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Investigating Seismic Deformation Associated with the Aging and Subduction of Oceanic Lithosphere Using Teleseismic Earthquake Relocation and InSAR Techniques

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INVESTIGATING SEISMIC DEFORMATION ASSOCIATED WITH THE AGING AND SUBDUCTION OF OCEANIC LITHOSPHERE USING TELESEISMIC EARTHQUAKE RELOCATION AND INSAR TECHNIQUES

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INVESTIGATING SEISMIC DEFORMATION ASSOCIATED WITH THE AGING AND SUBDUCTION OF OCEANIC LITHOSPHERE USING TELESEISMIC EARTHQUAKE RELOCATION AND INSAR TECHNIQUES

A Dissertation Presented to the Graduate Faculty of Dedman College Southern Methodist University in Partial Fulfillment of the Requirements for the degree of Doctor of Philosophy with a Major in Geophysics by Kevin Kwong B.S., Geology, California State Polytechnic University, Pomona M.S., Geophysics, University of Utah May 18, 2019
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Global seismic monitoring and the resulting teleseismic earthquake catalogs first adopted in the early 1960’s from the Bulletin of the International Seismological Centre (ISC) has become invaluable for mapping and delineating the large scale (1000s of kms) interior structures of plates subducting at convergent plate margins around the world. Based on seismicity and deformation patterns, the evolution and subduction of oceanic lithosphere can be divided into three domains: 1) the incoming oceanic lithosphere plate, 2) the megathrust plate boundary, and 3) the down-going slab. Well-located hypocenters illuminate fault complexity in space and time and verify details of surface deformation within or associated with the subduction system. This dissertation presents high-resolution earthquake catalogs derived using teleseismic double-difference (DD) relocation to best resolve earthquakes in absolute space, while providing relative locations for more detail fault studies. In the Wharton Basin, a new earthquake catalog allows more detailed examination of complex oceanic intraplate faulting and slip processes within un-subducted lithosphere (Domain 1). A revised earthquake catalog and associated 3D velocity model for the Ecuador subduction zone is paired with satellite surface deformation measurements to provide detailed information on rupture on the subduction megathrust (Domain 2) over the seismic cycle. Deeper, the Ecuador catalog provides insight into intraplate earthquake generation at
intermediate depths (Domain 3). Both the Wharton Basin and Ecuador studies include regionally significant large-to-great magnitude 7+ mainshocks that occurred over this decade.

Chapter 2 presents DD relocation results of the earthquake sequence generated by the April 11, 2012 (Mw 8.6) Wharton Basin mainshock, the largest intraplate strike-slip earthquake recorded. Aftershocks, including a second M8+ earthquake, activate a complex set of faults. Along all faults, aftershocks occur outside the highest coseismic slip associated with the largest earthquake on each fault and tend to cluster in low slip or slip edge areas. Events with depth phases (pP) generally correspond to predicted pP-P time observations in the ~5 km to 35-40 km depth range, such that the deepest limit occurs within the expected depth limit of brittle seismic failure at 600°C. The relocated earthquakes provide supporting evidence that the Wharton Basin sequence ruptured the entire oceanic lithosphere.

In Chapter 3, I use joint teleseismic DD relocation and tomography to examine seismic locations on the Ecuador subduction margin that includes a detailed study of the 2016 Mw 7.8 Pedernales megathrust earthquake sequence. I model interferometric synthetic aperture radar (InSAR) data derived from the Sentinel-1A satellite for the Mw 7.8 mainshock and Mw 6.7 and Mw 6.9 aftershocks. Results are compared with the high-resolution earthquake catalog and published interseismic coupling models to investigate fault segmentation along inferred seismic asperities. The mainshock initiated off-shore and most likely ruptured the same asperity region as the 1945 Mw 7.8 event. The two largest aftershocks ruptured northeast of the mainshock, overlapping in a region with moderate to high plate coupling and outside areas of aseismic slip. The seismic and geodetic data illuminate a highly segmented megathrust fault.

In Chapter 4, focus shifts to the subduction system at intermediate depths. Teleseismic DD tomography results improve images of slab structure under Ecuador compared to the 3D
starting model derived from global tomography. Localized clustering of intermediate depth earthquakes in Ecuador is controlled by contortion and bending on the Farallon Plate. In contrast, dehydration embrittlement and intermediate depth earthquakes have a strong global link from the correlation on seismicity rate (using the ISC catalog) with incoming plate fault throw measurements at major subduction zones. But no correlation is found with intermediate depth earthquakes and b-value.

I conclude in Chapter 5 how the results presented in this dissertation can improve global teleseismic catalog locations. The contribution of InSAR observations and teleseismic catalogs are useful tools in understanding earthquake hazard.
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To my Mom and Dad
1.1 Characteristics of Large Earthquakes

Large earthquakes (Mw 7.0+), which exhibit slip over long and wide fault dimensions hundreds of kilometers away from the hypocenter, uniquely provide observations of rupture heterogeneity and insight into the mechanics of faulting and seismogenic processes. Earthquakes with the same moment can produce slip over variable rupture dimensions, with or without directivity, over a range of slip rates, durations and amounts along the fault (e.g. Kanamori and McNally, 1982; Ye et al., 2018). What controls earthquake rupture initiation, termination and propagation are still poorly understood across major fault systems around the world. Specifically, what geophysical or dynamic properties of an earthquake-generating fault influence the observed seismic rupture or regions of high-slip, termed asperities (e.g. Lay and Kanamori, 1981; Ruff, 1992)? Rate-and-state friction models suggest earthquakes nucleate at asperities where coupling is high and/or frictional stability is unstable, but that slip may propagate into regions of conditional frictional stability under high slip rate (Dieterich et al., 1992).

Spatial correspondence between interseismic coupling and slip patches of subsequent large earthquakes has further supported the conventional “asperity model,” which posits that asperities on faults are localized patches (Figure 1.1) that are reused over multiple seismic cycles from accumulating strain during the interseismic period to hosting high slip during an earthquake
(e.g. Chlieh et al., 2011, Chlieh et al., 2014). In this paradigm, the regions between asperities creep without accumulating stress and behave in a velocity strengthening frictional mode. The asperity model suggests that rupture segmentation defined by persistent asperities exists over multiple seismic cycles and that the asperities fail in a “characteristic” manner in regard to timing and rupture extent (e.g. Schwartz and Coppersmith, 1984). Still, observations of one-to-one correlation between interseismic locking (strain accumulation) and coseismic slip over multiple seismic cycles and across major subduction zones are limited due to the long recurrence intervals and relative short observation period of modern onshore or offshore seismic and geodetic instrumentation.

Imaging slip patterns and locations of seismicity between large earthquakes provides observations on the fault-rupture history. Finite-fault modeling and back-projection of teleseismic waves have provided unprecedented details of seismic slip and rupture propagation during large earthquakes, thanks in part to the increasing number of dense seismic arrays globally (e.g. Ishii et al., 2005). Teleseismic waves, traveling at distances greater than 1000 km, provide the most rapidly accessible data for observing remote earthquakes, particularly oceanic events along subduction megathrust faults, transform faults, fracture zones, and mid-ocean ridges. The International Seismological Centre (ISC) aims to compile and review hypocenter locations and phase information reported in teleseismic catalogs with the mission to provide the most detailed and definitive record of the Earth’s seismicity (http://www.isc.ac.uk/).

In this dissertation, I explore how high precision teleseismic locations can be used to characterize fault rupture history and lithospheric structure with comparisons to other geophysical datasets, such as coseismic slip and fault geometry models. Using aftershock and background seismicity hypocenters, I investigate how deformation, fault segmentation and
earthquake rupture are accommodated in the incoming plate (prior to subduction), along the plate interface and within the slab at intermediate depths (Figure 1.1). Precise earthquake locations can infer the thermal and mechanic properties of the fault, as well as information about where earthquakes can nucleate. For subduction zones with accurate earthquake focal depths, information can be gathered about the steepness of the megathrust and the bending structures of down-going plates that can be related to the mechanics and deformation characteristics of the subducting oceanic lithosphere. Large intraplate and interplate earthquakes provide aftershock sequences for such studies.

The occurrence of large seaward intraplate earthquakes (off the trench) is also of interest as they may provide information relating to seismic coupling on the plate interface and partitioning of stresses in the subducting plate. The 11 April 2012 Mw 8.6 Wharton Basin earthquake remains the largest intra-oceanic plate strike-slip earthquake sequence ever recorded. The sequence started with the 10 January 2012 Mw 7.2 foreshock and the largest aftershock, an Mw 8.2, occurred ~two hours following the mainshock on a neighboring fault. Oceanic intraplate faulting outside of plate bending (outer-rise) regions is infrequent. These rare intraplate earthquakes result from the diffuse deformation within the Indian and Australian oceanic lithosphere plate (Gordon et al., 1998). The Wharton Basin, located off the west coast of Sumatra, is an extensively deforming ocean basin (e.g. Delescluse and Chamot-Rooke, 2007). Long north-south trending fracture zones populate the north Wharton Basin and originally appeared the most likely fault(s) to host the large 2012 strike-slip event. However, the wide-ranging studies that followed illustrated that the mainshock ruptured in multiple segments that trend parallel and near orthogonal to the fracture zones. A complex strike-slip fault system was illuminated from seismic observations, which indicated that possibly only a small segment of the
Mw 8.6 rupture (or possibly none at all) had taken place on the N-S oriented fracture zones (e.g. Hill et al., 2015; Singh et al., 2017). Observations of this earthquake led to a reevaluation of the maximum magnitude for intraplate strike-slip faulting and indicated that large earthquakes could generate considerable slip below the oceanic Moho within the oceanic mantle (McGuire and Beroza, 2012). The maximum depth of earthquakes has been observed to be inversely related to heat flow such that the brittle-ductile transition is defined by a threshold temperature limit (e.g. Bonner et al., 2003; Prieto et al., 2017). The plethora of great earthquakes (M8+) in the Wharton Basin provides the opportunity to study the seismogenic extent of oceanic intraplate faults.

Broadly speaking, the 2012 Wharton Basin earthquake sequence is thought to have been triggered by the great 2004 Mw 9.2 Sumatra megathrust subduction zone earthquake (Delescluse et al., 2012). Therefore, studies of the oceanic plate response to stress perturbations caused by large megathrust earthquakes becomes a relevant issue for understanding earthquake and tsunami hazard.

The largest earthquakes in the historic record, and those associated with the potential to generate tsunamis, occur on the megathrust fault between downgoing oceanic lithosphere and continental upper plates. The 16 April 2016 (Mw 7.8) Pedernales, Ecuador, megathrust earthquake resulted in over 650 fatalities and associated economic loss and recovery of ~ $3.3 billion USD (Meltzer et al., 2019). The mainshock appeared to rupture a portion of the subduction interface within the rupture limits of the great Mw 8.6 earthquake of 1906. The Pedernales event occurred inboard of subduction of the expansive Carnegie Ridge under north-central Ecuador and had a number of strong aftershocks. Subduction of the thicken lithosphere associated with the Carnegie Ridge had been associated with the recurrence time between large megathrust earthquakes, with the observation that the interplate boundary is partially decoupled.
and segmented (Chlieh et al., 2014). Initial studies indicated that the 2016 event re-ruptured the megathrust fault where a similarly sized (Mw 7.8) event occurred in 1945. Understanding the sequence and the evolution of seismogenesis over time bears on broader questions of fault segmentation and asperity reuse over multiple seismic cycles. The Pedernales earthquake resulted in a rapid deployment of onshore and offshore seismometers, accelerometers, ocean bottom seismometers (OBS) and GPS receivers, in a true international rapid response (Font et al., 2016; Meltzer et al., 2019). As a result, in addition to teleseismic catalogs, the onshore-offshore local and regional country-wide seismic data and hypocenter catalogs are or will be available for comparison study. Finally, the Sentinel 1-A satellite acquired data in-between the time of the three largest events of the Pedernales sequence, which allows detailed comparisons of surface deformation and modeled fault slip between the three largest events and offers a complementary but independent dataset to verify earthquake catalog hypocenters.

The Wharton Basin and Pedernales earthquakes illuminate the ways in which large intra-oceanic and interplate earthquakes are influenced between the incoming plate (seaward from trench) structure and the megathrust. One open debate is whether the earthquake size and rupture limit is primarily controlled by tectonic stresses or the structural irregularities along the plate. Marine seismic experiments using multibeam bathymetry and/or seismic reflections image the crustal structure off-shore of Sumatra (e.g. Singh et al. 2017) and of the Ecuador (e.g. Collot et al., 2017) convergent margin. While high-resolution bathymetry and seismic experiments provide detailed images of the structural heterogeneities in the faulting region, the pattern and distribution of seismicity helps to confirm and map the spatio-temporal location of deformation. Similarly, both the Wharton Basin and Ecuador earthquakes rupture multiple segments or asperities along strike and all or a significant percentage of downdip fault. Mapping fault
segmentation aids in understanding the maximum earthquake magnitude produced within a segment and across multiple segments. The seismic and tsunami hazard potential can then be refined.

Figure 1.1. Schematic diagram of subducting oceanic lithosphere (not to scale) highlighting three seismic deformation domains: incoming plate faults (orange lines), megathrust fault (green rectangle) intraslab faults on the down-going plate (red lines). The black patches on the megathrust fault represent multiple asperities of high-slip areas where large earthquakes occur. Arrows indicate coupling of the overriding plate to the downgoing plate during the interseismic portion of the seismic cycle.
1.2 Double-Difference Earthquake Relocation and Tomography

Earthquake location is an inherently non-linear problem. The arrival time, $T_k^i$, of the earthquake $i$ recorded at seismograph $k$ is calculated as the travel time, $T$, which is itself a function of latitude $X$, longitude $Y$, depth $Z$ of both earthquake and recording station added to the earthquake origin time $t_i$.

$$T_k^i = T(X_k, Y_k, Z_k, X_i, Y_i, Z_i) + t_i$$  \hspace{1cm} (1)

Since the seismograph location is already known $(X_k, Y_k, Z_k)$, there are four unknowns $(X_i, Y_i, Z_i, t_i)$ that need to be determined. The most fundamental numerical approach to earthquake location is Geiger’s method (1910) that linearizes the problem by using an iterative least squares technique to estimate the earthquake hypocenter location and origin time. For single event location hypocentral parameters for earthquake $i$ are found by minimizing the travel time residuals, $r_k^i$, for station $k$ through perturbations $\Delta m$ for $(X_i, Y_i, Z_i, t_i)$.

$$\frac{\partial T_k^i}{\partial m} \Delta m^i = r_k^i$$  \hspace{1cm} (2)

The travel-time residual $r_k^i$ is the difference between the observed and calculated travel-time, $T^{obs} - T^{cal}$. While modifications to the Geiger method and advanced relative location and Bayes statistical approaches developed, teleseismic catalogs like the Advanced National Seismic System (ANSS) Comprehensive Earthquake Catalog (ComCat) and the International Seismological Center (ISC) Bulletin still rely on single-event approaches based on Geiger’s method to produce rapidly available catalogs.
I improve teleseismic earthquake locations in this dissertation by using the relocation technique known as the double-difference (DD) method. Here, absolute travel times for single earthquakes are combined with differential travel-times (DT) of closely spaced event pairs (Waldhauser and Ellsworth, 2000; Zhang and Thurber, 2003). The combination of data yields improvements in relative relocation of neighboring earthquakes while retaining control of earthquake location in absolute space. Neighboring event pairs are assumed to have similar travel-paths from source to station and that the differences in arrival times for each station are representative of the difference in hypocentral location. The original double-difference equation (Waldhauser and Ellsworth, 2000) can be used to minimize the arrival-time residuals, $dr_{ki}^{ij}$ of event pair $i$ and $j$ and is expressed as the difference between the observed and calculated travel-times for station $k$.

$$dr_{ki}^{ij} = (t_k^i - t_k^j)^{obs} - (t_k^i - t_k^j)^{cal}$$ (3)

Adjustments to hypocenters from event pairs resolve relative hypocenter estimates as expressed as:

$$\frac{\partial t_k^i}{\partial m} \Delta m^i - \frac{\partial t_k^j}{\partial m} \Delta m^j = dr_{ki}^{ij}$$ (4)

Zhang and Thurber (2003) modified the DD method to retain the absolute arrival time information and simultaneously solve for three-dimensional velocity structure. The concept was extended to include the use of teleseismic data by Pesicek et al. (2010). Details for applying DD relocation for teleseismic events follow in Chapters 2 and 3.
1.3 InSAR Applications for Coseismic Deformation

Interferometric synthetic aperture radar (InSAR) is a satellite remote sensing application that can be used to measure the ground deformation of moderate to large earthquakes. An InSAR image containing the coseismic deformation signal is produced with at least one satellite radar image acquired before and after the earthquake. The observed interferometric phase ($\varphi_{insar}$) represent the differences of the phases returning to the satellite with the ground displacement ($\varphi_{defo}$), topographic ($\varphi_{topo}$), atmospheric ($\varphi_{atmo}$) signals and noise superimposed in the phase image. The differential SAR phase is described as,

$$\varphi_{insar} = \varphi_{defo} + \varphi_{topo} + \varphi_{atmo} + \text{noise}$$

(5)

The topographic signal can be removed with a digital elevation model (DEM) and precise orbital data. Atmospheric disturbances (artifacts related to stratified and turbulent troposphere and ionosphere) are potential sources of error and the signal is strongly correlated in space but not correlated in time. These artifacts can be removed using atmospheric correction models. In the dissertation, InSAR processing algorithms and modeling approaches are described in Chapter 3.

