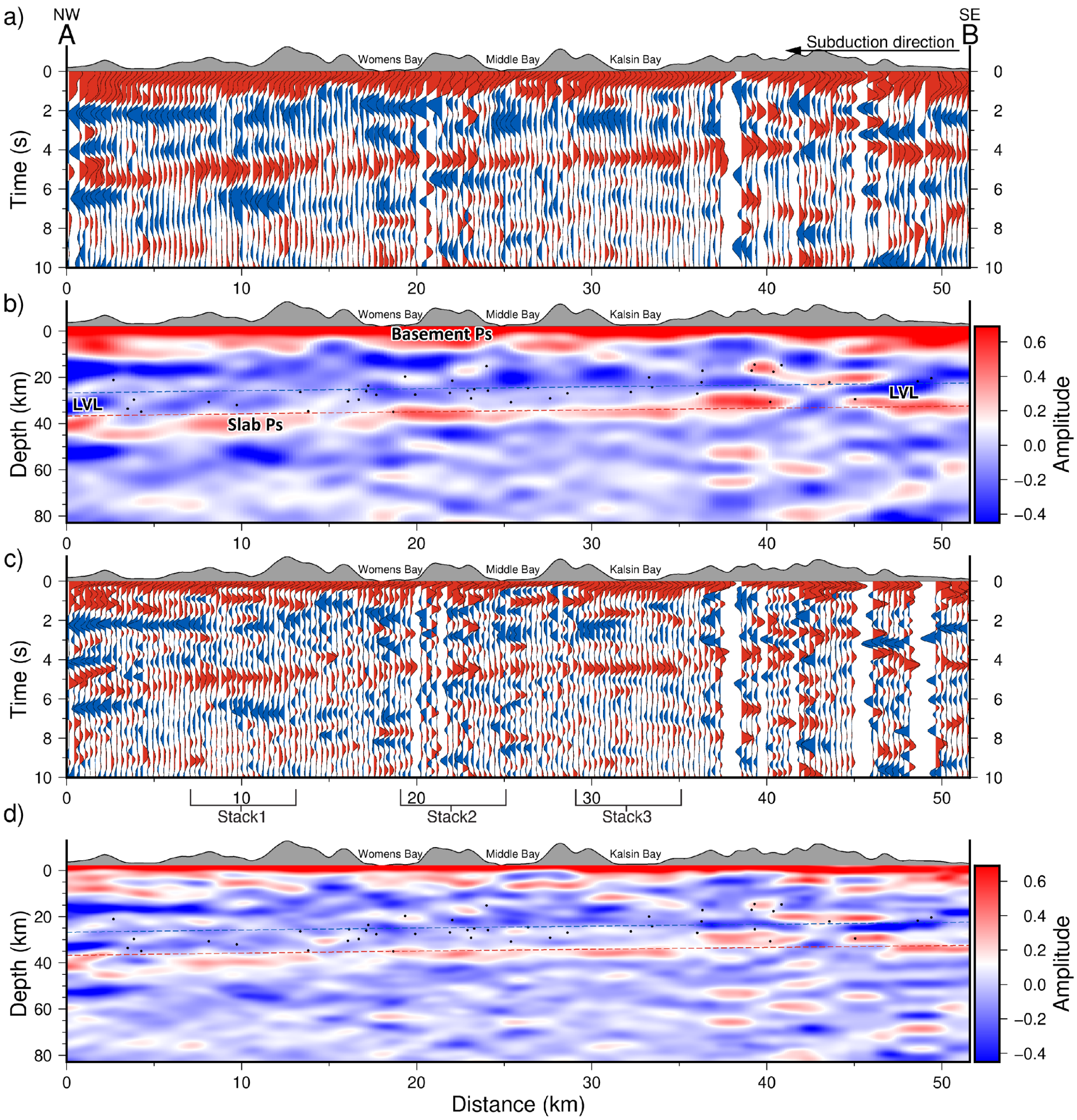

Figure 2. (a) Stacked radial receiver functions with a Gaussian value of 2.5 (~ 1.2 Hz). **(b)** ~1.2 Hz CCP image for transect A-B. Note the clear lack of a low velocity channel at the plate interface (Red = Positive, Blue = Negative). For reference, earthquakes from the AACSE catalog (black dots) and the top-of-slab depth from Hayes et al. (2018) is plotted as blue dashes and inferred Moho surface assuming an 8-km thick oceanic crust is plotted as red dashes. Vertical

exaggeration = 0.135. **(c)** Stacked radial receiver functions with a Gaussian value of 5.0 (~2.4 Hz). Stack1, Stack2 and Stack3 show the locations of the receiver functions stacked and plotted in Figure 3 to compare with synthetics. **(d)** ~2.4 Hz CCP image for transect A-B. Note the clear lack of a low velocity channel at the plate interface.

