Marine electromagnetic experiment across the Nicaragua Trench:
Imaging water-rich faults and melt-rich asthenosphere

A dissertation submitted in partial satisfaction of the requirements for the degree
Doctor of Philosophy

in

Earth Sciences

by

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2015
The dissertation of Samer Nasri Naif is approved, and it is acceptable in quality and form for publication on microfilm and electronically:

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2015
DEDICATION

To my parents, Nasri and Nora, and to my sister, Dena.
I want to learn more and more to see as beautiful what is necessary in things; then I shall be one of those who make things beautiful. Amor fati: let that be my love henceforth! I do not want to wage war against what is ugly. I do not want to accuse; I do not even want to accuse those who accuse. Looking away shall be my only negation. And all in all and on the whole: some day I wish to be only a Yes-sayer.

—Frederich Nietzsche
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I wish to thank my advisor, Kerry Key, for his enthusiasm and unwavering support. Kerry was generous to allow me to work on the SERPENT dataset for my graduate studies. He has provided me with countless hours of his time. I will always cherish our conversations together, which have instilled in me a deep passion for geophysics and a positive outlook on life. I also thank my co-advisor, Steven Constable, who was kind enough to allow me to volunteer in his lab as an undergraduate, and subsequently recruited me as a graduate student. I would not be where I am today without his support and trust. With the combined powers of Kerry and Steve, I have had the privilege to attend conferences and give presentations in Australia, Germany, Indonesia, Japan, New Zealand, Nicaragua, and Singapore, and at many fine cities across the United States. They have opened many doors for me without reservation. Their generosity is far more than any one person deserves. I am forever grateful. I wish to thank the remaining members of my committee, Steven Cande, Michael Holst, and Peter Shearer, for taking time out of their schedules to discuss my work. I also thank collaborator Rob L. Evans. Their help has improved this dissertation.

It would be remiss of me to not acknowledge David Myer, a past graduate of the Marine Electromagnetic Laboratory. David has developed and selflessly shared a large repertoire of computer codes, which has benefited me greatly. I suspect I would have encountered many more sleepless nights without them, and I thank him sincerely for that. I also thank past graduate Brent Wheelock for providing me with his wisdom.

I wish to thank the many people I have had the privilege to call friends and colleagues during my time at Scripps. I thank my cohort in particular, which consists of Dylan Connell, Diego Melgar, Fernando Paolo, Robert Petersen, and Valerie Sahakian. I am very fond of the countless one-of-a-kind adventures we have experienced together. I am optimistic there are still yet many more to come. We just need to remember to bring the cookies for Diego. I also thank Andy Barbour, John Blum, Brendan Crowell, Chris Takeuchi, and Brent Wheelock. The details of our friendship and numerous tirades are better told face-to-face with a beer in hand. I thank Anthony Dominguez for keeping me tethered to reality. I thank Dave Stegman for his counsel, and for always treating my ideas with respect no matter how lacking. I thank Brandon Razooky, my brother
from another father and mother; together, our Pinky and the Brain aspiration are sure to become reality.

I wish to thank my partner, Joyce Shi Sim. She has stood by me without trepidation. Her youthful soul and endless encouragement were critical to my sanity. Most importantly, she introduced me to Penang laksa, mangosteen, and durian. Last, but certainly not least, I wish to thank my parents Nasri and Nora, and my sister Dena. I am amazed that they have put up with me for this long, and yet continue to do so. Their patience and love knows no bounds.

Funding was provided, in part, by the Seafloor Electromagnetic Methods Consortium at Scripps Institution of Oceanography, and by the National Science Foundation grant OCE-0841114. Thanks go to the captain (M. Stein) and crew of R/V Melville and the governments of Nicaragua and Costa Rica for permission to work in their exclusive economic zones. The following are thanked for their participation in the research cruise: C. Armerding, C. Berger, E. Carruthers, B. Cohen, J. Elsenbeck, T. Matsuno, D. Myer, A. Orange, J. Perez, K. Shadle, J. Souders, K. Weitemeyer, B. Wheelock and S. Zipper; J. Lemire and A. Jacobs are thanked for their efforts with cruise planning, mobilization and demobilization.

The text of Chapter 2, in part, is a reprint of the material as it appears in Naif, S., Key, K., Constable, S., and Evans, R. L. (2013). Melt-rich channel observed at the lithosphere-asthenosphere boundary. *Nature*, 495(7441), 356-359. The dissertation author was the primary investigator and author of the published work.

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The text of Chapter 4, in part, has been submitted for publication of the material as it may appear in Naif, S., Key, K., Constable, S., and Evans, R. L., *Geochemistry Geophysics Geosystems*, 2015. The dissertation author was the primary investigator and author of this paper. The text of Chapter 4, in part, is currently being prepared for submission for publication of the material. The dissertation author was the primary investigator and author of this material.
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PUBLICATIONS


SELECTED ABSTRACTS


ABSTRACT OF THE DISSERTATION

Marine electromagnetic experiment across the Nicaragua Trench: Imaging water-rich faults and melt-rich asthenosphere

by

Samer Nasri Naif

Doctor of Philosophy in Earth Sciences

University of California, San Diego, 2015

Professor Kerry Key, Chair
Professor Steven Constable, Co-Chair

Electromagnetic (EM) methods have been widely applied in exploration geophysics and to study tectonics for several decades. Electrical resistivity, or its reciprocal conductivity, is a physical quantity that varies by several orders of magnitude. Bulk resistivity is highly dependent on the presence of fluids and ore bodies. While EM is primarily used to map the geoelectrical structures of terrestrial environments, advances over the last two decades in instrument technology and computing software have not only made marine EM experiments viable but also routine and reliable. In this dissertation, I explore the utility of the marine magnetotelluric and controlled-source electromagnetic techniques for probing subduction zone processes.
In the Spring of 2010, the Scripps Institution of Oceanography Marine EM group ventured on the R/V Melville to conduct the Serpentinite, Extension and Regional Porosity Experiment across the Nicaragua Trench (SERPENT). Over the course of 28 days, 54 sites of broadband marine magnetotelluric (MT) and 800 km of marine controlled-source electromagnetic (CSEM) data were collected, culminating in the first CSEM survey and the largest marine EM dataset at a subduction zone to date.

In this dissertation, I perform regularized two-dimensional inversions on the marine MT and CSEM data from SERPENT to model the electrical resistivity structure of the crust and upper mantle. The MT data revealed an unexpected conductive channel at a depth interval of 45-70 km. I apply measurements from laboratory studies and find that only partial melt can account for the electrical signature of the conductor. I conclude that the anomalous channel is a sheared partial melt layer at the lithosphere-asthenosphere boundary that decouples the lithosphere from the deeper mantle.

The CSEM data image sub-vertical conductive channels that correlate with outer rise fault scarps, providing the first observation to confirm bending faults behave as fluid pathways. I use Archie’s law to infer porosity and find that the crust subducts significantly more pore water than previously thought. The CSEM data also image the complete subduction of the incoming sediments along the megathrust plate interface, providing the first large-scale estimates of porosity at the megathrust. At 20 km into the forearc, a conductive anomaly extends from the plate interface into the overlying crust beneath a high concentration of active seafloor seeps, possibly imaging both the origin and migratory pathway of fluids escaping along the margin seafloor. The location of the anomaly correlates with a section of the seafloor that exhibits steepened bathymetric slope, suggesting a sediment underplating mechanism as its cause.
Chapter 1

Introduction

The surface of the Earth is distinguished by its diverse and unique topography, whether on continental land masses or on the seafloor beneath the oceans. The continents display mountain ranges that stretch for several thousand kilometers (km) in length and tower as high as 8,800 meters (m) above sea level, while the opaque oceans conceal a widespread network of trenches that run as deep as 10,600 m. Plate tectonics, the scientific theory that elucidates the underlying mechanisms behind the geologic and topographic makeup of the Earth, was established just over 50 years ago with the discovery of mid-ocean ridges and the seafloor spreading fabric. In the years since, significant scientific progress has been made and the body of knowledge has grown exponentially. However, many fundamental questions have yet to be answered or remain contentious within the geoscience community.