1.4 Summary of Studies

Teleseismic data and hypocenter locations featured in Chapters 2 through 4 include the GeoForschungsZentrum (GFZ) catalog, USGS ComCat and the ISC Bulletin. The ISC Bulletin dates to the early 1960’s and is the most comprehensive global seismic catalog; it includes data from the aforementioned catalogs and many other seismic networks in order to provide a uniformly reviewed and revised ISC hypocenter location.

Chapter 2 focuses on the 2012 Wharton Basin earthquake sequence and relocates the ISC and GFZ catalogs to examine intraplate faulting in the incoming plate region. Note that both the
ISC and the GFZ catalog includes depth phases, which is particularly important to studies of oceanic plate earthquakes. Comparison of aftershock locations to mainshock slip provides insight into moment release across the multiple fault segments. By constraining the depth distribution of the seismicity, I can also investigate whether slip reaches depths below the expected seismogenic zone based on temperature constraints. This work is submitted to the Journal of Geophysical Research under the following citation: Kwong, K.B., DeShon, H.R., Saul, J., and C. Thurber (2019). Full Lithospheric Rupture in the Wharton Basin Captured by Teleseismic Hypocenter Relocation of the Great 2012 Mw 8.6 Strike-Slip Earthquake Sequence. Journal of Geophysical Research.

Chapter 3 focuses on seismic deformation along the Ecuador megathrust fault using relocations of seismicity recorded from 1960 to 2018 and coseismic InSAR deformation associated with the 2016 Pedernales earthquake sequence. The aim is to understand how seismicity over the seismic cycle tracks fault segmentation and megathrust heterogeneity. The Ecuador subduction margin study uses joint teleseismic relocation and tomography. Relocation of both reviewed ISC Bulletin and the USGS ComCat links older events with depth phase control (ISC) to the NEIC catalog with no depth phases and shows the potential to improve teleseismic locations in a real-time procedure by linking catalog information via differential times. The relocation study is paired with InSAR observations of the mainshock and two large aftershocks, which provides high-resolution spatial locations of crustal deformation that can be used to compare and verify seismic source locations. The InSAR work presented also showcases the detectability and resolution of InSAR images for earthquakes with Mw > 6.5 < 7.0 in this region. The study is submitted for peer-reviewed publication under the following citation: Kwong, K.B., DeShon, H.R., Kim, J.W., and Z. Lu (2019). Resolving Teleseismic Earthquake
Chapter 4 revisits the Ecuador subduction zone and focuses on the intermediate depth seismicity and tomography. I discuss differences in intraslab deformation of two subducting plates (Nazca and Farallon). The Ecuador seismicity study is compared to global analysis of intermediate depth earthquakes that was a collaborative project exploring links between incoming plate faulting and intermediate depth seismicity (Boneh et al., 2019).

In Chapter 5, I summarize the major findings of each chapter that focuses on a different region on the subducting oceanic lithosphere. I also discuss ways to improve analysis of large earthquakes using InSAR and teleseismic data.
REFERENCES


CHAPTER 2

FULL LITHOSPHERIC RUPTURE IN THE WHARTON BASIN CAPTURED BY TELESEISMIC HYPOCENTER RELOCATION OF THE GREAT 2012 Mw 8.6 STRIKE-SLIP EARTHQUAKE SEQUENCE

2.1 Introduction

The nature of deep non-subduction related earthquakes within the continental versus the oceanic lithosphere are strikingly different. It is rare to observe continental earthquakes rupturing in the deep lithosphere below the Moho (e.g. Chen and Molnar, 1983; Prieto et al., 2017). Oceanic intraplate earthquakes (OIEs), however, can commonly occur in the upper mantle where brittle failure can be sustained at greater depths due to the temperature structure within oceanic plates (Abercrombie and Ekstrom, 2001). As the oceanic lithosphere cools with age away from the ocean ridge, ocean depth increases, lithospheric thickness increases, and heat flow decreases. As a result, thermal models of the ocean lithosphere can be approximated using the plate and half-space cooling models (e.g. Parsons and Sclater, 1977). The depths of OIEs generally do not occur at temperatures greater than 600°C (Abercrombie and Ekstrom, 2001; McKenzie et al., 2005), the favored isotherm that delineates the brittle-ductile transition zone and supports that temperature largely controls the depth extent of seismic failure in oceanic lithosphere. The interactions between aseismic and seismic slip in oceanic lithosphere have also raised questions about the depth extent of these processes. Recent observations on the Blanco Transform Fault from a temporary Ocean Bottom Seismometer (OBS) deployment suggests that the crust and shallow mantle on oceanic transform faults have different slip behaviors: mainshock-aftershock
sequences in the crust, and aseismic slip and swarm activity in the shallow mantle above the 600°C isotherm (Kuna et al., 2019).

Intraplate deformation in the Wharton Basin (WB) is observed to be almost exclusively strike-slip and has shown to be capable of generating large OIEs (Deplus et al., 1998; Delescluse and Chamot-Rooke, 2007). Located in the northeast Indian Ocean, the WB is an ocean basin that is subducting along the Sunda Trench and its seafloor is characterized by N-S oriented fracture zones. It is part of a broad distributed intraplate deformation zone within the Indo-Australian plate, but its regional plate motion is accommodated by subduction under Sumatra to the N-E and collision with India to the north. Until the 11 April 2012 mainshock (Mw 8.6), the two June 2000 Mw 7.9 earthquakes that struck the central WB off the coast of Sumatra were the largest observed strike-slip events in the region (Abercrombie et al., 2003). The 2012 mainshock is currently the largest observed strike-slip and intraplate event recorded, and it was followed by a Mw 8.2 aftershock approximately two hours later. The 10 January 2012 (Mw 7.2) foreshock was the first recorded earthquake to occur in this sequence (Figure 2.1).

The 2012 earthquakes are associated with the Ninety-East ridge (a proposed inactive hotspot trace that separates the Wharton Basin from the Central Indian Ocean), mapped fracture zones and unmapped faults, and characterized by deep coseismic rupture. Detailed mainshock finite-fault models (Yue et al., 2012; Wei et al., 2013; Hill et al., 2015) helped support information from aftershock locations and several back-projection studies (Meng et al., 2012; Satriano et al., 2012; Wang et al., 2012; Yue et al., 2012; Ishii et al., 2013) that show large rupture complexity on multiple faults. The initial studies largely argued for reactivation of N-S oriented fracture zones. Rupture observations on west-southwest trending faults were not previously supported by the mapped oceanic fabric until recently acquired bathymetry and
seismic reflection images revealed shear zone structures supporting the conjugate rupture observations and deformation on west-east oriented right-lateral faults (Singh et al., 2017). Global Centriod Moment Tensor (GCMT) solutions of the Mw 8.6 and Mw 8.2 events place the centroid depths at 45.6 km and 54.7 km respectively (http://www.globalcmt.org). Mainshock slip models resolve rupture on sub-faults extending to 60 km depth (Wei et al., 2013; Hill et al., 2015). For lithospheric age in this region (50 - 60 Ma), the expected brittle-ductile transition depth at 600°C is ~35 km, and rupture below this depth range indicates earthquake slip into the ductile regime.

The 2012 WB earthquake produced one of the most complex aftershock sequences ever recorded. However, the absolute locations, as for other earthquakes in remote oceanic settings, are not well constrained due to poor station coverage. The closest land-based seismic stations recording the 2012 intraplate sequence are on the Sumatra and Andaman Islands, ~ 400 to 500 km to the east. Global teleseismic data provide a more consistent arrival time and hypocenter dataset to study the seismicity distribution. There are a number of teleseismic catalogs, which vary in location procedure, magnitude determination, and uncertainty estimates for the 2012 WB earthquakes: the International Seismological Centre (ISC), U.S. Geological Survey (USGS) Comprehensive Catalog (ComCat), GeoForschungsZentrum (GFZ), and the Engdahl, van der Hilst and Buland (EHB) Bulletin (1960-2008). Based on studies following the 2004 Mw 9.2 Sumatra earthquake, hypocenter location uncertainties for teleseismic catalogs in the region can be greater than 20 km, in part due to unmodeled velocity structure associated with the subduction zone (Pesicek et al., 2010a). Strong tradeoffs between origin time and focal depth also present a challenge for earthquake location inversions. For shallow earthquakes (< 70 km depth) at teleseismic distances, accurate determination of focal depth from first arrivals is difficult without
other techniques; most common is the use of depth phase arrivals that reflect off the Earth’s surface. Global seismicity relocation studies have employed depth phase identification techniques to improve hypocenter location estimates (Engdahl et al., 1998). Rapid real-time teleseismic catalogs such as the ANSS ComCat mainly rely on and report out first arriving P phases but the reviewed ISC bulletin contains later arriving phases, including depth phases when available.

Regional teleseismic relocation results for the Sunda subduction zone show notable improvement in hypocenters and uncertainties over the initial teleseismic catalogs by using combined double-difference (DD) methods of catalog and cross-correlation (CC) derived differential times, a realistic global 3D velocity model, and depth phases (e.g. Pesicek et al., 2010b). Here, we employ teleseismic relocation on the 2012 WB sequence to improve hypocentral locations and to more accurately define the seismogenic structure of the region. The new high-precision hypocenter catalog is compared to mapped fracture zones and geophysical features to delineate fault geometry and structure. Improvement in earthquake focal depth helps to infer the thermal and mechanical structure of the deforming plate. The relocated catalog is compared with mainshock slip models and thermal models (half-space cooling) to understand seismic rupture extent in the WB.
Figure 2.1. Wharton Basin study area. Top inset map shows the bathymetry for the northeast Indian Ocean. Strike-slip global CMT solutions (beachballs colored by centroid depth) indicate deep moment release for the 2012 and 2018 Wharton Basin earthquakes and the 2014 Bay of Bengal earthquake. The shallow thrust CMTs highlight large ($M_w \geq 7.5$) megathrust events following the 2004 Sumatra-Andaman earthquake. The red outline box indicates the study area shown in the main map. Dashed black lines indicate plate boundaries (Bird, 2003); solid black lines indicate fracture zones (Matthews et al., 2011). In the lower map, epicentral locations from the new DD earthquake catalog are color coded by sections of time in 2012 and scaled by magnitude and the background color highlight ocean basin structure indicated by gravity anomalies.
2.2.1 Teleseismic Relocation

The differential time-based earthquake location methods are a long-standing tool for relative earthquake relocation (Fremont and Malone, 1987; Got et al., 1994). The rapid, joint inversion of catalog and waveform correlation-based differential times using DD approaches (e.g. Waldhauser and Ellsworth, 2000; Waldhauser and Schaff, 2008) has allowed for more precise relative hypocentral locations at regional and local distances. DD locations can yield smaller relative location uncertainties than locations determined from absolute times, under the assumption that the differential time catalog reflects only location differences, and not velocity differences between an event pair and the recording station. Subsequently, seismic tomography from both differential and absolute times was integrated with the DD method (tomoDD) to jointly invert for hypocenter locations and 3-D velocity structure (Zhang and Thurber, 2003). TeletomoDD was developed and applied for DD relocation and tomography using global teleseismic data (Pesicek et al., 2010b; Pesicek et al., 2014). A key difference in teletomoDD compared to tomoDD is the inclusion of a spherical earth tracer for a 3-D velocity model using the pseudo-bending (PB) method (Um and Thurber, 1987) in a spherical earth model with discontinuities (Koketsu and Sekine, 1998). TeletomoDD is used in this study to relocate seismicity in the WB using arrival time and differential time information from regional to teleseismic stations, without a joint tomographic inversion, due to the absence of seismic stations in the epicentral region.

The WB earthquakes occurred in the oceanic lithosphere west of the long-lived Sunda subduction margin. The standard global teleseismic earthquake catalogs uses 1-D global average velocity models that do not account for regional velocity variations. Recently developed global and regional 3-D tomographic models can help account for such large velocity variations in the
Earth. We adopt a nested regional-global 3-D velocity model developed for the Sumatra region to account for heterogeneous velocity structure within and outside the area of interest (Pesicek et al., 2010b, Pesicek et al., 2014). The global 3-D model MITP08 (Li et al., 2008) is downsampled to a grid spacing of 5° for computational efficiency and a regional 3-D velocity model (Pesicek et al., 2010a, Pesicek et al., 2014) is imbedded into the global velocity model from 88° to 96° longitude and -1° to 6° latitude, at 0 to 1075 km depth with a grid spacing of 0.5°. This regional P-wave model was built using arrival times from earthquakes recorded from 1964-2007.

As implemented in teletomoDD, the PB ray tracer accommodates primary phases (P, Pn, Pg, S, Sn and Sg) as well as pP and pwP depth phases that have a bounce point location representing where the upgoing depth phase reflect off the Earth’s surface. We use the ETOPO1 global relief model (Amante and Eakins, 2009) to correct for the topography (for pP) or bathymetry (for pwP) at the bounce point location. After the elevation corrections, the pwP observations reflect depth estimates from the earthquake depth with respect to ocean bottom.

We use a hierarchical weighting scheme for the inversion of multiple data types (Waldhauser and Ellsworth, 2000; Pesicek et al., 2010b) where the order of upweighted data relative to other data starts with absolute arrival times, then catalog differential times (CTDTs), and finally cross-correlation differential times (CCDTs). To test the effect of damping in the inversion, we ran multiple teletomoDD relocations of our dataset using different damping values to determine which values produced the lowest RMS residual. Table A.1 contains a complete description of weighting and damping parameters.
2.2.2 Catalog and Cross-Correlation Data

To build a catalog of absolute phase data and CTDT data, we use the reviewed ISC bulletin and the GFZ catalog. The ISC bulletin contains the most comprehensive set of globally recorded seismic hypocenters, phases, and magnitudes. The ISC review procedure re-examines seismic phases reported from different seismic networks and provides a consistently determined hypocenter catalog similar in methods to EHB-determined hypocenters (Engdahl et al., 1998, Bondar and Storchak, 2011). A rectangular area search from -1° to 6° latitude and 88° to 96° longitude returns 1100 events from January to December 2012 for ISC reviewed authored hypocenters. P, Pn, S, and Sn first arriving phases and pP and pwP depth phases were compiled.

The GFZ network provide a second hypocentral catalog. The GEOFON seismic network of the GFZ Potsdam program includes close monitoring in the Indian Ocean region and regional seismic networks in Sumatra close to the WB events. A GFZ aftershock catalog following the 11 April 2012 mainshock contains manually picked P and S phases augmented by pP and pwP depth phases where available. The GFZ dataset contains 282 events from April to August 2012.

We build a combined ISC and GFZ catalog by merging phases and hypocenter information and identify 238 overlapping events. Of these events, the travel-times of the GFZ phases were calculated based on the reported arrival time with respect to the ISC reported origin time. Therefore, the ISC location is used for the ISC-GFZ matched events. For any station-phase reported more than once, the average travel-time is used. The combined ISC-GFZ catalog contains 1144 events. For our relocation work, only 696 events have a station azimuthal gap ≤180° and number of P phases > 9 (Table 1). For the 696 events, CTDTs are calculated by linking a network of event-station pairs for each phase in the catalog. The criteria for linking
events include if the hypocentral distance between the event pair is less than 100 km. For primary phases, the event pair required at least 5 station DT observations, but for depth phases only 2 were required. The number of CTDT data is summarized in Table 2.1.