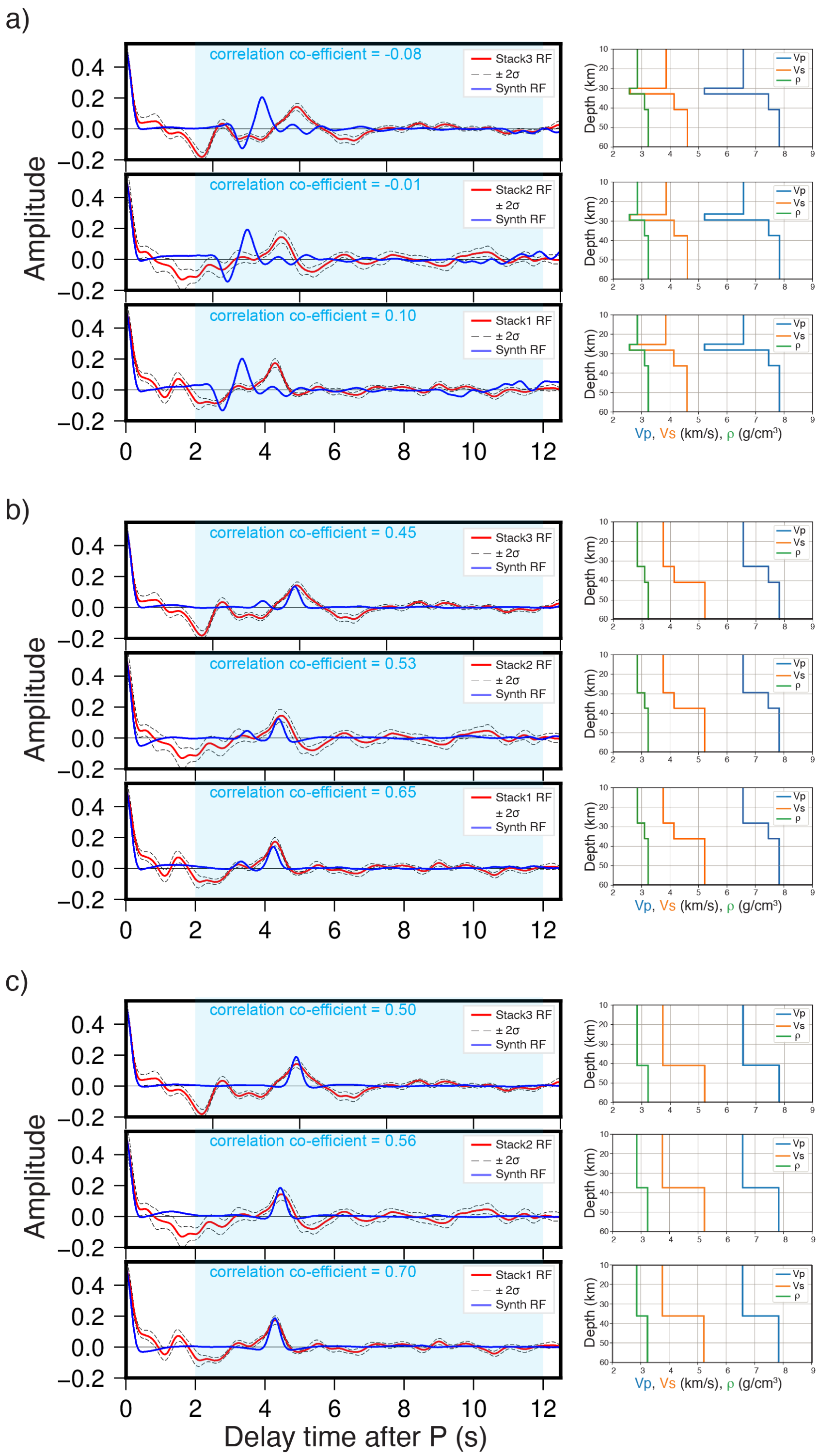

Figure 3. Each set of 3 plots represents synthetic modeling results (black dashed lines) overlaid on stacked field RFs (red lines) centered at 10 km (top), 22 km (middle) and 3 km (bottom), field RF uncertainties are plotted as black dashed lines. The right column contains the velocity models

used to calculate the synthetic RFs on the left. **(a)** Model 1 is analogous to the Kenai observations by Y. Kim et al., (2014). **(b)** Model 2 has no LVZ above the subduction slab. **(c)** Model 3 is the best-fitting model, it only contains the slab Moho.