Presently, a considerable number of research groups are devoted to the study of subduction zones, regions characterized by the convergence of two tectonic plates that lead to the formation of deep sea trenches and explosive volcanic arcs, such as the ‘Ring of Fire’ that borders the Pacific Ocean. Convergent margins are also responsible for the largest earthquakes ever recorded, some of which foment destructive tsunamis, including both the 2004 Bande-Aceh and the 2011 Tohoku-oki earthquakes. The last decade of natural disasters illustrates the destructive powers of subduction zones. In that regard, it is not only scientifically relevant but also societally beneficial to improve our understanding of the mechanisms that control such phenomena.

Much of what we know about subduction zones stems from a wide array of
geophysical, geochemical, and geological studies. The breadth of research consistently points to one overlapping ingredient: H$_2$O. Water plays a fundamental role in many coupled Earth systems, and particularly at subduction zones. The global deep-Earth water cycle is born out of the constant flux of material generated and destroyed at plate margins, where water is extracted from the mantle at plate spreading centers and recycled back into the mantle at subduction zones (Hirschmann and Kohlstedt, 2012). From its impact on seismicity to its capacity to promote mantle melting that later feeds arc volcanism, water is inextricably linked to the form of plate tectonics that is expressed on Earth. Advancing our knowledge of fluid-tectonic processes is essential to understanding the factors that govern a variety of phenomena at both active and passive margins, including the hazards associated with earthquakes, tsunamis, and volcanoes.

1.1 Probing the solid Earth with electrical resistivity

The central aim of this dissertation is to image the electrical resistivity of the Central American subduction zone offshore of Nicaragua with marine electromagnetic soundings. Electrical resistivity is an intrinsic material property that can help us to elucidate the geological and hydrological structure of the Earth’s interior. It is the inverse of electrical conductivity, which is the measure of a material’s ability to conduct electrical current. The resistivity of fluids decreases as a function of salinity. Because seawater is several orders of magnitude less electrically resistive than dry crystalline rock, the bulk resistivity of oceanic crust is largely dependent on porosity (Evans, 1994). In the mantle, bulk resistivity strongly depends on the presence of partial melt and the water content of nominally anhydrous minerals (NAMs) and to a lesser degree on temperature (Yoshino and Katsura, 2013). Hence, EM sounding not only has the capacity to image hydrogeologic structures, but it is also a powerful proxy for quantifying porosity, melt fraction, and the hydration state of the mantle.

Figure 1.1 shows the range of electrical resistivity for materials that comprise oceanic plates. A typical section of seafloor consists of a veneer of deposited sediment overlying basaltic crust (formed at a mid-ocean ridge) overlying peridotitic mantle.
Figure 1.1: A comparison of the electrical resistivity for various Earth materials. Resistivity is measured in units of Ohm meters ($\Omega \text{m}$), conductivity in Siemens per meter ($S/m$). Since seawater and partial melt are highly conductive, it is possible to estimate porosity and/or melt fraction with empirical and theoretical mixing models such as Archie’s law (Archie, 1942) or the Hashin-Shtrikman bounds (Hashin and Shtrikman, 1963).

Porosity controls the electrical signature in the sediments and crust. At greater depths, the oceanic lithosphere is effectively non-porous.

The upper mantle is primarily composed of peridotite, an ultramafic igneous rock that is highly resistive. At temperatures less than 1000°C, dry peridotite exceeds $10^4 \Omega \text{m}$. Thermal constraints place the mantle potential temperature (MPT) between 1300-1500°C (Hasterok, 2013). At the upper bound MPT, dry peridotite is 25 $\Omega \text{m}$ (Constable, 2006). More often than not, electromagnetic soundings of the oceanic upper mantle are too conductive to reasonably explain with high temperatures. One must invoke either interconnected partial melt or water dissolved in peridotite since both are mechanisms that reduce mantle resistivity (Yoshino and Katsura, 2013).

Electrical resistivity is remotely estimated with both passive and controlled-
source methods (e.g., Chave et al. 1991). During the past decade, the commercial application of marine magnetotelluric (MT) and controlled-source EM (CSEM) methods for hydrocarbon exploration has led to significant improvements in instrumentation and numerical software (e.g., Constable 2010). Reliable marine data acquisition with large 2D or 3D receiver arrays are now routinely performed for commercial applications (Myer et al., 2010; Key, 2012a). While 2D numerical solvers for marine MT data have been publicly available for more than a decade, the first and still only public 2D solver for marine CSEM data was only just released in 2013 (Key, 2012b). A generalized 3D EM code that can handle complex topography is currently being developed and should see the light of day within the next few years.

Seismic, thermal, geochemical, bathymetric mapping, and borehole drilling expeditions are extensively used to investigate fluid-tectonic processes in the offshore environment. These tools facilitate fundamental discoveries, but they lack the means to effectively and/or uniquely image subsurface fluids across large spatial scales (see Saffer and Tobin 2011). Electromagnetic soundings can fill that gap. Because electrical resistivity is highly sensitive to porosity yet insensitive to mechanical properties such as density and rheology, it offers valuable constraints that complement seismic methods.

1.2 Characteristics of subduction zones

Hidden from view beneath the oceans is a vast and continuous chain of seafloor mountain ranges, known as mid-ocean ridges, that stretch for over 65,000 km in length. The seafloor spreads in opposite direction away from the ridge axis and in its wake new oceanic crust is formed. As the crust spreads away from the ridge axis, it stretches as it cools and forms numerous ridge parallel extensional faults resulting in the ubiquitous abyssal hills fabric (Buck et al., 2005). The formation of abyssal hills may also be influenced by sea level oscillations caused by the Earth’s orbital cycles (Crowley et al., 2015; Tolstoy, 2015).

Subduction zones are formed by the convergence of two tectonic plates. Buoyancy forces thrust the denser lithosphere beneath the surface of its less dense sibling (Stegman et al., 2010). Since the lithosphere is rigid and behaves elastically, the down-
ward dip of the subducting slab causes it to bend under tension. The bending moment creates a bathymetric bulge seaward of the trench known as the outer rise, which is a characteristic feature of subduction zones (Turcotte and Schubert, 2014). The tensile stresses at the outer rise are large enough that the lithosphere experiences brittle failure, generating bending-induced normal faults that are visible as fault scarps on the seafloor (Levitt and Sandwell, 1995; Garcia et al., 2015). Although flexure at the outer rise is capable of generating new extensional faults, it is energetically more efficient to reactivate remnant abyssal hill faults formed near the ridge axis. The orientation of the abyssal hill fabric relative to the bending axis dictates whether new faults are formed or remnant faults are reactivated (Massell, 2002).

Sediments deposited on the seafloor are transported along with the subducting lithosphere. At a number of margins, some fraction of the sediments are accreted onto the overriding plate at the trench axis. Over geologic time scales, the accreted sediments accumulate to form an accretionary prism, which is marked by a gently sloping seafloor known as the continental slope. The seafloor gradually flattens and a shelf ensues. It is important to note that accretionary prisms are observed at less than half of the total length of convergent margins. The remainder of margins are non-accreting, where most (if not all) of the sediments are thrust beneath the overriding plate and the seafloor slope is steeper (Cloos and Shreve, 1988).