Waldhauser and Schaff (2007) showed that cross-correlation (CC) of regional to teleseismic waveforms using a 0.1-2.0 Hz filter and -5 to 5 s time window can be used to improve DD locations of subduction zone seismicity. The CC data provides more precise differential time measurements to control the last stage of inversion for relative relocation. The large WB teleseismic dataset requires a significant amount of broadband waveform data and computational time to perform CC over every possible event-station pair. Similar in procedure to Jones (2014), I use the Geophysical Institute of Seismology Matlab Objects (GISMO) toolbox (Reyes and West, 2011) to bring efficiency to the data downloading, organization, visualization, and correlation calculation procedure. GISMO provides tools to access and automate retrieval of continuous waveform data, station metadata, and event information from the Incorporated Research Institutions for Seismology (IRIS) Data Management Center (DMC) archive. We correlate waveform data for P, Pn, and pP phases for a time window of -3 to 3 s surrounding the ISC or GFZ catalog reported phase arrival times. Using a subset of stations, we test various bandpass filters and time windows and for the final set of differential times, we use a Butterworth filter with a bandpass of 0.75 to 2.0 Hz. We only calculate CC for events with no more than 100 km hypocenter separation in the initial catalogs and report DTs for CC coefficients greater than 0.7 (Table 2.1). Note that only 203 stations of 1769 have available data at the DMC for correlation, and the resulting CCDTs represent a small percentage of the overall dataset (Table 2.1).
Table 2.1. Number of Events, Stations and Phases for teletomoDD Relocation

<table>
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<th>S-arrivals</th>
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<td></td>
<td>21822</td>
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</table>

2.3 DD Location Uncertainties and Depth Constraint

The final relocation dataset contains 695 events from the 696 events in the starting catalog and will be referred to as the DD catalog. TeletomoDD does not provide formal hypocenter uncertainties, so instead, a bootstrap approach provides statistical analysis of the DD locations that reflect an estimate of absolute location uncertainty. The bootstrap analysis incorporates 50 additional teletomoDD runs with each run having 10% of the absolute and differential time data randomly removed. We differenced the epicenter, focal depth and origin time for each event from their 50 boot-strapped locations. The change in location between the DD catalog and boot-strapped DD catalog generally fit a normal distribution (Figure A.2). For the full catalog, the location differences are normally distributed for focal depth (km) and origin time (s), but the epicenter differences follow a log-normal distribution (right skewed curve) because all values are positive. The mean standard deviations for epicenter, focal depth, and origin time difference is 2.6 km, 1.1 km and 0.17 s respectively and reflect the mean location uncertainty associated with the DD catalog. The standard deviation values for each event (latitude, longitude and depth) are shown as uncertainty bars (Figure A.4 and Figure A.5).

We compare pP-P delay time observations to predicted delay times prior to and after relocation to assess the robustness of the relocation depths. The travel-time curves shown in
Figure 2 illustrate all pP-P observations along with tau-p predicted pP-P curves for selected earthquake depths. Figure 2.2a shows that pP-P delay times for intraplate events are primarily associated with shallow, and in many cases default, hypocentral depths in the ISC+GFZ catalog (10-15 km depth). Following relocation, the earthquake depths better reflect the range in delay times and earthquakes extend from ~5 km to ~40 km (Figure 2.2b). This analysis indicates an important caveat: intraplate events between ~40-55 km in depth have many pP-P delay times that correspond to predicted depths of <40 km. In other words, the deeper events in the DD catalog are not consistent with reported depth phases. There are also depth inconsistencies for events occurring on the megathrust or within the Sunda Plate (subduction related events). For this region, the depth phase observations are more consistent for shallower events < 30 km but there is more scatter for deeper events > 30 km. In the depth phase catalog, the deepest event is at 58 km and 73 km depth for the ISC+GFZ and DD catalog respectively. Out of 42 subduction related events, 24 events have a 10 to 20 km depth change between the ISC+GFZ and DD catalog.
Figure 2.2. Travel-time plots for observed pP-P delay measurements versus epicentral distance. Each symbol represents a single station pP-P observation and is colored by the depth of the earthquake in the (a) initial reported catalog and (b) the DD catalog. The circles are oceanic intraplate events and diamonds are subduction related events. Predicted pP-P travel time curves for select event depths are also shown (solid black lines).

2.4 Results and Interpretation

The DD locations and the ISC + GFZ absolute catalog locations are shown in Figure 2.3. We differenced the locations between the two catalogs and found the mean epicenter, focal depth and origin time difference to be 13 km ± 6 km, 7 km ±5 km and 0.8s ± 0.7s, respectively. Seismicity is located in several different clusters that are oriented in a conjugate pattern. Using the Spatial Statistics Toolbox in ArcGIS, the kernel density function and standard deviational ellipse was used to calculate the density and directional distribution. The formal spatial analysis delineates 7 major seismic clusters and cluster azimuths (Figure A.6) to aid in visual comparison. The ISC + GFZ and DD catalogs are projected onto 7 sub-fault cross-sections denoted as K1 to K7 where events up to 50 km from each side of the line (Figure 2.3). Cross-sections show that the majority of hypocentral depths reported in the ISC + GFZ catalog were fixed either at a
shallow 10 km depth or deep 35 km depth estimate. Regional structural information of the ocean bottom and Moho is obtained from the LITHO1.0 model (Pasyanos et al., 2014). The 10 km depth events are within the oceanic crust and the 35 km depth events are well below the Moho (~16 to 18 km depth). In comparison, the DD catalog provides a more plausible image of the earthquake depth distribution.

The spatial relationship between the DD locations, which are primarily aftershocks, and the mainshock finite-fault model from Hill et al. (2015) is shown in Figure 2.4. The Hill model indicate a complex co-seismic slip distribution over six sub-faults labeled H1-H6. The Hill model is the only published finite-fault models of this sequence to resolve slip using both teleseismic and GPS observations. For each sub-fault, the stress drop from Hill et al., (2015) is indicated next to the slip cross-sections. Areas with the most slip and largest stress-drop, on average, are located on the H1 and H2 faults oriented west-southwest. The highest slip regions (> 20 m) are confined above the LITHO1.0-defined Moho on all faults. Very few aftershocks occur within regions of high slip. Instead, aftershocks cluster in low slip regions or near edges of slip patches. On the fault where the Mw 8.6 mainshock initiated (H1), the aftershocks primarily occur below the Moho, and the deepest events are located away from significant mainshock slip. The fault segment that shows the deepest mainshock slip is H2, indicating rupture down to 60 km. Each of the six sub-faults show slip and aftershock locations simultaneously in the oceanic crust and upper mantle to depths between ~35 to 40 km.

We calculate a half-space cooling (HSC) model to explore the temperature limits of seismicity in the region. The HSC model includes a thermal diffusivity of 1e6 m²/s and a mantle temperature of 1350°C, as similarly modeled for this region in Aderhold and Abercrombie (2016). The hypocenter depth of each intraplate earthquake in the DD catalog is plotted against
ocean lithosphere age (Figure 2.5). The 2012 WB earthquake sequence spans lithosphere \( \sim 45 \) to \( \sim 65 \) MA in age, where the \( 600^\circ C \) isotherm boundary falls at \( \sim 40 \) to \( \sim 50 \) km depth. There are 64 events within the \( 600^\circ C - 800^\circ C \) isotherm and 10 events within the \( 800^\circ C \) to \( 1000^\circ C \) isotherm. The events below the \( 600^\circ C \) isotherm, the assumed brittle-ductile transition, tend to occur along specific fault segments. They cluster near the mainshock epicentral region (K1), the “outer rise” region near the Sunda trench (K3), the eastern side of the Ninety-East Ridge (K6) and the fault associated with the \( M_w \) 8.2 aftershock (K7) (Figure 2.3 and Figure 2.5a). The centroid locations from the gCMT catalog place the mainshock within the \( 600^\circ C - 800^\circ C \) isotherm range and the \( M_w \) 8.2 aftershock below the \( 800^\circ C \) isotherm. The deepest aftershock locations are located near these centroid depths and are consistent with seismic slip models (Wei et al., 2014, Hill et al., 2015), which also indicate deep co-seismic rupture between 50 to 60 km depth.
Figure 2.3. Comparison of ISC + GFZ catalog (yellow circles) vs. DD catalog (purple circles) locations. Left map shows epicenters determined in this study and the cross-section lines, labeled K1-K7 (blue), and the finite fault geometry used by Hill et al. (2015), labeled H1-H5 (red). Fracture zones (solid black lines) and plate boundary (dashed black lines) are shown as in Figure 2.1. Right cross-sections show seismicity from K1-K7 lines projected from 50 km on each side. The two solid black lines in each cross-section indicate ocean bottom (upper line) and Moho (lower line) from LITHO1.0 model (Pasyanos et al., 2014).
Figure 2.4. DD Aftershock distribution relative to mainshock slip. Slip model (H1-H6) from Hill et al. (2015) showing no slip (purple) to peak slip (red) at ~50 m. White circles indicate DD relocated aftershocks in relation to the mainshock rupture. The ocean bottom (black line), sediment layer (gray line) and Moho (white line) are from LITHO1.0 (Pasyanos et al., 2014). The stress drop associated with each sub-fault slip model are from Hill et al. (2015). White star indicates Mw 8.6 mainshock DD location.
Figure 2.5. DD locations for oceanic intraplate events only are shown with associated isotherm range according to the half-space cooling model. Events in 0-200°C are shown in purple, 200-400°C in blue, 400-600°C in green, 600-800°C in orange, and 600-800°C in red. Stars are events with depth phase data and circles are without. (a) Epicentral locations with oceanic lithosphere age (Ma) in grayscale. (b) Half-space cooling model showing the distribution of earthquake depth with half-space cooling isotherms. The gCMT lower hemisphere solution for the Mw 8.6 mainshock and Mw 8.2 aftershock are also shown.
2.5 Discussion and Conclusion

Usually, OIEs in transform fault and fracture zone settings generate a small number of aftershocks reported in teleseismic catalogs (e.g. Das and Henry, 2003; Antolik et al., 2006). Exploring the spatial and temporal relationship between mainshock and aftershock rupture can be difficult and linking that information to data on temperature and lithospheric strength can be more difficult. Still, the 2012 WB sequence was unusually large and complex and occurred near a number of high-quality seismic stations deployed following the 2004 Sumatra-Andaman Islands earthquake. We can therefore better explore the full range of oceanic crust and lithosphere rupture using the DD catalog.

Generally, aftershocks concentrate in low-slip areas or along edges of high-slip areas that may represent unbroken asperities with implications of where the stress is focused due to the static stress changes after the mainshock co-seismic rupture (e.g. Das and Henry, Mendoza and Hartzell, 1988). The 2000 M$_w$ 7.8 WB OIE earthquake was argued to have ruptured with significant moment energy release simultaneously on two conjugate faults, but aftershocks almost exclusively occurred on the N-S oriented fault plane (Robinson et al., 2000). Low aftershock productivity was seen in low-slip regions between large slip asperities for great strike-slip events such as the 2003 M$_w$ 7.6 Mid-Indian Ocean (Antolik et al., 2006) and the great 1998 M$_w$ 8.1 Antarctic Plate earthquakes (Henry et al., 2000). For megathrust faults, low uncertainty and high-quality datasets of co-seismic slip models and aftershock catalogs have shown more systematically that there is a deficiency of aftershocks occurring inside high co-seismic slip regions (e.g., Weltzer et al., 2018). This relationship also holds for large OIE ruptures associated with the 2012 WB sequence. Our refined DD aftershock locations are not co-located with high
slip. The Hill model for the mainshock resolves a high-slip asperity with maximum slip of ~50 m and there are no aftershocks within this maximum slip asperity.

There are a number of deep intraplate earthquakes occurring below the 600°C isotherm in the DD catalog. We previously discussed that pP-P delay times largely agree with the relocated depths in the DD catalog (Figure 2). However, deep intraplate events in the DD catalog are constrained by fewer phases (Figure 2.6) and exhibit more inconsistent pP-P times (Figure 2). This is surprising since we expect depth phases to be easier to identify for deeper events as their phases should be more impulsive and less dominated by high-frequency coda. Possibly for earthquakes in the mantle, pmP phases arriving as precursors to pP might cause depth phase misidentification. Picking pmP arrivals as depth phases would yield apparent pP-P delay times consistent with shallower predicted earthquake depths. However, pmP reflections off the Moho tend to be weak (McGlashan et al., 2008), and a pP arrival should be stronger than pmP. What seems to control the smaller number of delay time observations below the 600°C isotherm (Figure 2.6) is that these events also exhibit a smaller number of P-wave station observations (Figure 2.6a). Use of the delay times alone would argue for caution in interpreting the deeper earthquakes in the DD data. However, we also note that the depth phases used in this study were largely identified manually and incorporated into the GFZ dataset. Efforts to relocate the deep earthquakes to shallower depths more consistent with delay times by using a range of starting positions and combinations of phases still results in a DD catalog with some deep earthquakes.

The WB events that occurred below the 600°C isotherm (Figure 2.5) signify that deep earthquakes are not restricted to a particular temperature boundary. Apart from subducted slabs, unusually deep isolated earthquakes in oceanic and continental lithosphere are rare but raise
implications on alternative controls of seismic slip not present in the brittle regime. The 2013 M\text{w} 4.8 earthquake in the Wyoming continental craton occurred at a focal depth of \(~75\text{ km} \) at temperatures modeled to be above 750°C (Prieto et al., 2017). A year later, the 2014 Bay of Bengal (M\text{w} 6.0) earthquake occurred in the oceanic lithosphere at a focal depth of \(~50\text{ km} \) at temperatures predicted between 600°C and 700°C (Rao et al., 2015, Aderhold and Abercrombie, 2016). More recently, the 2017 M\text{w} 8.2 Tehuantepec, Mexico intraslab event was inferred to have ruptured down to the 1,100°C isotherm from the reactivation of preexisting bending faults under wholesale deviatoric tension of young oceanic lithosphere (Melgar et al., 2018). All these events are anomalous in that laboratory evidence for the brittle-ductile transition zone show olivine transitions from velocity strengthening to velocity weakening at 600°C, agreeing with the temperature limit and focal depths of most oceanic earthquakes (Boettcher et al., 2007). Earthquakes occurring between 600°C and 800°C specifically have been explained by a thermal runway feedback mechanism from rapid strain rates (Kelemen and Hirth, 2007). McGuire and Beroza (2012) appealed to this mechanism to explain the high stress drop, large magnitude and deep rupture extent of the 2012 WB mainshock.

Seismic slip can propagate below the seismogenic layer under rate-and-state friction assumptions (e.g., Scholz 2012). As was shown in Shawn and Wesnousky (2008), who used elastodynamic modeling, large earthquakes can produce rupture penetrating the stably sliding (aseismic and velocity strengthening) region of the fault. When large coseismic slip penetrates below the seismogenic layer, the stiffness of the fault decreases and rapidly allows additional slip to be accommodated in the underlying aseismic region. This is argued to create a deep conditionally stable layer that can rapidly accommodate large co-seismic rupture, but the initiation of deep aftershocks in this layer has been questioned. Evidence for aftershocks below
the seismogenic layer has indicated that the brittle-ductile transition zone may temporarily change in depth depending on strain-rate dependent post-seismic processes (Rolandone et al., 2004). After a large earthquake, aftershock depths can increase over time but is followed by post-seismic shallowing, indicating that the brittle-ductile transition zone is temporarily lowered, although, the brittle failure limit has only been seen to deepen temporarily by a few kilometers in depth. Earthquake generation in the ductile lithosphere might also be explained by presence of heterogeneities in the upper mantle. Localized seismic deformation within the lower crust and upper mantle in the Newport-Inglewood fault in southern California has been argued that compositional or mineralogical heterogeneity play a role in forming a network of seismic asperities imbedded in the ductile regime (Inbal et al., 2016).

The 2012 WB mainshock generated one of the most complex aftershock sequences ever recorded in the oceanic lithosphere. Spanning over at least seven fault zone regions, this diffuse and complex aftershock sequence provided the opportunity to gain a better understanding of seismogenesis of moderately sized intraplate earthquakes in the region. Our study improved hypocenter locations, particularly for focal depth determination in the WB for seismicity preceding and following the mainshock. Although the mainshock exhibited complex faulting, we also find that the mainshock-aftershock sequence shares similar characteristics seen in continental and megathrust earthquake sequences. The refined DD catalog shows that the aftershocks occur in areas outside or at the edge of high co-seismic slip, a consistent pattern in many other earthquake sequences that suggest aftershocks nucleate where the stress is redistributed after mainshock rupture. We also find that the earthquake depths are generally constrained by the 600°C isotherm, the expected depth limit of seismic rupture. The faults
associated with the mainshock have seismicity occurring to the base of the inferred seismogenic zone illustrating that these faults can fail throughout the entire lithospheric thickness.

Based on the pP-P observations and small number of deep intraplate earthquakes, we conservatively argue that the depths of earthquakes in the DD catalog are best constrained above the 600°C isotherm, that faulting in the WB is largely thermally controlled, and earthquakes initiate from the shallow crust to the base of the lithosphere. In the Hill model, all 6 sub-faults contain segments of notable co-seismic slip reaching 40 km in depth, corresponding to right below the 600°C isotherm, but aftershocks that occur in this depth range are mostly located in areas with no co-seismic slip. Faults in the WB could be analogous to the Newport-Inglewood fault where seismicity is mostly limited to above the brittle-ductile transition except in very localized areas where earthquakes can nucleate in what is modeled to be in the ductile regime.
Figure 2.6. Number of P and S phases with respect to half-space cooling model. (a) The number of P phases for events with depth phases and (b) the number of P and S phase station pairs for all intraplate events in the DD catalog are plotted with lithospheric age and relocated focal depth. The sizes of the purple circles are scaled according to the number of phases. The black lines indicate isotherms from the Wharton Basin half-space cooling model.
REFERENCES


Amante, C. & Eakins, B.W., 2009. ETOPO1 1 arc-minute global relief model: procedures, data sources and analysis, NOAA Technical Memorandum NESDIS NGDC-24


CHAPTER 3

RESOLVING TELESEISMIC EARTHQUAKE CATALOG AND INSAR DATA DISCREPANCIES IN ABSOLUTE SPACE TO EXPLORE RUPTURE COMPLEXITY ALONG THE ECUADORIAN MEGATHRUST FAULT

3.1 Introduction

Studies of megathrust earthquakes along the Columbia-Ecuador subduction zone were influential to derive the asperity model for rupture heterogeneity (e.g. Lay et al., 1982; Kanamori and McNally, 1982), here defined as the location of increased strength and high stress on a fault that in turn produces the largest slip during an coseismic event. Modern geodetic models confirm the presence of a heterogenous fault with discrete areas of high coupling, interpreted as asperities frictionally locked and accumulating strain, surrounded by areas with variable coupling (e.g. Moreno et al., 2014). Geodetic data from Ecuador establishes a subduction plate interface with several large asperities (Chlieh et al., 2014, Nocquet et al., 2014, Nocquet et al., 2017) surrounded by areas of aseismic slip (Vallee et al., 2013; Rolandone et al., 2018; Vaca et al., 2018). Variable resolution between geodetic and seismic datasets of megathrust faults, which lie largely offshore, continues to limit efforts to map and compare coupling and high slip. Geodetic and seismic slip modeling requires a known geometry for the subducting slab, which in turn is frequently constrained by teleseismic and regional to local earthquake catalogs. The catalogs can exhibit tens of kilometers spatial bias, especially in depth, due to incorrectly modeled velocity structure (e.g. Engdahl et al., 1998 and references therein). The Columbia-Ecuador subduction zone now provides key datasets to resolve these issues due to the growing availability of far-field
and near-field seismic and geodetic sensors deployed prior to and following the 2016 $M_w$ 7.8 Pedernales earthquakes. Combined, the data provides improved resolution to image spatial discrepancies in the teleseismic and satellite geodetic datasets, with independent local seismic data for verification.

Subduction of the Nazca plate below Ecuador has generated a history of large to great earthquakes, and the 16 April 2016 $M_w$ 7.8 Pedernales earthquake is the most recent (Figure 3.1). The largest, the 1906 $M_w$ 8.8 earthquake, likely ruptured all or most of the megathrust fault (~500 km) (Kelleher, 1972; Kanamori and McNally, 1982). Subsequent smaller events (1942 $M_w$ 7.8; 1958 $M_w$ 7.7; 1979 $M_w$ 8.2; 1998 $M_w$ 7.1) are thought to have ruptured several discrete asperities within the 1906 slip region based primarily on aftershock locations and coseismic slip models (e.g. Kanamori and McNally, 1982; Mendoza and Dewey, 1984; Ye et al., 2016), although a recent slip model of the 1906 event suggests it did not overlap with the later earthquakes (Yoshimoto et al., 2017) (Figure 3.1). The 2016 event occurred on the northern edge of the subducting Carnegie Ridge and the 1942 earthquake. Here, the subduction rate is ~47 mm/yr (Nazca Plate with respect to the North Andean Silver) and the convergence direction is slightly oblique to trench strike (Chen et al., 2001; Nocquet et al., 2014; Yepes et al., 2016; Nocquet et al., 2017) (Figure 3.1). Whether the 2016 rupture overlapped with the 1942 event has been discussed in several studies (Ye et al., 2016; Nocquet et al., 2017; He et al., 2017; Yoshimoto et al., 2017; Yi et al., 2018) but remains a point of scientific debate.

Along the Ecuador-Columbia subduction zone, interplate seismicity, positive Bouguer anomalies, and marine terraces spatially correspond to areas with significant or high plate coupling, resulting in a highly segmented margin (Gutscher et al., 1999; Font et al., 2013). Different scales of bathymetric high features entering the trench along Ecuador, such as the large
Carnegie Ridge and smaller seamounts, produce variable plate coupling reflected in the geodetic interseismic coupling models (Chlieh et al., 2014; Collot et al., 2017; Nocquet et al., 2017) (Figure 3.1) and mapping of aseismic slip (Nocquet et al., 2017; Rolandone et al., 2018; Vaca et al., 2018). From –2°S to 1°N, the megathrust fault has been divided into three distinctive regions (Figure 3.2). The Galera alignment is an active region of seismicity that trends perpendicular to the trench; the 2016 Pedernales epicenter lies within or at the edge of this segment (Figure 3.2b). The Jama and Manta Puerto Lopez segments are associated with two regions of strongly clustered seismicity on the subducted Carnegie Ridge. Seamount subduction occurs between the Galera and Jama seismic region, in an area interpreted as weakly coupled and relatively aseismic (Marcaillou et al., 2016). Subducted physical features such as seamounts and ridges, changes in age and thermal profile of the upper or lower plate, and variable changes in composition or thickness of incoming sediments have been used to explain heterogeneity along the South America subduction zone (Bilek, 2010), but improved earthquake locations in absolute space are required to fully map seismogenic variability along Ecuador (e.g., El Hariri et al., 2013).

We relocate teleseismic events along the Ecuador subduction zone region with two goals: 1) provide improved hypocenters prior to and after the Pedernales mainshock and InSAR derived deformation models to constrain shallow subduction zone processes; and 2) show how teleseismic double-difference (DD) tomography links phase data from different catalogs to remove spatial bias in global hypocenters. Here, we jointly solve for teleseismic relocations and a regional compressional 3D velocity model embedded within a static global 3D velocity model to better account for the strongly varying velocity field associated with the subducting Nazca lithosphere. The reviewed International Seismological Centre (ISC) Bulletin combined with the United States Geological Survey (USGS) Advanced National Seismic System (ANSS)
Comprehensive Catalog (ComCat) spanning from 1961 to 2016 (Figure 3.2a,b) provides the earthquake data. The ComCat includes rapid analysis and distribution of phase data for significant earthquakes, and we link ComCat to the comprehensive reviewed ISC catalog via phase differential times. We observe and model coseismic InSAR images of the 2016 $M_w$ 7.8 mainshock, the large $M_w$ 6.7 and the 6.9 aftershocks on 18 May 2016 (hereafter, aftershock 1 and aftershock 2 respectively) to provide spatial information of surface deformation during the sequence and confirm DD relocation results. We compare the DD catalog paired with the InSAR coseismic images (Sentinel-1A repeat interval of 12 days) with interseismic coupling models (Chlieh et al., 2014; Nocquet et al., 2017) to identify persistent and/or new asperities along the megathrust. The relationship between spatial patterns of coseismic slip, aftershock locations, interseismic plate coupling and aseismic slip models provide important assessment of the long-term earthquake and tsunami hazard of this margin.
Figure 3.1. Ecuador-Columbia subduction zone marks subduction of the Nazca plate under the northern part of South America (inset). Main map: Along Ecuador, the Carnegie Ridge and inactive spreading center structures subduct obliquely where the Nazca plate moves ~47 mm/yr with respect to the North Andean Silver with the South America Plate. Magenta indicates historic large earthquakes (magenta diamonds and lines) (Mendoza and Dewey, 1984); inferred rupture length of the 1906 earthquake (solid magenta line) (Yoshimoto et al. 2017); and alternative maximum rupture lengths from earlier studies (dashed magenta line). Recent earthquakes with gCMT solutions (beachballs) and USGS reported epicenter (stars) follows color coding: 1998 Mw 7.1 (teal), 1979 Mw 8.2 (purple), 2016 Mw 7.8 mainshock (red), Mw 6.7 aftershock (blue), Mw 6.9 aftershock (green). Interseismic plate coupling model is from Chlieh et al. (2014). Contour lines for the Slab2.0 model (Hayes, 2018) is in 20 km intervals with negative numbers indicating depth to top of slab.
3.2 Methods and Data

3.2.1 Earthquake Catalogs

Global and country-wide earthquake catalogs report seismic activity in Ecuador, and each dataset provides unique hypocenters, uncertainties, magnitudes, and number of events calculated through global or regional 1D velocity models. The ISC Bulletin, the ANSS ComCat, and the Global Centroid Moment Tensor (gCMT) catalogs provide easily accessed hypocenter and moment tensor solutions. The ISC is the most complete global collection of local through teleseismic phase onset times, including depth phases, and associated hypocenters with magnitudes. The ISC also provides a relocated, reduced uncertainty catalog called the reviewed ISC Bulletin derived using the EHB method (Engdahl et al., 1998). The reviewed Bulletin is generally published 2 years in arrear (Bondar and Storchak, 2011), however, making it difficult to study mainshock-aftershocks sequences in near real-time. In contrast, rapid publication of phase data used by the USGS National Earthquake Information Center and published in the ANSS ComCat allows real-time analysis but contains only primary first arrivals and no depth phases. In both catalogs, location procedures use global 1D velocity models (ak135 for the revised ISC bulletin and iasp91 for the ComCat). Significant spatial biases on the order of tens of kilometers in epicenter and depth have been shown to result from unmodeled structure within subduction zones (e.g., Engdahl et al., 1998; Storchak et al., 2000; Syracuse and Abers, 2009 and references therein). Additionally, teleseismic earthquake catalogs can have hypocenters shifted toward dense network coverage, such as toward land networks for offshore subduction-related events, and the natural trade-off between depth and origin time in the inverse solution can result in significant depth uncertainty, especially when depth phase times are not available. For the relocation study of the Ecuador-Columbia margin, we link the reviewed ISC and rapid ComCat
phase catalog data via differential times, and when available, correlated waveforms extracted from the Incorporated Research Institutions for Seismology Data Management Center (IRIS DMC) allow us to produce higher-accuracy differential time measurements. For initial hypocenters, the reviewed ISC bulletin takes precedence over the ComCat; the transition between catalogs takes place in late 2013 for our dataset.

The gCMT catalog (http://www.globalcmt.org/) provides moment tensor information, usually plotted as a beachball centered on the moment centroid, and for large magnitude earthquakes, the moment centroid may not correspond to the hypocenters reported by ISC and ComCat. Uncertainty in moment tensor catalogs has been explored by others (e.g. Frohlich and Davis, 1999; Kagan, 2003 and references therein), and we show only well-constrained CMT solutions following the criteria from Frohlich and Davis (1999) (Figure 3.2, bottom row). The gCMT catalog of shallow thrust earthquakes provides supplemental information about the subducting slab geometry. Shallow dip thrust mechanisms with trench-parallel striking clusters in the Galera, Jama and Manta segments are generally consistent in dip with the global Slab2.0 model (Hayes, 2018) (Figure 3.2). In central Ecuador, a cluster of normal faulting events associated with intermediate depth earthquakes (150-250 km) is also prominent in the dataset (Figure 3.2e).

The Ecuador Seismic Network (IG-EPN) provides earthquake monitoring for the region (Font et al., 2013; Alvarado et al., 2018). The IG-EPN seismic catalog extends from the early 1990s to present (e.g. Beauval et al., 2013), and the 2016 mainshock-aftershock sequence was well recorded; see Table 3.1 for IG-EPN hypocenters of 2016 Mw 7.8, 6.7 and 6.9 events. Local to regional earthquake location uncertainties from dense seismograph networks such as the IG-EPN can be an order of magnitude smaller than the uncertainties in the teleseismic catalogs.
(Wyss et al., 2011). However, the accuracy of arrival times, unmodeled velocity structure such as subducting slabs, and by limited azimuthal station coverage affects uncertainty in hypocenter determination at all scales. The IG-EPN catalog should provide more accurate epicenter determination in absolute space relative to teleseismic catalogs but may still exhibit spatial bias for offshore earthquakes due to the land-based station coverage. In this study, I use the IG-EPN catalog to assess the quality of the revised DD catalog.

**Table 3.1.** Comparing event locations of the 16 April 2016 mainshock and 18 May aftershocks between catalogs

<table>
<thead>
<tr>
<th>Catalog</th>
<th>Origin Time (UTC)</th>
<th>Latitude (°N)</th>
<th>Longitude (°W)</th>
<th>Depth (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td><strong>Mainshock (Mw 7.8, 16 April 2016)</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>USGS ComCAT</td>
<td>23:58:36.98</td>
<td>0.38</td>
<td>79.92</td>
<td>20.6</td>
</tr>
<tr>
<td>DD Relocation</td>
<td>23:58:34.92</td>
<td>0.33</td>
<td>80.17</td>
<td>35.8</td>
</tr>
<tr>
<td>IG-EPN</td>
<td>23:58:34.31</td>
<td>0.31</td>
<td>80.12</td>
<td>17.4</td>
</tr>
<tr>
<td>Nocquet et al., 2016</td>
<td>23:58:33.00</td>
<td>0.35</td>
<td>80.17</td>
<td>17.0</td>
</tr>
<tr>
<td>gCMT Centroid</td>
<td>23:58:57.00</td>
<td>0.12</td>
<td>80.25</td>
<td>22.3</td>
</tr>
<tr>
<td></td>
<td><strong>Aftershock 1 (Mw 6.7, 18 May 2016)</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>USGS ComCAT</td>
<td>07:57:02.65</td>
<td>0.43</td>
<td>79.79</td>
<td>16.0</td>
</tr>
<tr>
<td>DD Relocation</td>
<td>07:57:02.09</td>
<td>0.38</td>
<td>80.03</td>
<td>40.7</td>
</tr>
<tr>
<td>IG-EPN</td>
<td>07:57:00.43</td>
<td>0.43</td>
<td>80.01</td>
<td>17.2</td>
</tr>
<tr>
<td>gCMT Centroid</td>
<td>07:57:08.10</td>
<td>0.43</td>
<td>80.04</td>
<td>27.5</td>
</tr>
<tr>
<td></td>
<td><strong>Aftershock 2 (Mw 6.9, 18 May 2016)</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>USGS ComCAT</td>
<td>16:46:43.86</td>
<td>0.49</td>
<td>79.62</td>
<td>29.9</td>
</tr>
<tr>
<td>DD Relocation</td>
<td>16:46:41.38</td>
<td>0.39</td>
<td>79.88</td>
<td>43.3</td>
</tr>
<tr>
<td>IG-EPN</td>
<td>16:46:42.47</td>
<td>0.47</td>
<td>79.82</td>
<td>20.8</td>
</tr>
<tr>
<td>gCMT Centroid</td>
<td>16:46:49.20</td>
<td>0.30</td>
<td>80.02</td>
<td>33.6</td>
</tr>
</tbody>
</table>
Figure 3.2. Earthquake and centroid catalogs from 1961 through 15 April 2016 (left column: a,c,e) and from 16 April through 31 December 2016 (right column: b,d,f). Top row: the combined ISC and ComCat catalog; Middle row: DD relocations (this study); Bottom row: Global Centroid Moment Tensor solutions. Color coding indicates depth. Contour interval for the Slab2.0 model (Hayes, 2018) follow Figure 3.1. Right column also shows the slip contours (red at 1 m intervals) for the 2016 Mw 7.8 earthquake (earthquakes.usgs.gov, 2016) for reference.
3.2.2 Teleseismic Earthquake Relocation

Broadly, improvements to teleseismic catalog locations in South America incorporate the use of global 3D velocity models to remove large-scale bias due to unmodeled velocity structure, waveform cross-correlation to improve arrival time accuracy, and/or use of differential time data to constrain relative hypocenters (e.g., Rietbrock and Waldhauser, 2004; Waldhauser and Schaff, 2007; Pesicek et al., 2014). The double-difference (DD) relocation and tomography method has been widely used in a variety of settings and scales to produce higher-resolution images of seismicity via improved clustering and relative location to illuminate fault zone structures and/or slip patterns that were not apparent in absolute catalog locations (e.g., Waldhauser and Ellsworth, 2000; Zhang and Thurber, 2003; Waldhauser and Schaff, 2007; Pesicek et al., 2010; Diehl et al., 2013). The further improvements in the relative locations can be achieved using precise differential arrival times via waveform cross-correlation (CC) (Waldhauser and Schaff, 2008). In the DD relocation procedure, minimizing the residuals between observed and calculated differential travel times yields improved relative locations between neighboring earthquakes (event pairs) (Waldhauser and Ellsworth, 2000). The program teletomoDD is a modified version of DD tomography (Zhang and Thurber, 2003) than incorporates absolute time data and 3D velocity structure with the differential time data adapted for global and regional teleseseimic phases (Pesicek et al., 2010; Pesicek et al., 2014). TeletomoDD uses a 3D velocity model and the pseudobending (PB) method (Um and Thurber, 1987) for ray tracing through a spherical Earth model with discontinuities (Koketsu and Sekine, 1998). The global P-wave perturbation model MITP08 from Li et al. (2008) serves as the reference static 3D Earth model. For Ecuador, a regional model derived from the MITP08 re-gridded to 0.7° spacing is nested with the 3D global model and extends from -4.5 °S to 4.5 °N latitude, -84.7 °W to -75.5 °W
longitude. We test location sensitivity to velocity by holding the regional model fixed and by conducting single iteration joint inversion for a new regional compressional wave model following the procedure described in Pesicek et al. (2014). Table A.1 and Figure A.1 provides the optimal damping parameters, and trade-off curves used to derive those parameters, for location only and joint inversion including velocity.