All subduction margins are characterized by a seismogenic zone. Subduction zones contain the largest low angle thrust faults ever observed at the boundary between sinking and overlying plates, appropriately given the name megathrust. The frictional properties at the plate interface are such that a confined segment of the megathrust becomes locked, accumulating strain that is eventually released during large to giant magnitude earthquakes. Interstitial and mineral bound water subducted in the sediments and crust alter the megathrust frictional properties by modulating the effective normal stress at the plate interface (Scholz, 1998), which has significant consequences for the behavior of seismogenic zones.

Another characteristic feature of subduction zones is a volcanic arc that runs parallel to the trench axis at 200-300 km into the margin. The top of the sinking slab is found at 125-175 km depths below the arc, where the corresponding pressures and
temperatures cause subducted minerals to undergo metamorphic dehydration reactions that release water into the overlying mantle wedge (Hacker et al., 2003). Because water reduces the melting temperature (Kushiro et al., 1968), the subducting slab promotes magma production and thus is the mechanism responsible for the volcanic arc.

1.3 The Nicaraguan subduction zone

The Nicaraguan portion of the Central American subduction zone is unique in that it has been found to be unusually rich in fluids by a multitude of data. Samples from its arc volcanoes contain the largest geochemical signatures found to date that link the erupting material to subducted sediments (Morris et al., 1990; Patino et al., 2000). Beneath the volcanic arc, seismic studies detect a velocity anomaly at depths coincident with the subducting oceanic crust, suggesting that the sinking slab is highly hydrated (Abers et al., 2003; Syracuse et al., 2008).

The Central American landmass is an uplifted block on the western portion of the Caribbean oceanic plate. It includes the Central America Volcanic Arc (CAVA), formed by the ongoing subduction of the Cocos oceanic plate beneath the western edge of the Caribbean plate (Mann et al., 2007). The border between the two plates is marked by the Middle America Trench (MAT).

The Cocos plate crust is born out of two independent mid-ocean ridges: the north-south trending East Pacific Rise (EPR) and the east-west trending Cocos-Nazca Spreading center (CNS). The boundary between the two origins of the Cocos oceanic crust runs southwest-northeast and meets the MAT off of Costa Rica’s Nicoya Peninsula (Barckhausen et al., 2001). Figure 1.2 shows the tectonic setting of Central America.

The distinction between crustal origin is important because the thickness, seismic velocity, and heat flow signatures of the EPR and CNS crust are starkly different. The EPR crust emits anomalously low heat flux, while the CNS crust shows typical heat flux for its age (Fisher et al., 2003a; Hutnak et al., 2008). The EPR crust is also thin (~5.5 km) and seismically fast relative to the ~8-11 km thick CNS crust (Ivandic et al., 2008; van Avendonk et al., 2011).

The incoming EPR-sourced oceanic plate begins to bend as it dives into the
Figure 1.2: The 280 km long SERPENT EM profile (boxed) crosses the Middle America Trench offshore of Nicaragua, where 24 Ma Cocos plate subducts beneath the Caribbean plate at a rate of 83 mm/yr (DeMets et al., 2010). The black dashed line is the boundary between the portions of the Cocos plate produced by the East Pacific Rise (EPR) and the Cocos-Nazca spreading center (CNS).

MAT, producing an outer rise flexural bulge that is cut by numerous normal faults, often referred to as bending faults. Their associated scarps are clearly seen in ship-track bathymetric maps. The scarps reach several hundred meters offset along the MAT, which indicates that fresh basaltic basement is exposed to seawater.

A seismic reflection survey off of southern Nicaragua imaged outer rise bending faults extending several kilometers into the upper mantle (Ranero et al., 2003). Out-
cropping basement highs on the EPR Cocos crust have been shown to recharge and
discharge fluids, fueling hydrothermal circulation in the relatively permeable upper
oceanic crust (Fisher et al., 2003b; Hutnak et al., 2008). Naturally, the observation of
fluid circulation through basement outcrops combined with the possibility that faults
generate wide damage zones suggests that exposed seafloor faults may also provide
permeable pathways for seawater (Peacock, 2001). This hypothesis is supported by
a numerical model that shows sub-hydrostatic pressure gradients arise from outer
rise extensional stresses, allowing seawater to percolate down to upper mantle depths
(Faccenda et al., 2009).

When H\textsubscript{2}O comes into contact with fresh ultramafic mantle at temperatures
below $\sim$450°C, it sparks a variety of chemical reactions that form hydrous alteration
minerals, particularly serpentinites (Moody, 1976). If bending faults promote the mi-
gration of seawater to Moho depths at the outer rise, a late-stage of serpentinization
will ensue. Serpentinite phases are unstable at higher pressures and temperatures,
undergoing dehydration reactions during subduction to supply a substantial flux of
H\textsubscript{2}O to the overlying mantle wedge (Hacker, 2008; van Keken et al., 2011). Since water
lowers the solidus of mantle minerals (Kushiro et al., 1968; Gaetani and Grove, 1998), outer
rise serpentinization could increase magma production by enhancing the flux of H\textsubscript{2}O
in the mantle wedge. The larger flux of magma ultimately increases volcanic activity in
the arc.

Serpentinites are significantly less dense and as such have lower seismic velocity
than their parent rock, peridotite (Carlson and Miller, 2003). Wide-angle seismic data
collected along the MAT detect velocity reductions in the upper mantle, indicating that
some amount of serpentinized mantle is likely present (Ivandic et al., 2008; van Avendonk
et al., 2011; Ivandic et al., 2010).

1.4 The Serpentinite, Extension and Regional Porosity Experiment across the Nicaragua Trench

In April-May 2010, Scripps performed the Serpentinite, Extension and Regional
Porosity Experiment across the Nicaragua Trench (SERPENT) on a global class research
Figure 1.3: Map of the electromagnetic survey. Our transect begins 80 km seaward from the onset of outer rise faulting and ends 25 km from the coastline. Squares show the location of ocean bottom EM receivers and the blue line shows the CSEM transmitter towpath. Filled circles show venting seafloor fluid seeps. The star is the hypocenter of the 1992 M$_{w}$ 7.6 tsunamigenic earthquake.

deployment of 8 receivers spaced 10 km apart on the Cocos plate abyssal plain, 36 receivers spaced 4 km apart on the Cocos plate outer rise and Caribbean plate forearc slope, and 6 receivers spaced 10 km apart on the Caribbean plate shelf (Figure 1.3). The receivers continuously recorded data for 19 days. The instrument and data recovery success rates were 96% and 91%, respectively.

The instruments recorded passive magnetotelluric data and controlled-source electromagnetic data to image the electrical resistivity structure of the oceanic plate and forearc margin. Our objectives were to: 1) quantify the degree of upper mantle serpentinization in the outer rise and the magnitude of water subducted; 2) map
porosity structure, fluid pathways, and the pattern of hydrothermal circulation to detect changes in the hydrogeological system and relate them to tectonic processes; and 3) identify the sources and migration pathways of fluids venting on the forearc slope seafloor.

1.5 Marine EM instrumentation

Electromagnetic sounding requires time-series measurements of the electric and magnetic fields. Both land and marine applications use the same family of sensors. Scripps designs and builds ocean-bottom instruments in-house. The current fleet of instruments are the 3rd iteration of the original design (Constable et al., 1998). For a detailed discussion regarding marine instrumentation, see Constable (2013).

The electric field is measured with a pair of electrodes separated by some nominal distance to increase signal-to-noise. Considering that electric fields have units of V m\(^{-1}\), electrodes are essentially voltmeters. To accurately measure the potential difference across the electrodes, the input impedance must be larger than that of the contact material. Since seawater is highly conductive, we are able to utilize lower impedance electrodes that limit unwanted Johnson thermal noise.