To build a catalog of differential times using teleseismic data, we use the phase onset times of first arrivals for individual earthquakes accessed through publicly available earthquake catalogs. The reviewed ISC Bulletin provides a unified and consistent teleseismic dataset that includes the reexamination of phases, residuals and focal depth solutions (Bondar and Storchak, 2011). Within the study area, the reviewed ISC Bulletin reports 1512 events from 07-28-1961 to 12-31-2013. The ComCat data from 01-01-2014 to 11-19-2016 reports 229 events. The combined catalog includes a total of 1741 events, including 141 reported aftershocks following the 2016 mainshock (Figure 2, top row). Catalog phase data includes P, Pn, Pg, S, Sn, and Sg arrivals (primary phases) and pP (depth phase, ISC only). Catalog traveltimes from the ISC and ComCat and catalog-based differential time data (CTDT) calculated for events within a 100 km hypocentral separation for primary phase arrivals and 150 km separation for depth phase arrivals are used in inversion for the revised catalog (Table B.1).

Differential time data is also calculated using CC of phases recorded on broadband stations archived with the IRIS DMC. For CCDTs, we apply cross-correlation on P, Pn and pP phases reported in the teleseismic catalog using a filter passband of 0.75 to 2 Hz within the Geophysical Institute of Seismology Matlab Objects (GISMO) toolbox (Reyes and West, 2011). Only CCDTs associated with CC coefficients above 0.80 for events with an event separation of 150 km are retained. These thresholds are based on manual review of data. There is
significantly less S data relative to P data for use in teleseismic DD location (Table B.1). The linkage between the ISC and ComCat reported phases in our catalog shows that 15% of the total number of stations are reported in both catalogs. For cross-correlation data, 25% of differential times are linked between an ISC and ComCat event pair. 801 ISC and 219 ComCat events contained cross-correlation data.

In summary, the absolute arrival times, CTDTs and CCDTs for primary and depth phases are independent sets of datatypes used in the DD inversion process. These data types vary in quantity (Table B.1) and quality. Therefore, we apply a hierarchical dynamic data weighting scheme to the inversion as similarly described in other DD studies (e.g. Waldhauser and Ellsworth, 2000; Pesicek et al., 2010). For all inversion iterations, the P-wave times are always upweighted relative to the S-wave data (Table B.2). The initial iterations of the inversion update the locations and velocities primarily by absolute arrival time data. Subsequent iterations involve the CTDT data controlling the inversion. In the final iterations, the CCDT data provide the most control in determining the final hypocentral locations.

3.2.3 InSAR Observations and Modeling

A coseismic InSAR image represents the range distance difference measured in the satellite's line-of-sight (LOS) direction between two or more synthetic aperture radar (SAR) images acquired before and after the earthquake. This difference in LOS range is sensitive to the surface deformation on the order of the radar wavelength and is represented by the relative change in the interferometric phase on a pixel-by-pixel basis. As a result, InSAR has the capability to measure ground surface displacements with sub-centimeter level precision. Descending track passes (heading angle: -168.04 degree) from the Sentinel 1-A sensor map the ground surface deformation due to the Pedernales mainshock and the large aftershocks. The
Sentinel 1-A satellite is a C-band sensor operating at a wavelength of 5.6 cm with a ~20 (azimuth) x 5 (range) m spatial resolution (De Zan and Guarnieri, 2006). The incidence angle of the acquisitions is ~ 33.9° at the center of the image swath.

The Sentinel-1A satellite has an orbital repeat time of 12 days and acquired SAR images after the mainshock and between the two large aftershocks on 18 May 2016. As a result, the 18 May Sentinel-1A acquisition provides a SAR image that allow us to separate the two aftershock ruptures. Previous studies used the Sentinel-1A and the ALOS-2 satellites data of the Pedernales mainshock and reported peak LOS displacements of 60-70 cm (Nocquet et al., 2017; He et al., 2017). Funning and Garcia, 2019 analyzed the Sentinel 1-A InSAR pair for the Mw 7.8 mainshock and Mw 6.7 aftershock, but not the Mw 6.9 aftershock, and argued that the rainforest vegetation causes expected InSAR decorrelation in the region.

For consistency, we process deformation images for all three earthquakes using the same processing technique. SAR images acquired on 12-April-2016 and 24-April-2016 produce the interferogram for the Pedernales mainshock (Figure 3.3a). For the 18 May events, the SAR image acquired on 18-May-2016 at 11:00:7.73 UTC paired with the 06-May-2016 and 30-May-2016 SAR image produce the interferograms for the Mw 6.7 (7:57:2.65 UTC) and the Mw 6.9 (16:46:43.86 UTC) events, respectively (Figure 3.3b-c). The April 12/24 InSAR pair has a perpendicular baseline of -17.59 m, whereas the May 6/18 and May 18/30 InSAR pairs have a perpendicular baseline of -76.17 m and 59.61 m. Due to the small perpendicular baseline less than 100 m, the spatial decorrelation might not be significant, but the dense vegetation in tropical climate attributed to the low coherence (<0.2) of Sentinel-1A interferograms required a strong spatial filtering (Goldstein and Werner, 1998) to increase coherence. We remove topographic phases using the precise orbit ephemerides (https://qc.sentinel1.eo.esa.int) and the 1-arcsec (~30
m) global SRTM digital elevation model (DEM). The spectral diversity method provides precise coregistration of Sentinel-1A to avoid the discontinuous interferometric changes between bursts and sub-swaths (Yague-Martinez et al., 2016). We confirm that the primary interferometric phases in all interferograms do not contain significant atmospheric effects by checking existing weather forecast models (i.e., European Center for Medium-Range Weather Forecasts (ECMWF)’s European Reanalysis (ERA)-Interim, Generic Atmospheric Correction Online Service (GACOS)) and assessing the effects of the tropospheric delay on all the generated interferograms (Bekaert et al., 2015; Yu et al., 2017a, 2017b). Residual phase ramps (due to baseline error or large-scale atmospheric artifact) in the original interferograms covering a large region are removed using a second order polynomial fitting assuming that the far-field deformation is negligible. We crop out a portion of the original interferogram over the epicenter area for further modeling.

We model the observed interferograms to verify the fault geometry and coseismic slip distribution. The most common approach for inverting earthquake interferograms is to numerically model a single rectangular dipping fault with uniform slip in an elastic half-space (Okada, 1985) by minimizing the misfit between the observations and model predictions from a least squares approach to find the optimal model parameters. Optimization approaches like these can lack the quantification of uncertainties associated with the model parameters. The model uncertainties are important to characterize when similar inversion results can yield different model parameters especially if the model is highly nonlinear. Therefore, we use a Bayesian approach to solve the inverse problem because the formulations of posterior probability density functions (PDFs) provide statistical meaning on the model parameters. The Markov Chain Monte Carlo (MCMC) algorithm with the Metropolis sampling approach is used to find the PDFs of
physics-based source parameters (e.g., Hastings, 1970; Mosegaard and Tarantola, 1995; Anderson and Segall, 2013).

Interferograms containing a significantly large number of data points (in the millions) can limit the speed and efficiency of the modeling process. An effective way to reduce data points without losing resolution is to apply a two-dimensional quantization of the dataset known as quadtree (e.g., Jonsson et al., 2002). The quadtree partitioning starts with the dataset divided into four quadrants and the calculation of quadrant mean. If the rms scatter about the quadrant mean exceeds a variance threshold, the quadrant is subdivided into four subdivisions with associated means and subdivision continues until all quadrants meet threshold criteria. For quadtree analysis of earthquake deformation, the densest squares should sample the region nearest to and incorporating the surface projection of the fault (Figure B.3).

We use the Geodetic Bayesian Inversion Software (GBIS) to model the 2016 mainshock and the May 18th aftershock interferograms (e.g. Bagnardi and Hooper, 2018). This recently developed software follows a Bayesian approach using the MCMC algorithm and quadtree data sampling to invert for surface deformation from a variety of analytical models. For application to earthquake sources, we apply the Okada model for a dipping fault with uniform slip. Further information about GBIS, quadtree and PDF results appears in the supporting information (Figure B.3, B.4, B.5, B.6).
Figure 3.3. Wrapped Sentinel 1-A interferograms of the (a) 16 April 2016 mainshock (April 12/24 InSAR pair) and the (b) M_w 6.7 (May 6/18 InSAR pair) and (c) M_w 6.9 (May 18/30 InSAR pair) aftershocks of 18 May 2016.
3.3 Results

3.3.1 Teleseismic DD Locations

The merging of two teleseismic catalogs, inclusion of an updated 3D velocity model and use of differential time data to constrain relative location results in significant hypocentral changes in the Ecuador teleseismic catalog (Figure 3.2 and Figure 3.4). We compared relocation through the global DD model (DD-only) with relocation solutions using joint inversion (DD-tomo). For the Ecuador-Columbia shallow subduction zone, the \textit{MITP08} model does not image a dipping high velocity anomaly indicative of a subducting slab. Our tomographic inversion (Figure 3.4), plotted relative to \textit{ak135}, does image dipping velocity anomalies that trend parallel to the Slab2.0 earthquake-derived slab. Tomographic inversion results in significant residual reduction from 0.55 root mean square (rms) residual time for DD-only to 0.28 rms for DD-tomo, but changes in hypocentral parameters appear minor (Table B.2). Within the seismogenic zone, here defined as extending from trench to \textasciitilde 100 km, hypocentral changes were \textasciitilde 8 km in epicenter and \textasciitilde 3 km in depth, which are on par with the absolute uncertainties reported for the Sunda subduction zone (10, 11 km latitude, longitude and 7 km depth) (Pesicek et al., 2010). Both DD catalogs showed similar spatial patterns in seismicity, and we present the DD-tomo catalog as preferred for discussion.

There are significant hypocentral changes between the DD catalog and the original ISC and ComCat teleseismic catalogs. First, the median epicentral difference records a shift of the seismicity \textasciitilde 26 km to the southwest (Figure 3.4a, Table B.2). The large cluster of intermediate depth earthquakes biases catalog comparisons, so we separate the changes to those reflecting the full catalog and changes reflecting the shallow subduction zone (0-100 km depth). Within the shallow subduction zone, the systematic shift is \textasciitilde 25 km to the southwest (Table B.2). The DD
locations appear more tightly clustered, as expected for a relative location procedure, especially in the Galera and Jama seismic regions (Figure 3.2). It is also apparent that many aftershocks occur in the same seismicity clouds defined by earthquakes occurring prior to 2016 (Figure 3.2), a finding noted in previous studies of the 2016 Perdernales sequence (e.g., Nocquet et al., 2017; Vaca et al., 2018).

The depth resolution of the ISC and ComCat is poor, as reflected in the large set of fixed 10 km depth solutions for shallow events (Figure 3.4 and Figure 3.5). The DD hypocenters are deeper than the ISC and ComCat catalogs by a median ~21 km (median absolute deviation 10 km), and in the shallow subduction zone of ~15 km (median absolute deviation 10.5 km). The DD locations form an east dipping feature that is consistent with the Slab2.0 model at shallow depths (<100 km), and down to 250 km depth, the DD catalog defines a more smoothly dipping slab than Slab2.0 (Figure 3.4). However, the derived regional velocity model contains slow velocities below 100 km that need to be further explored (Figure 3.4) but is left for a future study. Interpretation and exploration of uncertainty focuses on the shallow subduction zone above 100 km depth.

*TeletomoDD* uses a Least Squares QR (LSQR) inversion technique that precludes the calculation of formal location uncertainties. We use a bootstrap approach to estimate uncertainty by running 50 inversions that each have 10% of the phase and differential time data for each event randomly removed. The difference in epicenter location (km), focal depth (km) and origin time (s) reflects uncertainty in location due to phase information but does not reflect absolute location uncertainty due to velocity model. Location changes due to use of different regional 3D velocity models were discussed previously. We take the standard deviation of the location differences for each event (full data relocated event differenced from 50 relocations with 10%
data randomly removed) and compute the mean of those events to derive an estimate of relative location uncertainty for the entire DD catalog (Table B.2). The mean change in epicenter is ~2 km, in depth is 1.8 km and in origin time is 0.37 s. Within the seismogenic zone (above 100 km depth), the mean change is 1.30 km in epicenter, 1.33 km in depth and 0.24 s in origin time. These values best estimate the relative location uncertainty in the dataset in areas of highly clustered seismicity and do not reflect absolute location uncertainty for events not well-linked through differential time data.

Network locations from the IG-EPN (http://www.igepn.edu.ec/solicitud-de-datos, last accessed 09/04/2017) provide events from 2012 through 2016 for comparison with the teleseismic catalogs. We discard fixed event depth solutions (10 km) from the IG-EPN catalog and match IG-EPN events to the ISC/ComCat (214 events) and DD catalog (208 events) by associating in space and requiring origin time within 60 seconds. Both the ISC/ComCat and DD catalogs have a median epicentral difference of ~20 km relative to the IG-EPN catalog (Table S3). Depth differences between the catalogs are larger (Table B.2). Within the seismogenic zone, Figure 2.5b illustrates that the DD locations are able to define the location of the slab and show that the seismicity delineates a dipping feature parallel to the Slab2.0 model (~20° to the east) and consistent with the dip of gCMT solutions across the Galera, Jama and Manta clusters rather than the sub-horizontal trends notable in the IG-EPN and ISC/ComCat catalogs (Figure 3.5). Relocated shallow seismic clusters locate near the bottom and top of the Slab2.0 subduction interface dipping 20°-25° east, similarly, proposed in Font et al. (2013) for the plate interface.
Figure 3.4. Comparison of the teleseismic ISC/ComCat catalog (black circles) and the DD catalog (white circles). The top epicentral map highlight the ~25 km southwest shift in epicenters relative to the oblique convergence (arrow) of the Carnegie Ridge. Red bars link initial and relocation to highlight systematic trends. Cross-section locations are taken perpendicular to Slab2.0 depth contours (20 km interval, negative numbers indicate depth to top of the slab) that crosses two distinct seismic clusters on the megathrust (A-A’ and B-B’) showing; Cross-sections of earthquake catalogs (circles), Slab2.0 top of the slab (green) and regional velocity perturbation relative to AK135 model.
Figure 3.5. (a) Epicentral comparison of in common ISC/ComCat (purple diamonds), DD (yellow circles) and IG-EPN (brown hexagon) catalog events. The 2016 M\textsubscript{w} 7.8 mainshock and M\textsubscript{w} 6.7 and M\textsubscript{w} 6.9 aftershocks are outlined in red, blue and green respectively. (b) Cross-sections for the Galera, Jama and Manta seismic regions is shown comparing the three catalog locations with respect to the Slab2.0 model (dotted line). gCMT solutions are shown at centroid location.

3.3.2 InSAR Models

The three model interferograms generally fit the data quite well and the small residual deformation (Figure 3.6) mostly reflects tropospheric delay and does not contain earthquake signals. Inherent with the data, both the mainshock and aftershock 1 can only resolve a one-sided InSAR lobe, whereas aftershock 2 can be resolved with a complete lobe. Although the interferograms are only limited to onshore observations, we invert for uniform slip on a fault and calculated moment for the three events based on their InSAR model (Table 3.2).

InSAR slip information is summarized in Table 3.2. The optimal solution for the mainshock defines a 79 km x 45 km fault slip area (Figure 3.6a-c). Relative to the gCMT and USGS slip model, the InSAR mainshock strike is 16°-17° more northward and 15° to 20° steeper (Figure 2.6b). Although there are discrepancies in the fault geometry between the InSAR and
seismic solutions, the NNE strike direction and thrust fault dip range (< 45°) are similar. The InSAR derived fault depth (26.5 km) reflects the depth to the bottom of the modeled fault, which is 4-6 km deeper than the center of slip. For aftershock 1, the uniform slip area is resolved inland on a shallow, 49 km x 39 km fault (Figure 3.6d-f), and when compared to the gCMT solution, the strike is 10° north and 11° shallower. The seismic moment derived from InSAR corresponds to a Mw 6.59 coseismic event, underestimated compared to the gCMT derived Mw 6.7. Of the three InSAR models, the optimal fault area for aftershock 2 (Figure 3.6g-i) is located farthest from the coast and has the smallest fault width (50 km x 33 km). The InSAR derived seismic moment corresponds to an Mw 6.6 event, underestimated compared to the gCMT Mw 6.9, but still larger than the first aftershock. The optimal fault strike is 10° NNE and ~19° dip to the east. The InSAR fault strike for aftershock 2 is similar to the mainshock and oriented 18° north from the gCMT strike. The dip however is close to the gCMT dip (21°).

The fault area was the most problematic parameter to constrain for the inversion of aftershock 2 and a minimum fault length of 50 km was set in order to derive a considerable moment larger than aftershock 1. Both aftershocks have underestimated InSAR derived moments, contradictory, InSAR studies of several earthquakes that have shown a slight tendency for InSAR derived moments to be smaller than those reported in the gCMT catalog, attributed to low signal-to-noise ratios, substantial off-shore deformation, and small surface deformation due to earthquake depth (Weston et al., 2011).

The relocated teleseismic (DD) and local (IG-EPN) epicentral locations of the mainshock, aftershock 1 and aftershock 2 agree well with each other and fall within the InSAR derived slip areas (Figure 3.6). The ComCat epicenters lie outside the fault area for the mainshock and aftershock 2 and within the slip area of aftershock 1. The InSAR fault depths for
the three events are more consistent with the ComCat hypocenter and gCMT centroid depths but our DD results place these events about 10 to 15 kms deeper.

**Figure 3.6.** Unwrapped InSAR line-of-sight observations (left column), unwrapped synthetic model interferogram (middle column) and model residual interferogram (right column) for the mainshock (a,b,c), aftershock 1 (d,e,f) and aftershock 2 (g,h,i). For each event, the ComCat (purple star), DD (yellow star) and IG-EPN epicenters (brown star) are shown with the gCMT centroid (red beachball). The DEM elevation model is plotted under the interferogram. All earthquakes (open circles) occurring between the SAR acquisition times for each interferogram are also shown. Dashed boxes represent the surface projected fault area of the best-fitting dislocation plane from co-seismic InSAR modelling and the thickened edges represent the shallow end of the fault plane.
3.4 Discussion and Conclusion

3.4.1 Teleseismic Locations Along the Ecuador-Columbia Subduction Zone

One goal of this study is to improve teleseismic event catalogs in order to better define the geometry and extent of the seismogenic zone where local data are not available. The uncertainty calculations via bootstrap for the Ecuador DD catalog assume the 3D velocity model correctly reflects large-scale heterogeneity, such as subducting lithosphere. The global MITP08 (Li et al., 2008) model contains fast velocity, subducting slab under Peru and Chile but not in the Ecuador region. Instead, the $V_p$ perturbation calculated relative to ak135 indicates slow velocities where a fast subducting slab is expected and likely reflects a lack of earthquake data prior to 2008. The revised single iteration regional velocity model calculated in this study, shown in Figure 2.4 plotted relative to ak135, does contain faster velocities within the shallow seismogenic zone and provides a more realistic 3D velocity model under Ecuador, but large areas of slow velocity present in the starting MITP08 model remain. Future work to further improve the tomographic images for the Ecuador subduction zone are required to yield more accurate absolute depths. However, the general agreement with DD catalog dip with the expected slab structure of the subducting Nazca plate supports the bootstrap-derived low relative uncertainty values.

The ~25 km systematic shift to the southwest in the DD catalog (Figure 2.4) is possibly due to azimuthal bias in station coverage to the north and east. Globally, observations from the coastal inland region and North America provide dense observations for earthquakes in Ecuador from the north and east direction. Earthquake catalogs can be biased toward dense station coverage, such as offshore earthquakes being pulled landward due to stations located on land and not offshore. Such spatial bias in global earthquake catalogs have been documented in other
subduction zones (e.g. Syracuse and Abers, 2009). The epicentral agreement between the DD epicenters for the three largest earthquakes, InSAR solutions, and relocation using local seismic stations (Nocquet et al., 2017) supports the epicentral shift noted in the DD catalog. It is also of note that the systematic shift occurs when either the global MITP08 model or the tomography derived regional velocity model is used, suggesting that any model that contains some slab structure yields this change.

### 3.4.2 Interseismic Earthquakes

Recent studies provide new interpretations on the rupture histories of the 1906 to 2016 great megathrusts. The 1942, 1958 and 1979 rupture areas are inferred to be neighboring each other and cumulatively overlap with the 1906 rupture extent, a characteristic termed “the Ecuador supercycle” by Nocquet et al. (2017). However, Yoshimoto et al. (2017) estimated the slip distribution of the 1906 event (corresponding to a $M_w$ 8.4) to be exclusively located up-dip near the trench, an area behaving aseismically with recent afterslip and slow slip events triggered by the 2016 mainshock (Rolandone et al., 2018). Their study concluded that the 1906 rupture did not overlap with the deeper 1942, 1958, 1979 and 2016 rupture areas and/or asperities. Yi et al. (2017) suggested that the 1942 and 2016 Ecuador events, similar in size and location, did not overlap, whereas other studies (Ye et al. 2016; Nocquet et al. 2017; He et al. 2017; Yoshimoto et al. 2017) suggest they did.

Prior to the 2016 $M_w$ 7.8 Perdernales earthquake, interseismic seismicity indicates persistent seismic clusters along the megathrust (Figure 2.2). These strong clusters appearing in the Galera, Jama and Manta seismic regions have been well observed before the Pedernales event (Font et al., 2013). Teleseismic events recorded since ~1960 show that these clusters located off-shore central to northern Ecuador are predominately characterized by alignments of
north-northwest (perpendicular to trench) trending seismicity with notable seismic gaps or small clusters in between. The DD epicenters confirm that the seismic clusters terminate near the coast, particularly for the Galera and Jama region, and that onshore interseismic cycle earthquakes are more diffuse. The Pedernales mainshock slip occurred within the less seismically active area bounded by the Galera and Jama clusters.

### 3.4.3 Large Megathrust Earthquakes and the 2016 Pedernales Sequence

Observations of overlapping seismic asperities can reflect persistent frictional or mechanical fault properties of the megathrust that can be reused over multiple seismic cycles. Geodetic observations of asperities, or strong coupling, just prior to the 2016 earthquake were imaged via interseismic locking (coupling) in Chlieh et al. (2014) and Nocquet et al. (2017) (updated model from Nocquet et al., 2014). Both sets of plate locking models feature similar asperity regions relative to locations of historic earthquakes. Nocquet et al. (2017) suggest that propagation of large earthquake ruptures in this region does not reach near the trench. In this discussion, we focus on the rough model (minimum coupling) over the smooth model (maximum coupling) presented in Chlieh et al. (2014) as the former offers evidence of several discrete asperities (up to 7) whereas the latter tends to average neighboring asperities, resulting in up to 3 main asperities in the model. In the vicinity of the Carnegie ridge subduction region, we plot the rough coupling model (also referred here as the Chlieh model) and identified four highly coupled asperity regions (A1-A4) where plate coupling is >0.4 (Figure 3.7b). We also show the coupling model from Nocquet et al. (2017) as contour intervals for comparison.

Both the 1942 epicenter (Mendoza and Dewey, 1984) and the 2016 mainshock are located off-shore within asperity A3. The coseismic InSAR deformation shows that the mainshock ruptured on asperity A3 (Figure 3.7). If the rapidly calculated USGS mainshock
(finite-fault) slip model (earthquakes.usgs.gov, 2016) were located relative to the DD location rather than the initial ComCat mainshock location, slip moves off-shore and the down-dip edge agrees better with the InSAR deformation data (Figure 3.7a). This again confirms the DD epicenter is more accurate in space. The off-shore slip extent matches spatially to the fault area from InSAR modeling and aligns better with the local data derived mainshock slip distribution and epicenter from Nocquet et al. (2017).

The 18 May aftershock (aftershock 1 and 2) epicenters are located northeast from the mainshock (Figure 3.7). Also, the InSAR models show that the two largest aftershocks both ruptured northeast from the northern-east edge of the mainshock rupture in a narrow seismic gap between the 1958 and 1942/2016 rupture regions. In the Chlieh model, a distinct highly locked asperity region (A2) is located on the peninsula within the Galera region. There is small overlap between the asperity region A2 and the 18 May aftershock InSAR fault region. Chlieh et al. (2014) noted this asperity and considered a future large earthquake in that area, but it is now regarded as the aseismic Punta Galera-Mompiche Zone (PGMZ) as referred to in Vaca et al. (2018). The Nocquet and Chlieh models indicate that the asperity region A2 is moderately to highly coupled where the two largest Pedernales aftershocks occur in an unstable seismic slip region abutting next to the off-shore PGMZ.

The afterslip hosted by slow slip events in the PGMZ region is exclusively located off-shore such that we expect the three earthquake interferograms in this study to be unaffected by afterslip. The Pedernales aftershocks show that strong (magnitude 6.0 - 6.9) seismic events present considerable hazard for coastal to inland regions in Northern Ecuador and their interactions between the aseismic slip region might reveal processes that lead to the next large
earthquake. Currently, asperity regions A1 and A3 have hosted the locations of large megathrust earthquakes.

In the teleseismic catalog, the Pedernales aftershocks occur mostly within the Galera and Jama seismic clusters that bounds the entire segment of the mainshock rupture instead of occurring within or surrounding the slip asperities. Seismicity after the May 18th aftershocks continue to occur within those interplate segments as well but also to the northeast from the Galera cluster, separate from the mainshock area. The teleseismic locations shown in this study highlight the spatial relationships between the great $M_w > 7.5$ epicenter locations and aftershock seismicity in relation to interplate coupling, revealing that great earthquake rupture in Ecuador is segmented from the generation of moderate to smaller earthquakes.

**Table 3.2. Summary of InSAR and Seismic Source Models**

<table>
<thead>
<tr>
<th>Model</th>
<th>Longitude</th>
<th>Latitude</th>
<th>Strike</th>
<th>Dip</th>
<th>Length</th>
<th>Width</th>
<th>Depth</th>
<th>Slip</th>
<th>Moment ($\times 10^{20}$ N*m)</th>
<th>Mw</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mainshock-USGS</td>
<td>-79.926</td>
<td>0.352</td>
<td>26</td>
<td>16</td>
<td>-</td>
<td>-</td>
<td>20.6</td>
<td>-</td>
<td>7.05</td>
<td>7.8</td>
</tr>
<tr>
<td>Mainshock - gCMT</td>
<td>-80.25</td>
<td>-0.12</td>
<td>27</td>
<td>21</td>
<td>-</td>
<td>-</td>
<td>22.3</td>
<td>-</td>
<td>5.93</td>
<td>7.8</td>
</tr>
<tr>
<td>Mainshock - InSAR</td>
<td>-80.1208</td>
<td>-0.04</td>
<td>10.3</td>
<td>36.3</td>
<td>79</td>
<td>45</td>
<td>26.5</td>
<td>5.05</td>
<td>5.74</td>
<td>7.81</td>
</tr>
<tr>
<td>Aftershock 1 - gCMT</td>
<td>-80.04</td>
<td>0.43</td>
<td>28</td>
<td>18</td>
<td>-</td>
<td>-</td>
<td>27.5</td>
<td>-</td>
<td>0.141</td>
<td>6.7</td>
</tr>
<tr>
<td>Aftershock 1 - InSAR</td>
<td>-79.64</td>
<td>0.437</td>
<td>18</td>
<td>7</td>
<td>49</td>
<td>38</td>
<td>14.8</td>
<td>0.143</td>
<td>0.085</td>
<td>6.59</td>
</tr>
<tr>
<td>Aftershock 2 - gCMT</td>
<td>-80.02</td>
<td>0.3</td>
<td>28</td>
<td>21</td>
<td>-</td>
<td>-</td>
<td>33.6</td>
<td>-</td>
<td>0.253</td>
<td>6.9</td>
</tr>
<tr>
<td>Aftershock 2 - InSAR</td>
<td>-79.69</td>
<td>0.233</td>
<td>10</td>
<td>19.2</td>
<td>50</td>
<td>33</td>
<td>25.0</td>
<td>0.215</td>
<td>0.114</td>
<td>6.67</td>
</tr>
</tbody>
</table>
Figure 3.7. (a) DD teleseismic catalog locations and coseismic InSAR observations for the northern Ecuador margin. Seismicity before the 2016 mainshock (black circles). 2016 seismicity sequence locations are indicated by the following: 16 April mainshock and aftershocks before 18 May events (red), 18 May events for aftershock 1 and following (blue) and aftershock 2 and following (green). The 1942 and 1958 epicenters (Mendoza and Dewey, 1984) are shown (magenta diamonds). The 2016 mainshock coseismic interferogram is shown in the copper to black gradient. The InSAR fault area of the mainshock, aftershock 1 and 2 are outlined in red, blue and green respectively and the stars show their DD epicentral locations. The mainshock slip model (red 1 m contour lines) from the USGS is plotted relative to the relocated DD epicenter (red star), as discussed in the main text. (b) The interseismic coupling from Chlieh et al. (2014) is illustrated in the yellow to dark red scale indicating low to high coupling (1 is 100% coupling). Four distinct regions of high interplate coupling are noted (A1-A4). Black contour lines (20% coupling interval) and cyan contour lines (1 m slip interval) indicate the coupling model and mainshock slip model values respectively from Nocquet et al. (2017) respectively. The relocated DD mainshock location agrees with the epicenter from Nocquet et al. (2017) (cyan star). The InSAR fault areas and associated DD event locations are shown in white. The DD seismicity are shown as unfilled circles.
3.4.4 Conclusion

Unique to this study, we show that surface displacements detected by the Sentinel 1-A satellite was imparted by two large aftershocks, but precise quantification of slip and moment is difficult to resolve owing to unmodeled errors in the interferogram. The InSAR data were here used to confirm DD catalog locations and explore depth uncertainty. The InSAR results, however, show that large aftershocks were focused to the northeast of the mainshock along areas of the megathrust fault that had moderate to higher coupling, but had failed to slip during the 2016 Mw 7.8 Pedernales earthquake. InSAR observations provide slip locations for the 18 May aftershocks where seismic slip models are not available. Our result confirms previous studies showing that InSAR can contribute to assess the quality of global seismic earthquake catalogs (e.g. Weston et al., 2012).

We present a high-resolution teleseismic catalog of relative DD hypocenter locations (1961-2016) along the Ecuador subduction margin that exhibit epicentral shifts on average 25 km southwest. The shift, while large, yields teleseismically derived locations that are consistent with local seismic network solutions and better match InSAR derived deformation data. The DD catalog shows tighter alignment of persistent event clusters along the megathrust over the seismic cycle and reproduces many of the segmentation features of the Ecuador subduction zone being reported for the Pedernales sequence using local seismic data (Font et al., 2013; Rolandone et al., 2018). The DD relocations confirm that the 2016 mainshock epicenter occurred off-shore, similar to Nocquet et al., (2017), rather than the near-on-shore location provided in the ComCat, for example. The DD catalog relative locations that link the ComCat rapid locations to the reviewed ISC historic catalog better reproduce the slab dip required by gCMT data and the InSAR data presented here, suggesting that the relative relocations are improved relative to the
standard rapid global catalog. This study affirms that using differential times that link new
seismicity to established catalogs can yield important improvements to absolute locations in
rapidly produced earthquake catalogs. Improvements to the Ecuador DD catalog, especially in
depth, can be calculated in the future through improved 3D regional velocity models constrained
by local seismic catalogs (i.e., Pontoise and Monfret, 2004; Beauval et al., 2013; Font et al.,
2013) and by incorporating the seismic data obtained by the international rapid response team
(e.g., Font et al. 2016).
REFERENCES


CHAPTER 4

INTRASLAB EARTHQUAKES

The last two chapters focused on understanding deformation in the oceanic lithosphere from shallow-focus earthquakes that occur at depths less than 70 km. This chapter will focus on deep-focus earthquakes below 70 km that exist uniquely within most down-going slabs of subducted lithosphere on Earth. The mechanisms of these deeper earthquakes are not yet fully understood, but their existence provides direct observations to investigate the dynamics of intraslab deformation. The genesis of deep-focus earthquakes within subducting slabs (intraslab) remains enigmatic compared to shallow-focus earthquakes on the plate interface or in the overriding plate. Intraslab earthquakes occur as deep as ~660 km, and at these great depths lithostatic pressure is assumed to be too large to allow earthquakes to be generated by normal frictional processes (e.g., Houston, 2015). Still, deep-focus earthquakes are a unique global feature on long-lived subduction zones, regardless of slab dip, convergence velocity and thermal parameter. Deep-focus intraslab earthquakes are observed in two distinct depth domains. Intermediate-depth earthquakes generally begin at 70 km depth and continue to ~300 km depth. The frequency of deep-earthquakes increases from ~550 km, rates peak around 600 km depth and the maximum depth reaches 660 km.