Measuring the potential difference requires a small amount of current to be drawn. Therefore, we require non-polarizing (reversible) electrodes to avoid contaminating the recordings with galvanic distortions (Corwin, 1973). Reversible electrodes fall under one of two main categories. Polarizing ‘electrodes of the first kind’ make use of bare metals, such as stainless steel. Such electrodes are non-polarizing when the metal is submerged in an ionic solution that is saturated in its own cations. Land instruments often use electrodes of this type, such as the popular copper-copper sulfate ‘porous pot’ electrodes. Marine instruments rely on ‘electrodes of the second kind’ that consist of a bare metal coated with a low solubility salt and submerged in an ionic solution of the same anion. Silver-silver chloride (Ag-AgCl) electrodes are ideal for marine applications since seawater is a sodium chloride solution. Scripps constructs a version of the Webb et al. (1985) Ag-AgCl electrodes.

The magnetic field is measured with either fluxgate or induction coil vector field
magnetometers. Long period instruments use fluxgate magnetometers that measure the total magnetic field. Fluxgates run current through a pair of coils that are each wound around a core of high magnetic permeability. The cores are saturated in opposite directions. The core that saturates sooner is better aligned with the external magnetic field, such that a time varying magnetic field corresponds to a time varying magnetization.

Induction coil magnetometers work by winding a metal wire as many times as possible around a high magnetic permeability core. Scripps uses cores made from mu-metal alloy with aluminum wire to reduce weight (Constable, 2013). The time-varying external magnetic field induces a current in the wire that is measured as a potential difference. Fluxgates can measure the magnetic field down to 10 s periods, and are better behaved than induction coils at periods greater than 500 s. The performance of an induction coil improves as a function of increasing frequency, but is relatively well behaved out to $10^4$ s periods.

The amplifiers, data logger, batteries, and other electronics are housed inside an aluminum alloy cylinder. The cylinder and sensors are attached to a high density polyethylene frame equipped with glass spheres for flotation. An electronic compass attached to the side of the frame records the geographical heading as well as the pitch and dip of the instrument for the first 24 hours after it lands on the seafloor. The instrument is also equipped with an acoustic transponder, which is used to navigate the absolute position of the receiver on the seafloor as well as to trigger the release mechanism that allows the instrument to rise to the sea surface for recovery. An assembled instrument is shown in Figure 1.4. The entire system can withstand pressures at 6 km seafloor depths.
Figure 1.4: SIO ocean bottom EM receiver.
1.6 References


Chapter 2

Marine Magnetotelluric Method

The geomagnetic field is continuously bombarded by charged particles, or plasma, that originate from the sun. This stream of plasma is known as the solar wind. The interaction of Earth’s magnetic field with the solar wind generates large-scale electrical currents in the magnetosphere and the ionosphere (Chapman, 1951; Campbell, 2003). To first order, the currents flow as uniform horizontal sheets. Solar activity, sometimes referred to as space weather, determines the force that the solar wind exerts on the geomagnetic field. Since space weather is highly chaotic, it modulates the amplitude of the geomagnetic field over a wide spectrum of frequencies. The geomagnetic spectrum, dubbed as the “Grand Spectrum”, is shown in Figure 2.1.

It is the external forcing of the geomagnetic field that the MT method seeks to utilize as its excitation source. Most of the energy at frequencies relevant to MT ($10^{-4} - 10^0$ Hz) are generated by the van Allen radiation belts of the magnetosphere at altitudes greater than 100 km (Campbell, 2003). Some of the plasma particles that strike the magnetopause leak into the magnetosphere and flow in an orbit centered around the equator. The motion of charged particles around the equator is effectively a ring of electrical current, hence the name ‘ring current’. The ring current generates a large-scale relatively uniform external magnetic field that opposes the internally generated geomagnetic field. Recall that the solar wind strength fluctuates chaotically according to solar weather patterns. As such, the strength of the external magnetic field also fluctuates.

An electrically insulating atmosphere separates the oscillating external magnetic
Figure 2.1: The amplitude spectrum of geomagnetic field variations. The source-fields in the MT range of the frequency spectrum are strengthened by solar storm activity. From Constable and Constable (2004).

field from the surface of the Earth. Thus, it is reasonable to assume that a uniform horizontally polarized plane-wave propagates through the resistive atmosphere and strikes the surface of the Earth. This oscillating plane-wave induces secondary electric fields in the Earth, which in turn produce secondary magnetic fields. The theoretical basis of the MT method hinges on this assumption, otherwise known as the plane-wave assumption. As Tikhonov and Cagniard discovered independently more than 60 years ago, it is possible to estimate the subsurface electrical resistivity structure from electric and magnetic field measurements, so long as the plane-wave assumption holds true.
2.1 Electromagnetic induction

Below, I present a brief review of the well developed magnetotelluric theory (e.g., Ward and Hohmann 1988). Maxwell’s equations, which describe the dynamics of electric and magnetic fields, constitute the theoretical foundation of the MT method. The differential formulation of the general Maxwell’s equations are:

\[ \nabla \times \mathbf{E} = -\frac{\partial \mathbf{B}}{\partial t} \]  
\[ \nabla \times \mathbf{H} = \mathbf{J} + \frac{\partial \mathbf{D}}{\partial t} \]  

where \( \mathbf{E} \) is the electric field \([\text{V m}^{-1}]\), \( \mathbf{B} \) is the magnetic flux density \([\text{T}]\), \( \mathbf{H} \) is the magnetic field \([\text{A m}^{-1}]\), \( \mathbf{J} \) is the electric current density \([\text{A m}^{-2}]\), and \( \mathbf{D} \) is the displacement current density \([\text{C m}^{-2}]\). Equation (2.1) is Faraday’s law and states that a time-varying magnetic flux density induces a loop of current. Equation (2.2) is Ampere’s law, which when viewed from the context of Faraday’s law states that an electric current induced by a time-varying magnetic field will in-turn generate its own ‘secondary’ magnetic field with amplitude proportional to the total current density. Originally, Ampere’s law lacked the inclusion of displacement currents until Maxwell discovered that time-varying electric fields also generate magnetic fields.

We must specify the constitutive relations to apply Maxwell’s equations to Earth materials. For the simple case of a homogeneous isotropic medium:

\[ \mathbf{B} = \mu \mathbf{H} \]  
\[ \mathbf{J} = \sigma \mathbf{E} \]  
\[ \mathbf{D} = \varepsilon \mathbf{E} \]  

where \( \mu \) is the magnetic permeability \([\text{N A}^{-2}]\), \( \sigma \) is the electric conductivity \([\text{S m}^{-1}]\),
and $\varepsilon$ is the electric permittivity [F m$^{-1}$]. Note that equation (2.4) is Ohm’s law. The constitutive relations reduce the general form of Maxwell’s equations to:

\[
\nabla \times \mathbf{E} = -\frac{\partial \mathbf{B}}{\partial t} \tag{2.6}
\]
\[
\nabla \times \mathbf{B} = \mu_0 \sigma \mathbf{E} + \varepsilon \frac{\partial \mathbf{E}}{\partial t} \tag{2.7}
\]

We can combine equations (2.6)-(2.7) by applying the handy \textit{curl-curl} vector identity:

\[
\nabla \times \nabla \times \mathbf{A} = \nabla (\nabla \cdot \mathbf{A}) - \nabla^2 \mathbf{A} \tag{2.8}
\]

Taking the \textit{curl} of equation (2.6):

\[
\nabla \times \nabla \times \mathbf{E} = -\frac{\partial}{\partial t} (\nabla \times \mathbf{B}) \tag{2.9}
\]

and applying the \textit{curl-curl} identity yields:

\[
\nabla \times \nabla \times \mathbf{E} = \nabla \cdot \nabla \cdot \mathbf{E} - \nabla^2 \mathbf{E} \tag{2.10}
\]