Transformational faulting in the mantle-transition zone from 410 to 660 km depth is a popular model and hypothesis to explain the mechanisms of deep earthquakes (e.g., Kirby, 1987). For intermediate-depth earthquakes, dehydration embrittlement of hydrous minerals under
conditions of high pressure and temperature has been a promising hypothesis to explain a mechanism to decrease frictional strength within the slab and generate slip (Green and Houston, 1995; Hirth and Guillot, 2013; Houston, 2015; Meade and Jeanloz, 1991; Yamasaki and Seno, 2003). Supporting evidence for this hypothesis come from studies that argue that a significant amount of water is contained in subducted slabs through the hydration of the crust and mantle at the incoming plate region. Bending of the pre-subducted incoming plate creates extensional faults inferred to be viable pathways for seawater to enter the crust and lithospheric mantle. Regional seismic reflection and wide-angle refraction datasets collected offshore Costa Rica and Alaska show a spatial correlation between well-developed outer-rise faults, slab hydration indicated by reduced seismic velocities, and increased intraslab seismicity rates (Ranero et al., 2005, Shillington et al., 2015).

As shown in Chapter 3, intermediate-depth earthquakes do occur within the Ecuador subducting slab, but mainly appear in spatially distinct clusters rather than smoothly along dip or strike. I use joint relocation and tomography to relocate these earthquakes and interpret deep slab structure in the region. I compare these interpretations in Ecuador to a global study of intermediate depth seismicity, derived by using the ISC Bulletin to analyze seismicity rate, magnitude of completeness and b-value estimates. This latter work represents an extension of my calculations provided for the global summary presented recently in Boneh et al. (2019). The combined analyses of this chapter explore possible explanations for the observed heterogenous spatial distribution and/or clustering of intraslab seismicity evidenced in the historic record of intermediate depth earthquakes in the ISC Bulletin.
4.1 Intraslab Seismicity and Tomography of the Ecuador Subduction Zone

Intermediate-depth earthquakes are distributed in central to southern Ecuador from -1° to -3° latitude and reflect subduction of two tectonic plates of different rheology and age. In this region, intermediate depth earthquakes exclusively occur on the subducted Farallon plate (e.g. Yepes et al., 2016) (Figure 4.1). Slab-pull stresses is hypothesized to be the driving force for the separation of the Farallon plate in the beginning of the Miocene that in turn created the Nazca plate, as seen today, and the plate complexity noted under Ecuador (Lonsdale, 2005). The Grijalva rifted margin marks the boundary between the Farallon and Nazca plates, and the inactive rift margin shares a parallel subduction trajectory of the Carnegie ridge, which has been entering the subduction zone for the past 3-6 Ma (Gutscher et al., 1999). In the USGS Slab2.0 model of subducting slab shown in Figure 4.1, the bend to the south-west in the depth contours indicate the Nazca/Farallon divide. Note that this is also near a highly clustered region of intermediate depth earthquakes, the so-called El Puyo cluster, where normal mechanisms are interpreted as indicating flexural bending within the downgoing slab (Yepes et al., 2016). In contrast, intermediate-depth earthquakes are largely absent on the younger subducted Nazca plate, consistent with global observations of young down-going plates (Syracuse and Abers, 2006).

I present tomographic images and relocated seismicity to investigate the complex structure of the oceanic lithosphere subducted beneath Ecuador. As in Chapter 3, earthquake absolute and differential-time data are jointly inverted to image velocity and details in seismic structures. Pesicek et al. (2014) showed that multiscale tomography incorporating absolute and differential-time data from the local, regional and teleseismic distances provides improved resolution of subducting lithosphere and earthquakes within the slab, on the plate interface and
off-shore; that study focused on the Maule, Chile, region of South America. The 2016 Pedernales earthquake sequence was the most well-recorded earthquake in the Ecuadorian portion of the same subduction zone, and thus provides a dataset that can improve the resolution of subducting slab structure. The global P-wave tomography model MITP08 (Li et al., 2008), published before the Pedernales events, does not clearly image a fast velocity perturbation expected for subducting oceanic lithosphere and improvement in velocity is necessary to reduce spatial bias in absolute space.

I perturb an embedded regional up-sampled version of the MITP08 model (0.7° grid spacing, -4.5 °S to 4.5 °N latitude, -84.7 °W to -75.5 °W longitude) and leave the MITP08 down-sampled global version of the model fixed. The teletomoDD joint relocation and tomography procedure begins in the first iteration by updating the velocities and locations using the absolute data. In the second iteration, inversion for velocity perturbation is turn off and hypocenter relocation uses the first iteration regional velocity model. The two-step process continues over 6 additional iterations. I regularize the inverse problem by applying independent damping parameters to the global and regional velocity models and the location terms following Pesicek et al. (2014). I test damping parameters using single teletomoDD iterations to the location and regional velocity terms, and find 100 and 2500 to provide optimal damping (Figure 4.2). For the global velocity term, I select a damping parameter of 700. Table 4.1 shows a summary of the weighting and damping parameters. An alternative approach to select damping and weighting using multiple iterations remains as future work.

I present tomography results from the teletomoDD study described in Chapter 3 (DD-tomo model) by comparing velocity perturbations of the MITP08 and DD-tomo models relative to the 1-D AK135 model (Figure 4.3). As noted previously, intermediate-depth seismicity is
prevalent in central to southern Ecuador compared to the north. I illustrate this difference by showing tomographic and seismicity cross-sections (1) across the Pedernales mainshock area (A-A’), (2) across the Jama cluster and intermediate depth seismicity (B-B’), and (3) across a more continuous band of shallow to intermediate depth seismicity in the southern region (C-C’). The MITP08 model shows velocity perturbations up to 4%, whereas the DD-tomo model show much larger perturbations up to 14% relative to ak135. Below the intermediate depth seismicity, velocity significantly increases relative to ak135, whereas in the MITP08 model the region appears as a minor, slow perturbation.

The seismicity across A-A’ contains a small number of events deeper than 100 km depth. At intermediate depths, there are no fast velocity perturbations beneath the slab, defined here by the Slab2.0 depth contours, but rather a high amplitude slow velocity anomaly appears. In contrast, both regions with many intermediate depth events (B-B’ and C-C’) image fast velocity anomalies below the seismicity and the velocity feature connects to the trench, as expected if the feature represents the subducting slab. However, in both regions, the amplitude of the feature relative to ak135 varies. At the top of each cross-section, I note the boundary location between a major upper plate division noted in the literature: the North Andean forearc Sliver (NAS) and South American Plate (SAP) are separated by a large shear zone. Strong velocity anomalies appear below this boundary, but the pattern and location are not consistent for each of the three regions. The regions with intermediate-depth seismicity, however, do show fast anomalies starting below the NAS/SAP boundary location which is also near the inferred Nazca and Farallon plate divide. The fast anomalies appear to be “sinking” down to near 700 – 800 km depth below the Slab2.0 model.
The El Puyo earthquake cluster at -1° to -2° latitude has been associated with plate contortion and bending of the Farallon slab (Yepes et al., 2016). Seismicity south of the El Puyo cluster is more diffuse, and in cross-section C-C’, the shallow and intermediate-depth earthquakes define a smooth, continuous slab. Yepes et al. (2016) argues that the lack of intermediate depth earthquakes in the subducting Nazca plate reflect that the younger, hotter slab cannot efficiently generate earthquakes. Similarly, the relatively young and hot subducting Juan De Fuca slab along Cascadia does not generate high rates of true intermediate depth earthquakes. Due to resolution issues, it remains unclear as to whether the large velocity perturbations imaged in the DD-tomo model near the slab can be linked to temperature, age or density differences between the subducting Nazca and Farallon plate or if the velocity anomalies reflect bias due to an increased density in raypaths because there are earthquakes.

The along-strike differences in velocity, subducting plate age and intermediate depth seismicity rates remain a tantalizing target for future research. Pesicek et al. (2014) cautioned that DD data with closely spaced events in between wider grid node separation can provide only little additional velocity constraint. In Figure 4.1, the velocity grid node spacing used here is in fact larger than the seismicity spacing. Future work is required to understand the effect of grid spacing and differential-time data selection on the resulting velocity image. In order to understand the uncertainty of the model, synthetic recovery and checkerboard tests could illustrate the solution non-uniqueness. Bootstrap and jackknifing statistical analysis could also be used estimate the model uncertainty.
Figure 4.1. DD locations (colored by depth) and velocity grid nodes (red squares) in north-central Ecuador. Straight black lines indicate cross-sections (lines A-A', B-B', and C-C') across the subduction zone shown in Figure 4.3. Plate boundary locations (green line) include the Nazca plate, the North Andean Sliver and the South American Plate. Slab2.0 (black contour lines, 20 km depth interval) shows the top of the subducting slab. Blue dashed lines represent the general boundaries of the Carnegie ridge. Red dashed line indicates noticeable bend of the slab, coinciding with the inferred Grijalva rifted margin (Yepes et al., 2016) and the boundary between subducted Nazca and Farallon lithosphere.
Table 4.1. TeletomoDD single iteration joint relocation and tomography weighting and damping parameters

<table>
<thead>
<tr>
<th>WTCCP</th>
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<th>WDC</th>
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* WTCCP, WTCCS: weight cross-correlation P, S data
* WTCTP, WTCTS: weight catalog P, S data
* WRCC, WRCT: residual threshold in sec for cross-correlation, catalog data
* WDCC, WDCT: max distance [km] between cross-correlation and catalog linked pairs
* DAMP, D2, D3: damping for relocation, regional, and global parameter
Figure 4.2. TeletomoDD damping trade off curves. (a) Damping values versus absolute variance for first teletomoDD iterations. (b) Damping values for regional tomographic model showing the data variance versus the model variance between the MITP08 (Li et al., 2008) starting model and updated tomographic model. I chose damping values of 100 (relocation), 2500 (regional tomography) and 700 (global or non-joint inversion regional parameters) (see Table S2).
Figure 4.3. Compressional wave velocity perturbations relative to the 1D AK135 reference model. MITP08 (left panel) and DD-tomo (right panel) are shown along cross-sections A-A’, B-B’ and C-C’ (see Figure 4.1 for line locations). DD locations are colored following Figure 4.1. Slab2.0 (orange line) indicates top of the slab. The green bar indicate boundary between the North Andean Silver and the South American Plate.
4.2 Global Observations of Intermediate-depth Seismicity

I became involved in a graduate student led research group on intraslab intermediate depth earthquakes formed at the 2017 Cooperative Institute for Dynamic Earth Research (CIDER) Summer Program on “Subduction Zone Structure and Dynamics”. In supporting the assumptions that the temperature and pressure conditions in the slab enable the release of fluids during dehydration embrittlement, our work sought to investigate whether the amount of incoming plate faulting is a mechanical parameter that influences the seismicity rate of intraslab intermediate depth earthquakes.

The results of our study are presented in Boneh et al. (2019), who argues that maximum fault throw (MFT) measurements can be used as a proxy for fault damage zone and hydration state on the incoming plate by showing a positive global correlation between fault throw and intermediate-depth seismicity rate. Measurements of MFT and intermediate-depth earthquake rates were sampled on several subduction segments, including the Aleutians, Cascadia, Japan, Java, Middle America, Marianas, South America and Tonga (Figure 4.4). Information about the multibeam bathymetry and seismic-reflection datasets used to measure MFT is summarized in Boneh et al. (2019). My contribution to the study involves gathering a historic to recent record of reported intermediate depth seismicity that spans the extent of the incoming plate faulting dataset for different subduction zone segments and calculating a seismicity-rate value for each. This chapter will exclude comparisons of the seismicity-rate with the MFT data done by co-authors of the Boneh et al. (2019) study. Instead I add additional analyses to the catalogs regarding magnitude of completeness and b-value calculations and explore links to spatial variation trends or presence of fluids.
The reviewed ISC catalog contains a consistent database of ISC authored hypocenter locations (http://www.isc.ac.uk/iscbulletin/review/). Events that occurred between 1964-2015 (reviewed bulletin ~2 years behind date accessed: 06 March 2018) with focal depth 70 – 300 km, were the initial criteria to retrieve a global intermediate depth earthquake catalog. As described in Boneh et al. (2019), the incoming plate faulting data for each segment correspond to a trench length and location, the number of intermediate depth earthquakes occurring perpendicular to the trench length. The selected earthquakes for each segment are shown in Figure 4.4. The number of earthquakes is divided by the trench length in kms and time span of the catalog (51 years) to derive a seismicity rate in (km^{-1}*year^{-1}).

Seismic events reported in the ISC catalog for 50 + years result in different magnitude threshold and location detection capabilities over time. Di Giacomo et al. (2015) found that the ISC-GEM Reference Global Instrumental Earthquake Catalogue is complete down to magnitude 5.6 starting from 1964. As a result, they define a single magnitude of completeness (M_c) for the ISC catalog. M_c is defined as the minimum magnitude that can be spatially and temporally detected, theoretically at a 100% chance. I additionally calculate a magnitude of completeness value for each of the 16 subduction zone segments, and derive a seismicity-rate to compare with results using one magnitude threshold value. I use the maximum curvature method to calculate M_c for each segment. Wiemer and Wyss (2000) showed that maximum curvature tends to provide a good estimate of M_c. The M_c estimate is the maximum value of the highest first derivate slope from a frequency-magnitude distribution. Using the same frequency-magnitude distribution, I also calculate the b-value using the maximum likelihood method (Aki, 1965; Utsu, 1965). The magnitude frequency-distributions indicating M_c and b-value estimates are shown in
Figures 4.5, 4.6, 4.7, 4.8. A summary of the new $M_c$ and b-values and seismicity rates using $M_c$ and $m_b \geq 4.5$ and 5.6 (Boneh et al., 2019) cutoffs are shown in Table 4.2.

The Japan and the North Marianas segment have the lowest $M_c$ (3.8) and the Central Marianas segment has the highest (4.8). This indicates great variability of $M_c$ in the Marianas including the Southern segment ($M_c=4.2$). Except the Central Marianas and Northern Peru ($M_c=4.7$) segments, all other segments have $M_c \leq 4.5$. The lowest $M_c$ segments had the lowest b-values (0.69-0.76) and the highest $M_c$ segments had the highest b-values (1.25). The presence of fluids and high pore-pressure is hypothesized to increase the b-value (i.e., Wiemer and Wyss, 1997), but here, b-values seem to deviate regionally and globally from 1 throughout several subduction segments. I test to see if there is a similar linear correlation between the seismicity rate and b-value estimates. However, I find no linear correlations between b-value and the seismicity rate using magnitude cutoffs $M_c$ and $m_b \geq 4.5$ and $m_b \geq 5.6$ (Figure 4.9). The linear correlation between intermediate-depth seismicity rate and increased throw on incoming plate bending faults found in Boneh et al. (2019) is interpreted to reflect a mechanical control on the extent of hydration.

4.3 Conclusion

The Ecuador study and global study for regions with incoming plate faulting indicate different controlling factors of intermediate depth earthquakes. The presence of two different subducting slabs under Ecuador manifests in the observed seismicity spatial extent and rates. The older and more contorted Farallon plate exhibits high clustering and rates of intermediate depth normal faulting seismicity, while the young Nazca plate lacks the prevalence of intermediate depth seismicity, as noted by Yepes et al. (2016). Interestingly, initial velocity modeling indicates a similar spatial variation in subducting slab velocity and within the upper
plate, but additional testing of resolution is required before velocity perturbations can be definitively linked to regional tectonics. Dehydration embrittlement remains the most likely control on intermediate depth earthquake occurrence within subduction zones but may be modulated by incoming plate faulting and other complex plate interactions that change the availability of fluids within the system.

Figure 4.4. (from Boneh et al., 2019) (supporting document, Figure S1) Subduction zone segments and intermediate-depth earthquakes from the ISC catalog. Black boxes delineate area used to calculate seismicity-rate.
Figure 4.5. Cumulative (black squares) and non-cumulative (white triangles) frequency-magnitude distribution with $M_c$ and b-value fit for South American segments 1-4.
Figure 4.6. Cumulative (black squares) and non-cumulative (white triangles) frequency-magnitude distribution with $M_c$ and $b$-value fit for the Aleutians West and East, Tonga and Javas segments.
Figure 4.7. Cumulative (black squares) and non-cumulative (white triangles) frequency-magnitude distribution with Mc and b-value fit for the Marianas segments and Kuril segment.
Figure 4.8. Cumulative (black squares) and non-cumulative (white triangles) frequency-magnitude distribution with $M_c$ and b-value fit for the Middle American trench (MAT), Mexico/Guatemala and Japan segments.
Table 4.2. Seismicity-rates using different magnitude thresholds and b-value estimate.

<table>
<thead>
<tr>
<th>Subduction Segment</th>
<th>Seismicity Rate <em>1000</em>km-1*year-1 for Mb &gt;= 4.5</th>
<th>Seismicity Rate <em>1000</em>km-1*year-1 for Mb &gt;= 5.6</th>
<th>Magnitude of Completeness Mc</th>
<th>Seismicity Rate <em>1000</em>km-1*year-1 for Mc</th>
<th>b-value</th>
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<tr>
<td>Aleutians(East)</td>
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<td>Aleutians(West)</td>
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<td>0.00</td>
<td>0.000</td>
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<tr>
<td>Japan(Tohoku)(all)</td>
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<td>0.681</td>
<td>3.80</td>
<td>38.681</td>
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</tr>
<tr>
<td>Java</td>
<td>10.000</td>
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<td>16.081</td>
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<td>Kuril</td>
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<td>Middle America Trench</td>
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<td>1.010</td>
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<td>North Mariana</td>
<td>24.510</td>
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<td>3.80</td>
<td>59.819</td>
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<td>Central Mariana</td>
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<td>0.237</td>
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<td>South Mariana</td>
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<td>Mexico/Guatemala</td>
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<td>4.20</td>
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<tr>
<td>S.A1(N.Peru)</td>
<td>17.020</td>
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Figure 4.9. b-value vs. seismicity-rate of intermediate depth events with (a) $m_b \geq 4.5$ (b) $m_b \geq 5.6$ and (b) $M_c$ estimates. Intraslab segments include: Aleutians, east (green circle), Aleutians, west (green diamond), Cascadia (green circle), Java (light blue), Kuril (pink circle), Middle American Trench (purple circle), Mariana, north (purple circle), Mariana, central (purple diamond), Mariana, south (purple square), Mexico (red circle), North Peru (orange circle), South Peru and North Chile (orange diamond), Chile (orange square), South Chile (orange triangle), Tonga (yellow circle).
REFERENCES


CHAPTER 5

CONCLUSION

The relocation studies presented in this dissertation show that the use of a 3D velocity model and differential-time data can shift teleseismic earthquake hypocenters 10s of kilometers. The ISC Bulletin and ComCat contain a significant percentage of shallow events with fixed absolute hypocentral depth solutions (10 km, 20 km, 35 km depth). Within global catalogs, fixed depth solutions result from single event location using sparse teleseismic data and 1D models where convergence to a free depth solution cannot be found. Such artifacts make using the full catalogs to delineate the dip and seismogenic extent of the fault more difficult and can created biased images of seismogenic structures in absolute space. Incorporating differential times, depth phases and linking rapid solutions, such as the Perdionales aftershocks, to historic catalogs that contain depth phase data, overcame the fixed depth problem and resulting in more free depth solutions. The improve locations, in turn, can be used to assess finer details regarding seismogenesis across a range of domains spanning incoming oceanic plate, through the subduction megathrust, to intermediate depths within subduction systems.

The spatial distribution of relocated seismicity presented in Chapter 2 and Chapter 3 show highly heterogenous seismic clusters and segmented fault regions that participate in the earthquake rupture process for the Wharton Basin and Ecuador subduction zone. I highlight the key points of the two studies:
Chapter 2: Wharton Basin Earthquakes study:

- Aftershocks are located away from or at edges of significant coseismic slip consistently across multiple faults
- The depth distribution of aftershocks shows complete lithospheric rupture agreeing with coseismic slip models
- The 600°C isotherm provides a robust limit on the depth extent of faulting in the Wharton Basin

Chapter 3: Ecuador Subduction zone study:

- High resolution teleseismic double difference tomography shifts earthquakes tens of kilometers southwest along the Ecuador seismogenic zone
- The revised catalog better models satellite derived deformation data and confirms persistent megathrust segmentation over seismic cycles
- The 2016 M6.7 and M6.9 aftershocks of the M7.8 Pedernales event ruptured northeast of the mainshock in areas of moderate to high coupling

In Chapter 4, intermediate-depth seismicity along Ecuador was compared to upper plate and subducting plate major tectonic features. The dataset was compared to global comparison conducted as part of a unique collaborative study arising from the 2017 CIDER workshop. Key points in this Chapter are:

- Intermediate depth earthquakes in Ecuador are prevalent on the Farallon plate in an area with concentrated bending stresses and plate contortion
- A correlation between b-value and intermediate-depth seismicity rate cannot be found globally
• Dehydration embrittlement or regional intraslab stress partitioning provide explanations of the existence of intermediate depth earthquakes

The DD locations from Chapters 2 and 3 show that aftershock seismicity (Mb > 3) tend to occur outside regions of high coseismic slip and can occur far outside the boundaries of the mainshock rupture area. In both cases, this definition of aftershock is not limited to earthquakes occurring after the largest event and on the same fault as the mainshock. For the Wharton basin case, the oceanic plate responds to stress perturbations from a series of earthquakes dating back to the M9.2 2004 Andaman Islands - Sumatra earthquake and activates a complex conjugate system of faults that extend rupture into the oceanic mantle. The Pedernales event, while in many respects a common if large subduction megathrust earthquakes, exhibits a intersecting combination of aseismic slip, co-seismic slip, large aftershocks extending rupture north and downdip of the original earthquake, and post-seismic deformation. I will be interested to monitor how quickly and where the seismogenic portion of the megathrust begins to re-accumulate strain and re-enters the interseismic phase of the seismic cycle.

The 2016 Pedernales mid-magnitude aftershocks occurred primarily along previously existing seismic clusters in the background, interseismic activity. The event provided the opportunity to link historic events from the reviewed ISC Bulletin with the aftershock sequence reported in the USGS ComCat. Here, I purpose that the implementation of teleseismic DD relocation methods linking the reviewed ISC Bulletin and real-time USGS catalog can improve the rapid assessment of aftershock locations and identify spatio-temporal patterns to yield improved seismic and tsunami hazard forecasts. Steps for implementing teletomoDD in near-real time would require automating the retrieval of seismic phase and event information from
teleseismic catalogs. Also automating first iteration trial-runs for determining optimal damping parameter for various regions of the planet and exploring alternative 3D velocity models to assess uncertainty would be necessary. In this dissertation, the most time-consuming and computationally intensive procedure would be the waveform cross-correlation used to derive (a small subset of) the differential-time dataset. Jones (2014) combined the GISMO MATLAB toolbox with the IRIS DMC waveform retrieval services to semi-automate the cross-correlation of teleseismic waveforms from information in the absolute phase catalog. I used the same procedure but a detailed study of how to implement and make such a system efficient would be needed.

Finally, the next generation satellite radar missions coming online over the next 5 years carry the potential to improve images of earthquake rupture via inversion of measured surface deformation. The InSAR study applied to the Perdanes, Ecuador, earthquakes provided additional rupture constraints and proved to be a complimentary dataset to validate seismic source locations. The highly vegetated region of Ecuador introduced a loss of coherency issue to C-band Satellite observations but the L-band Satellite launch of the NASA-ISRO Synthetic Aperture Radar (NISAR) mission in 2021 should improve the resolution of InSAR images of dense foliage regions in the future.

In this dissertation I have compared teleseismic relocation catalogs with a range of complementary geophysical datasets, including finite-fault rupture models, InSAR-derived fault slip models, geodetically derived plate coupling models, half-space cooling models, and subduction geometry models. Joint interpretation of the datasets, anchored by a high-quality earthquake catalog, reveal how lithospheric structure inherent to creation and aging transition into complex deformation and seismogenic processes prior to subduction, and continue to affect
earthquake generation through the subduction system. I suggest future work jointly incorporating datasets such as high-rate continuous GPS and satellite-derived surface deformation measurements with seismic phase and waveform data from all distances can improve teleseismic catalogs. Refining the use of differential-time data in teleseismic tomography studies is also a future goal to improve 3D regional and global velocity structure is also a future, fruitful path forward. Ultimately, the work in this dissertation to improve earthquake catalogs and understand the subduction system holistically aim to improve real-time assessments of seismic and tsunami hazard.
REFERENCES

Jones K. (2014). Refining Fault Structure and Seismic Behavior Along the Kuril Islands, Russia, Using Teleseismic Double-Difference Relocation and High-Resolution Waveform Cross-Correlation. [1587779th]. Available from Dissertations & Theses @ Southern Methodist University; ProQuest Dissertations & Theses Global.
## APPENDIX A

### Table A.1. TeletomoDD weighting and damping parameters.

<table>
<thead>
<tr>
<th>NITER</th>
<th>WTCCP</th>
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* NITER: last iteration to use the following weights
* WTCCP, WTCCS: weight cross-correlation P, S data
* WTCTP, WTCTS: weight catalog P, S data
* WRCC, WRCT: residual threshold in sec for cross-correlation, catalog data
* WDCC, WDCT: max distance [km] between cross-correlation and catalog linked pairs
* WTDD: weight of damping
* DAMP, D2, D3: damping for relocation, regional, and global parameters
* JOINT: relocation only = 0, joint relocation and tomography = 1
* THRE: joint inversion residual threshold (s)
Figure A.1. Trade-off curve between damping values and absolute data variance after first iteration results in teletomoDD of the combined ISC and GFZ catalog (red circles). Damping value of 100 is chosen in this study.
Figure A.2. Histograms showing the distribution of location differences between the DD catalog and the DD catalog determined from boot-strap data resampling. A log-normal distribution is fitted for (a) the epicenter difference, and a normal distribution is fitted for (b) the logarithm of the epicenter difference. A normal distribution is fitted for (b) the focal depth difference and (c) origin time difference.
Figure A.3. DD epicenters with latitude and longitude error bars associated with the standard deviation of the location differences from the boot-strap dataset.
Figure A.4. Distribution of DD earthquake depths (km) vs origin time (monthly range). Vertical error bars are associated with the standard deviation of the event depth difference from the bootstrap dataset.
Figure A.5. Travel-time plots for observed pP-P time measurements with epicentral distance for earthquakes below the 600°C isotherm. The circles are events occurring between the 600°C and 800°C isotherms calculated from the half-space cooling model. The squares are events occurring between the 800°C and 1000°C isotherms. Each symbol represents a pP depth phase station observation and are colored by (a) relocated DD depth and (b) absolute catalog depth. Predicted pP-P travel time curves corresponding to event depth are also indicated (solid black lines).
Figure A.6. Earthquake density-map and ellipses showing earthquake clustering and orientation. Using the Spatial Statistics Toolbox in ArcGIS, a Kernel Density function was calculated using the DD catalog to calculate the magnitude-per-unit area shown where warmer colors denotes areas of high earthquake density. We chose 7 regions where there is high earthquake density (tight clustering) and use the Standard Deviation Ellipse feature in the Spatial Statistics Toolbox to calculate the direction distribution of the earthquake clusters.
APPENDIX B

Table B.1. Summary of *teleTomoDD* data input

<table>
<thead>
<tr>
<th>Data Description</th>
<th>Data Amount</th>
</tr>
</thead>
<tbody>
<tr>
<td>ISC Bulletin events</td>
<td>1492</td>
</tr>
<tr>
<td>USGS ComCat events</td>
<td>229</td>
</tr>
<tr>
<td>Stations</td>
<td>5690</td>
</tr>
<tr>
<td>Catalog P (S) times</td>
<td>127140 (3328)</td>
</tr>
<tr>
<td>Cross-Correlation P(S)</td>
<td>108586 (0)</td>
</tr>
<tr>
<td>Differential Catalog P(S)</td>
<td>5333913 (570)</td>
</tr>
<tr>
<td>Catalog pP times</td>
<td>7908</td>
</tr>
<tr>
<td>Cross-Correlation pP</td>
<td>499</td>
</tr>
<tr>
<td>Differential Catalog pP</td>
<td>13294</td>
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</table>
Table B.2. Median Location Differences Between Teleseismic and Local IG-EPN Catalog

<table>
<thead>
<tr>
<th></th>
<th>DDonly – DDtomo</th>
<th>ISC/ComCat - DDonly</th>
<th>ISC/ComCat - IGEPN</th>
<th>ISC/ComCat - DDtomo</th>
<th>IGEPN - DDonly</th>
<th>IGEPN - DDtomo</th>
</tr>
</thead>
<tbody>
<tr>
<td>Δepicenter (km)</td>
<td>7.55 (7.60)</td>
<td>30.58</td>
<td>19.43</td>
<td>25.97 (24.93)</td>
<td>21.32</td>
<td>18.78</td>
</tr>
<tr>
<td>Δdepth (km)</td>
<td>0.28 (3.1)</td>
<td>-22.86</td>
<td>5.74</td>
<td>-20.81 (-14.73)</td>
<td>27.16</td>
<td>19.03</td>
</tr>
<tr>
<td>Δtime (s)</td>
<td>-0.015 (-0.34)</td>
<td>-19.2</td>
<td>0.98</td>
<td>-2.18 (1.96)</td>
<td>1.68</td>
<td>1.89</td>
</tr>
<tr>
<td>Δrms</td>
<td>0.2 (0.18)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*Note: DDonly = Double-difference relocation only catalog, DDtomo = Joint Double-difference relocation and tomography catalog, ISC/ComCat = Absolute catalog locations from the ISC and USGS ComCat, IG-EPN = Local Ecuador earthquake catalog. Values within the parenthesis are for events located between 0-100 km depth.
Figure B.1. Histograms (blue bars) showing the differences in epicentral location (km), hypocentral depth (km) and origin time (s) between (top row) teleseismic ISC/ComCat catalog vs. DD locations derived using joint inversion including velocity model, (middle row) teleseismic catalog vs. IG-EPN locations (bottom row) DD locations derived using joint inversion including velocity model vs. IG-EPN locations. The red line shows the fitted normal distribution. Means for the full catalog comparison, shown here, are reported in Table S2.
**Figure B.2.** Bootstrap results shown as histograms of epicentral location difference (km), hypocentral depth difference (depth) and origin time difference (s) between the DD catalog and 50 perturbations, each having 10% of absolute and differential time data removed. The red line shows a fitted normal distribution. Mean and standard deviations are reported in the main text.
B.3 GBIS Modeling

For each of the InSAR dataset, the Geodetic Inversion Bayesian Software (GBIS) provides quadtree subsampling of the data on the unwrapped interferograms. Different threshold variance values for quadtree subsampling were chosen to fit the densest amount of quadtree squares within the assumed fault rupture locations and to eliminate over sampling for locations further away the fault. The preferred threshold variance for quadtree subsampling of the 2016 mainshock and aftershock 1 and 2 interferograms are $6.5 \times 10^{-5}$ m, $4.9 \times 10^{-5}$ m and $8.1 \times 10^{-5}$ m respectively. The quadtree sampling for the three interferograms is shown below. The number of data points of the new quadtree samples from the original number of pixels in the interferogram reduced dramatically. Quadtree reduced the number of data points for the mainshock, aftershock 1 and aftershock 2 from ~17 million to 478, ~9.5 million to 49 and 21 million to 96 respectively. We ran $1 \times 10^6$ simulations on each of the quadtree sampled interferograms. The results of the simulations for each model parameter can be summarized by their posterior density functions (PDFs) for each different source parameters (Figures B4-6). The optimal fault parameters are found by selecting the solution with the highest probability density or also known as the maximum likelihood solution. The PDFs provide statistical meaning that the errors in the data can be drawn from a zero-mean Gaussian distribution. The majority of model parameter simulations fit a normal distribution very well. The small spread in model parameter solutions indicates which fault parameters are well resolved. The strike solutions for aftershock 1 and 2 have a ~20° spread and was not as robustly constrained as the mainshock strike. Fault area was the most difficult to constrain for aftershock 2 and a 30 km length and width was used as the minimum threshold.
Figure B.3. Quadtree subsampled data points for (a) the mainshock, (b) aftershock 1 and (c) aftershock 2
Figure B.4. Mainshock posterior density functions showing histograms for the number of iterations vs value of each model parameter (blue bars). The optimal model parameter is shown (red line).
Figure B.5. Aftershock 1 posterior density functions showing histograms for the number of iterations vs value of each model parameter (blue bars). The optimal model parameter is shown (red line).
Figure B.6. Aftershock 2 posterior density functions showing histograms for the number of iterations vs value of each model parameter (blue bars). The optimal model parameter is shown (red line).