We make the assumption that the interior of the homogeneous medium is free of electric current sources. Mathematically, this translates to:

\[
\nabla \cdot \mathbf{E} = 0 \tag{2.11}
\]

Equation (2.11) reduces equation (2.10) and as a result also reduces equation (2.9) to:

\[
\nabla \times \nabla \times \mathbf{E} = -\nabla^2 \mathbf{E} = -\frac{\partial}{\partial t} (\nabla \times \mathbf{B}) \tag{2.12}
\]
Substituting equation (2.7) into equation (2.12) and applying the quasi-static approximation to neglect displacement currents yields:

$$\nabla^2 E = \mu \sigma \frac{\partial E}{\partial t}$$  \hspace{1cm} (2.13)

Similarly, we can apply the procedure outlined in equations (2.9)-(2.13) to the magnetic field in equation (2.7):

$$\nabla^2 B = \mu \sigma \frac{\partial B}{\partial t}$$ \hspace{1cm} (2.14)

Equations (2.13)-(2.14) show that the fields propagate diffusively and as such are governed by the diffusion equation. Therefore, electromagnetic induction in the solid Earth is analogous to heat flow. In an inhomogeneous medium, these equations are augmented with terms accounting for boundary charges.

The magnetic source-field is a periodic horizontally polarized plane-wave, which can be expressed mathematically in harmonic form:

$$B = B_0 e^{i\omega t}$$ \hspace{1cm} (2.15)

Substituting the harmonic form of $B$ into the right hand side of equation (2.14) yields:

$$\nabla^2 B = i\omega \mu \sigma B$$ \hspace{1cm} (2.16)

Similarly, we can substitute equation (2.15) into equation (2.12) to get the electric field:

$$\nabla^2 E = i\mu \omega \sigma E$$ \hspace{1cm} (2.17)

The product of the scalar quantities on the right side of the equation has the dimension m$^{-2}$. Hence, we can define a complex wavenumber as the square root of the scalar
product, \( k = \sqrt{i\omega \mu \sigma} \). Substituting the complex wavenumber gives:

\[
\nabla^2 B = k^2 B \tag{2.18}
\]

\[
\nabla^2 E = k^2 E \tag{2.19}
\]

For a vertically incident horizontally polarized plane-wave impinging on a homogeneous half-space, the horizontal fields will be constant. However, the fields will attenuate with depth, reducing equations (2.18)-(2.19) to:

\[
\frac{d^2 E}{dz^2} + k^2 E = 0 \tag{2.20}
\]

\[
\frac{d^2 B}{dz^2} + k^2 B = 0 \tag{2.21}
\]

which are homogeneous linear second-order differential equations with a well defined characteristic solution of the form:

\[
B(z) = C_1 e^{kz} + C_2 e^{-iz} \tag{2.22}
\]

where subscript \( z \) is the vertical direction in cartesian coordinates. Let us evaluate \( k \) as the sum of real and imaginary parts:

\[
k = (1 + i) \sqrt{\omega \mu \sigma / 2} \tag{2.23}
\]

and define the real part of \( k \), with dimension of meters, as:

\[
z_s(\omega) = \sqrt{\frac{2}{\omega \mu \sigma}} \tag{2.24}
\]
where \( \omega = 2\pi f \). Substituting equation (2.24) into equation (2.22) yields:

\[
B(z) = C_1 e^{z/z_s} e^{iz/z_s} + C_2 e^{-z/z_s} e^{-iz/z_s}
\]  

(2.25)

For the Earth system, we know the fields must decay with depth, which translates to \( C_1 = 0 \). Our boundary condition at the surface \( (z = 0) \) is the full amplitude of the sinusoidal magnetic field in equation (2.15), such that \( C_2 = B_0 e^{i\omega t} \). Therefore, the full solution for the magnetic field of a homogeneous half-space is:

\[
B(z, t) = B_0 e^{i\omega t} e^{-z(1+i)/z_s}
\]  

(2.26)

Having solved for the magnetic field \( B \), it is simplest to deduce the solution to the electric field \( E \) by substituting equation (2.26) into Ampere’s law, equation (2.7). Rearranging:

\[
E = \frac{\nabla \times B}{\mu \sigma}
\]  

(2.27)

Suppose we are substituting the solution for \( B_x \), then:

\[
E_y(z, t) = \frac{e^{i\omega t}}{\mu \sigma} \frac{dB_x}{dz} = -B_0 (1 + i) e^{i\omega t} e^{-z(1+i)/z_s} \left( \frac{\omega z_s^2}{2} \right)
\]  

(2.28)

where subscripts \( x \) and \( y \) are orthogonal horizontal axes in cartesian coordinates. As expected, the electric currents are horizontal but flow perpendicular to the magnetic field. The coupling between orthogonal electric and magnetic fields lends us the ability to see MT signals in real time-series data. Figure (2.2) shows an example of clean coherent MT signals in time-series data collected during the SERPENT survey.

Equations (2.26) and (2.28) describe how the fields decay. When the penetration depth of the fields \( z = z_s \), the amplitudes decay to \( 1/e \) (or \( \approx 37\% \)) and the phase shifts by one radian. This is known as the magnetotelluric skin-depth approximation. A more
Figure 2.2: Two hours of time-series data from a SERPENT receiver. The data were low pass filtered at 0.01 Hz to highlight the strong MT signals, which are coherent in orthogonal electric and magnetic channels.

A practical form of the skin-depth is:

\[ z_s(f) \approx 500 \left( \frac{1}{\sigma f} \right)^{1/2} \]  \hspace{1cm} (2.29)

where \( f \) is the frequency with the dimension \( s^{-1} \). The skin-depth offers insight into the depth that a particular frequency of the fields would be sensitive to, depending on the electrical conductivity of the excited material. Earth materials exhibit electrical conductivities that span more than six orders of magnitude. As an example, the skin-depth of seawater (\( \sigma \approx 3 \text{ S/m} \)) at 100 s period is almost 3 km while that of cold crystalline rock (\( \sigma \approx 10^{-6} \text{ S/m} \)) is 5,000 km! The source-field will induce large currents in regions with large electrical conductivities, whereby the magnetic fields attenuate rapidly with depth. Conversely, less current is induced in resistive regions, so the fields will attenuate less and penetrate greater depths. As such, lower frequencies penetrate greater depths than higher frequencies.
2.2 The MT impedance tensor

The pioneering works of Tikhonov (1950) and Cagniard (1953) determined that by measuring magnetic and electric field time-series in orthogonal directions, it is possible to estimate electrical conductivity as a function of depth. The frequency-dependent scalar transfer function is the ratio of an orthogonal pair of electric and magnetic fields:

$$Z_{xy}(\omega) = \frac{E_x}{H_y}$$

where $\mu H = B$. With $Z_{xy}$, we calculate the half-space resistivity – the reciprocal of conductivity – and the phase:

$$\rho_a(\omega) = \frac{1}{\sigma} = \frac{1}{\mu \omega} \frac{1}{|k|^2} = \frac{1}{\mu \omega} |Z(\omega)|^2$$

$$\phi = \text{arg} \frac{\text{Im}(Z)}{\text{Re}(Z)}$$

where $\rho_a$ is the bulk volume average of the resistivity, otherwise known as the apparent resistivity, and $\phi$ is the phase. We also see that:

$$Z_{xy} = \frac{E_x}{H_y} = \frac{\mu \omega}{k}$$

Since the complex wavenumber $k = \sqrt{i \omega \mu \sigma}$, then the phase of the magnetic field lags the orthogonal electric field by 45° in a homogenous half-space.

Cagniard (1953) derived equations (2.31) and (2.32) in detail and proceeded to solve for one-dimension horizontally stratified systems containing either two or three layers of variable conductivities. Cagniard also had the foresight to envision marine magnetotelluric soundings. He went so far as to suggest that it is possible to reliably measure magnetic fields in offshore environments if one could “install a self-recording magnetometer on a series of piles forming a foundation or in an immersed box on the
bottom of the sea”.

In a homogeneous half-space, measuring the transfer function of a pair of orthogonal electric and magnetic fields at a single frequency would suffice to estimate the apparent resistivity. A layered half-space would again require a single transfer function, but measured at several frequencies to estimate the apparent resistivity as a function of depth. This is because in a one-dimensional domain where resistivity only varies with depth, the transfer function of one orthogonal pair, say $B_x$ and $E_y$, is equal and opposite in sign to the transfer function of the other, $B_y$ and $E_x$.

When the resistivity varies in two or three spatial dimensions, we must measure both pairs of orthogonal electric and magnetic fields. The impedance tensor considers the transfer function between all four components (Swift, 1967):

$$
Z = \begin{bmatrix}
Z_{xx} & Z_{xy} \\
Z_{yx} & Z_{yy}
\end{bmatrix}
$$

where:

$$
E = ZH
$$

2.3 MT data processing

The goal of MT data processing is to estimate the complex impedance tensor across several decades of frequency, which is an exercise in digital signal processing. Scripps marine MT instruments record two channels of electric field time-series ($E_x$ & $E_y$) and two channels of magnetic field time-series ($B_x$ & $B_y$).

To estimate the four transfer functions that comprise the impedance tensor shown in equations (2.34) and (2.35), the time-series data are first decomposed into the frequency domain with the Fourier transform and then processed with a robust multivariate statistical scheme. A generalized processing procedure is illustrated in the flow diagram in Figure 2.3.
Figure 2.3: MT processing flowchart.
2.3.1 Preparing the time-series for processing

Each channel of time-series data is cascade decimated and then Fourier transformed to yield an ensemble of Fourier coefficients (FCs), typically in the $10^{-5}$ to 1 Hz frequency band. Cascade decimation minimizes computational expense while maximizing the number of independent FCs that are extracted from the time-series. To maximize the number of FCs at a given frequency, we segment the time-series into multiple overlapping sections that span a specified length of time. The window length is shorter when calculating higher frequency FCs and longer when calculating lower frequency FCs. We segment the time-series repeatedly, once per unique window length, to yield a series of “cascading” windows. Without additional filtering, different length windows that overlap in time will give redundant and correlated high-frequency FCs. To avoid redundancy, we low-pass filter and then decimate the windows. Longer time windows are subjected to lower cut-off frequencies when low-pass filtered, which minimizes computational expense.

Following cascade decimation, we decompose each window into the frequency domain with a discrete Fourier transforms (DFT), such as the Fastest Fourier Transform in the West (FFTW). Since the time-series data are recorded by electronic sensors whose response functions are unique to each sensor, we must apply a frequency-dependent calibration that is specific to each sensor. After the FCs are calculated and the calibration applied for a particular channel of time-series, we can plot the amplitude of the FCs per frequency as a function of the recording time to produce a time-dependent spectrogram. Figure 2.4 shows two high-quality FCs – collected during the SERPENT survey – from an orthogonal pair of electric and magnetic channels.

Besides the external magnetic fluctuations that comprise the MT source-field, few high-amplitude phenomena operate at the intermediate range of frequencies that are of interest to MT practitioners. Therefore, barring any problems with instrumentation or other anomalous factors, we may be able to discern moderate fluctuations in the strength of the MT source through visual inspection of the FCs. In the event that a large solar storm suddenly strikes the geomagnetic field, the dramatic onset of high-power FCs can be spectacular and lend significant improvements to the data quality. Although brief, we were fortunate enough to observe such an event during the SERPENT survey.
Figure 2.4: An example of high-quality Fourier coefficients for an orthogonal pair of electric and magnetic channels from the SERPENT survey. The band of high amplitudes on April 23rd is caused by the passage of the controlled-source transmitter over the receiver, which is treated as noise and excluded from MT processing. The more gradual onset of high amplitudes on May 2nd is caused by moderate solar storm activity, beginning May 2nd, the tail end of the data record. Upon closer inspection of Figure 2.4, we clearly see that the amplitudes begin to ramp up on May 2nd. The solar storm increased the source-power by at least two orders of magnitude!

Plotting the FCs can help to delineate a variety of other interesting phenomena, such as large-amplitude environmental ‘noise’ associated with heavy seas, tidal oscillations, and ocean-bottom water currents. Figure 2.5 illustrates how ocean waves are clearly manifested in the magnetic FCs at a site deployed in shallow water on the continental shelf.

Attempting to assess the data quality by visualizing the FCs, while informative,
Figure 2.5: The magnetic FCs for a shallow water site from SERPENT. The instrument, deployed at an ocean depth of 100 m, records higher field amplitudes at 10-50 s periods due to the propagation of ocean waves.

can also be misleading. A site that appears to be low-quality based on the FCs can produce high-quality impedance tensors after processing. In practical terms, we visually inspect the FCs to document first impressions of the overall data quality. Typically, a channel contains anomalous FCs that span the start or end of the time-series while the rest of the FCs are well-behaved, which is due to the instrument logging data while it is in the water column and not on the seafloor. We simply flag the offending time segments and discard their FCs. Once we have filtered outliers for all of the data, we are ready for processing.

Occasionally, a sensor will fail prior to recording any useable data and so all of the FCs are deemed corrupt for that channel. The MT method requires that we estimate all four transfer functions of the impedance tensor for the data to be viable for modeling. Therefore, when one sensor fails then all data from the corresponding receiver are unfit for modeling. I compare the FCs of a functioning and a defective electric field sensor installed on the same instrument in Figure 2.6. Although extracting any meaningful MT data from the affected site is likely hopeless, the computational expense of processing the data has become so cheap that it is always worth trying.

Of the 54 instruments deployed during SERPENT, sites s29 and s40 were lost at sea, site s27 had a defective magnetic field sensor, and sites s02 and s30 had defective electric field sensors. Otherwise, the majority of the instruments showed good to
Figure 2.6: Fourier coefficients from the electric channels of site s02, showing a defective $E_x$ sensor. The high amplitudes in the $E_y$ sensor are due to energy emitted by the controlled-source transmitter.

excellent quality FCs.

2.3.2 Univariate least squares algorithms

As previously discussed, we use the FCs to estimate the complex impedance tensor $Z$. Since the FCs are estimated as a function of frequency, we can estimate $Z$ for each frequency independently. There are a variety of methodologies to do so. The classical approach is limited to estimating the impedance at each measurement site with no outside constraints.

Expanding equation (2.35):
we have four unknowns but only two equations. It is useful to think of the electric field as the noisy output signal of two uncorrelated magnetic field input signals. In other words, the MT method is a two-input one-output system. We multiply equations (2.36)-(2.37) by the complex conjugate of any pair of field components and then take the mean of the measured data over a band of frequencies (Vozoff, 1972), which yields a total of 24 possible combinations with the form:

\[ E_x A^* = Z_{xx} H_x A^* + Z_{xy} H_y A^* \]  
\[ E_y B^* = Z_{yx} H_x B^* + Z_{yy} H_y B^* \]

where \( A^* \) and \( B^* \) are complex conjugates of different field components. We can solve for each of the four transfer functions six different ways with the set of 24 equations. Two of the six possible formulations for \( Z_{xy} \) are:

\[ Z_{xy} = \frac{(E_x^2)(H_x H_y^*) - (E_x E_y^*)(H_x E_x^*)}{(H_y E_y^*)(H_x E_y^*) - (H_y E_y^*)(H_x E_x^*)} \]  
\[ Z_{xy} = \frac{(E_x H_x^*)(H_x H_y^*) - (E_x H_y^*)(H_x^2)}{(H_y H_x^*)(H_x H_y^*) - (H_y^2)(H_x^2)} \]

This is essentially a least-squares regression that seeks to minimize the squared data residuals. While both formulations should yield equivalent estimates for \( Z_{xy} \), they are heavily biased in the presence of noise. Equation (2.40) is dominated by the power of \( E_x \) in the numerator so any noise in that channel would bias \( Z_{xy} \) upwards, and equation
(2.41) is dominated by the power of $H_y$ in the denominator so any noise in that channel would bias $Z_{xy}$ downwards (Vozoff, 1972).

Historically, MT practitioners preferred the magnetic complex conjugates to estimate the transfer functions since the noise was found to be better behaved (lower cross-power relative to electrics). Written in more familiar form, the least-squares regression seeks to minimize:

$$\min \| r \|^2 = \sum \| (E - H^\dagger \hat{Z}) \|^2$$  \hspace{1cm} (2.42)

where $(\dagger)$ is the complex conjugate transpose, $(\hat{\cdot})$ is an estimated variable, and $\| r \|$ is the euclidean norm of the residual vector. Rearranging to solve for the impedance:

$$\hat{Z} = (H^\dagger H)^{-1} H^\dagger E$$  \hspace{1cm} (2.43)

which is the well-known ordinary least squares estimator that takes the ratio of the cross-power and autopower spectral estimates. Although the ordinary least squares estimator is theoretically unbiased, implicit in equation (2.43) is the assumption of noise-free magnetic data. Considering equation (2.41), we again see that the classical approach produces systematically downward biased impedance tensor estimations.

Following the starting assumptions of the MT method, the electric and magnetic fields behave as zero-mean Gaussian distributed variables that are time-independent. Simply put, the MT method is a random stationary process. A more statistically elegant way to express the transfer functions is to utilize the covariance matrix. For a one-dimensional Earth composed of horizontally stacked layers, the diagonal components of the tensor are zeros, reducing equations (2.36)-(2.37) to:

$$E_x = Z_{xy} H_y$$ \hspace{1cm} (2.44)

$$E_y = Z_{yx} H_x$$ \hspace{1cm} (2.45)
so that the MT method is simplified to a single-input single-output system. With the addition of noise in the electric channels, equations (2.44) and (2.45) become:

\[ E_i = Z_{ij}H_j + n_i, \quad i \neq j \] (2.46)

where \( n \) is noise. In this way, we can treat the covariance of equation (2.46) as equivalent to that of a time-delay problem:

\[
C_{ij} = \begin{bmatrix}
\text{cov}(E_i, E_i) & \text{cov}(E_i, H_j) \\
\text{cov}(H_j, E_i) & \text{cov}(H_j, H_j)
\end{bmatrix} = \begin{bmatrix}
|Z_{ij}|^2 \sigma_{H_j}^2 + \sigma_{n_i}^2 & Z_{ij} \sigma_{H_j}^2 \\
Z_{ij}^* \sigma_{H_j}^2 & \sigma_{H_j}^2
\end{bmatrix} \tag{2.47}
\]

where \( \sigma^2 \) is the variance. Since estimating the covariance of the measured data is trivial with modern computers, we use equation (2.47) to estimate the 1D impedance:

\[
\hat{Z}_{ij} = \frac{\hat{C}(E_i, H_j)}{\hat{C}(H_j, H_j)} \tag{2.48}
\]

Although it is relatively safe to assume that the electric and magnetic fields behave as zero-mean Gaussian distributed variables, field measurements are certainly distorted by noise in all channels such that even estimates of 1D impedances are downward biased by noise in magnetic data.

The first breakthrough in MT processing was developed in the late 1970’s. The remote-reference method circumvents the problem of noise-power bias in impedance estimations by utilizing additional channels of magnetic data (Goubau et al., 1978; Gamble et al., 1979a). Ideally, the additional magnetic channels are recorded at a quiet site that is far enough from the primary measurement site that the noise is uncorrelated yet close enough that the source-field between the two sites is equivalent. Applying a remote channel \( H_R \) in the least-squares regression, then:

\[
\hat{Z} = (H_R^\dagger H)^{-1}H_R^\dagger E \tag{2.49}
\]
Assuming the noise is uncorrelated between $H$ and $H_R$, the cross-power will be unbiased and thus yield unbiased estimates of $\hat{Z}$ (Gamble et al., 1979b). However, the remote-reference method will suffer when the measured data begin to violate our implicit assumption of Gaussian-behaved variables, including for generally Gaussian distributed data that contain the occasional significant outlier. Luckily, a number of solutions exist that accommodate outliers, typically referred to as robust estimators.

The simplest way to identify the breakdown of a non-robust regression analysis is to test for Gaussian-distributed residuals:

$$rr^\dagger = \sigma_n^2 I \quad (2.50)$$

where $I$ is the identity matrix and $\sigma_n^2$ is the noise variance. When this equality does not hold, then the ordinary least squares estimate is no longer valid. Least squares seeks to minimize the sum of the squared residuals, or the $L_2$ norm, which magnifies the effect of outliers on the model misfit. One solution is to minimize the absolute residuals, or the $L_1$ norm. Initially introduced by Egbert and Booker (1986), an effective implementation of the $L_1$ norm is to apply a hybrid minimization to an adaptively weighted least squares regression using what is known as an M-estimator:

$$\hat{Z} = (H_R^t wH)^{-1} H_R^t w E \quad (2.51)$$

$$w_i(r_i) = \begin{cases} 
1, & |r_i| < r_o \\
\frac{r_o}{|r_i|}, & |r_i| \geq r_o 
\end{cases} \quad (2.52)$$

where $w$ is the weighting vector, $r_i$ is the residual, and $r_o$ is a threshold for down-weighting the residual that is set by the user. In practice, we initially determine the residual vector $r$ with an ordinary least squares regression to calculate the weighting vector. Then, we calculate the weighted least squares, update the weighting vector, and iterate until reaching convergence, a procedure generally referred to as iteratively re-weighted least squares.
As is often the case, MT measurements are distorted by correlated noise, such as when an instrument is subjected to motional forces that cause it to wobble. The robust remote-reference method will fold these distortions into the impedance tensor estimate. To address such shortcomings, we need to adopt a multivariate statistical framework.

### 2.3.3 The robust multiple-station algorithm

The method of choice that Scripps uses to process marine MT data and that I use for the SERPENT data is the robust multiple-station algorithm, or multi-station for short (Egbert, 1997, 2002). The multi-station method utilizes the multivariate errors-in-variables (MEV) model to explicitly approximate the dimension of the external magnetic source input. Recall that the univariate remote-reference method implicitly assumes a two-input source, which is the plane-wave MT source. With the MEV model, we avoid such a restriction and hence are able to estimate both planar and non-planar sources, of which the latter are treated as coherent noise.

Following the derivation of Egbert and Booker (1989), we define the MT system as:

\[
x_j = \begin{bmatrix}
H_{xj} \\
E_{xj} \\
H_{yj} \\
E_{yj}
\end{bmatrix}
\]  

where \( j \) denotes the site number. To estimate the two planar MT sources, we define each channel as a linear superposition of the two independent polarizations:

\[
\mathbf{H}_{xj} = \begin{bmatrix}
h_{xj,1} \\
h_{xj,2}
\end{bmatrix}
\begin{bmatrix}
\alpha_1 \\
\alpha_2
\end{bmatrix}
\]  

where \( \alpha \) is a vector of length \( p = 2 \) that contains all possible polarizations that together form the linear superposition of the source. Substituting equation (2.54) into equation (2.53) and adding incoherent noise:
\[ \begin{bmatrix} h_{xj,1} \\ e_{xj,1} \\ h_{yj,1} \\ e_{yj,1} \end{bmatrix} \alpha_1 + \begin{bmatrix} h_{xj,2} \\ e_{xj,2} \\ h_{yj,2} \\ e_{yj,2} \end{bmatrix} \alpha_2 + \begin{bmatrix} \varepsilon_{Hx,j} \\ \varepsilon_{Ey,j} \\ \varepsilon_{Ex,j} \\ \varepsilon_{Ey,j} \end{bmatrix} = u_j \alpha + \varepsilon_j \quad (2.55) \]

We can calculate the impedance tensor at site \( j \) from the \( 4 \times 2 \) matrix \( u_j \):

\[ Z_j = \begin{bmatrix} u_{21} & u_{22} \\ u_{41} & u_{42} \end{bmatrix}^{-1} \quad (2.56) \]

Since we cannot measure the external source \( \alpha \) independently of \( u_j \), we look to solve for \( u_j \). However, recall that we have more unknowns than knowns at a single site when assuming noise in both input and output channels. Instead of limiting ourselves to a remote-reference site, we consider all channels from a group of sites simultaneously:

\[ X = \begin{bmatrix} x_1 \\ x_2 \\ \vdots \\ x_J \end{bmatrix} = \begin{bmatrix} u_1 \\ u_2 \\ \vdots \\ u_J \end{bmatrix} \alpha + \begin{bmatrix} \varepsilon_1 \\ \varepsilon_2 \\ \vdots \\ \varepsilon_J \end{bmatrix} = U\alpha + \varepsilon \quad (2.57) \]

where \( J \) is the total number of sites. Setting the total number of channels as \( K = 4J \), then \( U \) is a \( K \times 2 \) matrix. We find the solution to \( U \) with a scaled eigendecomposition of the covariance. In the Fourier domain, the covariance is known as the spectral density matrix (SDM):

\[ S = XX^\dagger = U\alpha\alpha^\dagger U^\dagger + \varepsilon\varepsilon^\dagger = U\Sigma_{\alpha}U^\dagger + \Sigma_{\varepsilon} \quad (2.58) \]

where \( \Sigma_{\alpha} \) and \( \Sigma_{\varepsilon} \) are the covariance of the signal and the noise, respectively, and \( S \) is the SDM (\( K \times K \) matrix). We assume to know the noise covariance \( a \ priori \), which we define as a diagonal matrix of variances that represent the uncorrelated noise in each channel. We use the standard deviation of the estimated noise to scale the SDM:
\[ S' = \Sigma_{e}^{-\frac{1}{2}}S\Sigma_{e}^{-\frac{1}{2}} \quad (2.59) \]

Applying the decomposition to the scaled SDM:

\[ S' = Q\Lambda Q^\dagger \quad (2.60) \]

where \( \Lambda \) is the diagonal matrix of eigenvalues and \( Q \) is a unitary matrix with eigenvector columns. The eigenvectors correspond to linear combinations of the two non-zero eigenvalues that define \( U \). We partition \( Q \) into a primary and secondary matrix:

\[ Q = \begin{bmatrix} Q_1 \\ Q_2 \end{bmatrix} \quad (2.61) \]

where \( Q_1 \) is a \( p \times p \) matrix and \( Q_2 \) is a \((K - p) \times p\) matrix. The eigenvectors are equal to the product of scaled \( U \) and an arbitrary non-singular \( p \times p \) matrix \( A \). With minimal rearranging, we find that:

\[ \Sigma_{e}Q_2Q_1^{-1} = (U_2A)(U_1A)^{-1} = U_2U_1^{-1} = \tilde{Z} \quad (2.62) \]

The system that we have solved in equation (2.55) is specific to the polarized \( p = 2 \) dimensional plane-wave source. Realistically, we do not know the true dimension of the source since we cannot confidently rule out non-planar sources.

Subsequently, Egbert (1997) expanded the starting assumption in equation (2.55) to define the system in a more general form, one with an undefined source where we seek to estimate the magnitude of the term \( p \):

\[ x_j = \begin{bmatrix} U & V \end{bmatrix} \begin{bmatrix} \alpha \\ \gamma \end{bmatrix} + \varepsilon_j = W\beta + \varepsilon \quad (2.63) \]

where \( W \) is the combination of the primary source-field matrix \( U \) and the non-planar
source matrix $V$. In practice, equation (2.63) is solved iteratively. First, we robustly estimate the scaled SDM $S'$ and make an initial estimate of $W$ for the dominant eigenvectors. Then, we use $W$ to estimate $\beta$, after which we calculate the cleaned data (raw data minus estimated noise) and the residuals. We now use the cleaned data to recalculate the scaled SDM while using the residuals to robustly weight the data, iterating until convergence.

### 2.3.4 Processing results

The main advantage of the multi-station algorithm is its ability to capitalize on several channels of data from multiple receivers to improve the quality of impedance tensor estimates. Theoretically, the more channels of data we include, the better our impedance estimates will be.

There are some practical limitations to consider when deciding which receivers to process as a group. Because we can only process the subset of FCs that overlap in all channels for a given group, we should aim to maximize the data overlap when we allocate which sites to group together. In design, data overlap is rarely an issue because the most efficient way to perform a marine MT survey is to deploy all instruments consecutively. However, the lack of data quality equality across instruments means that after sifting through the FCs and discarding outliers, the extent of overlap varies on a site by site basis. If a site has had a significant fraction of its FCs discarded, its inclusion will severely limit data overlap across receivers. Therefore, a site with a limited set of useable FCs or that otherwise contains little data overlap should be processed with a unique group in order to best estimate the impedance for that one site alone.

With data overlap in mind, I created several groups that contained randomized combinations of three to five receivers with each receiver in a minimum of two groupings. After processing, I inspected the eigenvalues of each group. The eigenvalue “spectrum” roughly corresponds to the signal-to-noise ratio (SNR) of the processed data. Ideally, we expect to find two dominant eigenvalues that represent the primary plane-wave polarizations and have significantly higher power than the remaining eigenvalues. Figure 2.7 shows the eigenvalue spectrum of a group that exemplifies the good quality SNR generally representative of the SERPENT dataset.
Figure 2.7: Eigenvalues as a function of frequency for a group of sites. Note the increase in power for the third largest eigenvalue at around 1000 s, which is more pronounced in this group compared with others. This warrants caution when attempting to model data at periods longer than a few thousand seconds.

Since we use the two dominant eigenvalues to estimate the MT impedance tensor, it is worthwhile to confirm the validity of our plane-wave assumption by inspecting the polarization ellipse of the normalized eigenvectors. Figure 2.8 shows the polarizations of the three largest eigenvalues for the same group of sites in Figure 2.7. The short period eigenvectors (at 4 s) reflect the poor SNR of the impedance tensor estimates, the intermediate period eigenvectors (at 512 s) confirm that the two primary eigenvalues are those of the plane-wave source, and the long period eigenvectors (at 2980 s), while consistent from site to site, are not ideal plane-wave polarizations and should be treated with caution.

The eigenvalue spectrums and eigenvector polarizations provide enough information to use them as a basis for data filtering. I discard data at periods shorter than 20 s from further analysis due to their poor SNR, and at periods longer than 3000 s due to the increasing power and non-ideal polarization ellipse seen in the third eigenvalue.
Figure 2.8: The eigenvector polarization ellipses at 4 s, 512 s, and 2980 s periods for each site in a group of sites. At the shortest period, the dominant signal is likely coherent noise from ocean currents and tidal motions. At 2980 s, we see signs indicative of 3D data, which require 3D numerical solvers to model.
Figure 2.9: The electric and magnetic FCs from trench site s32 are strongly contaminated by tidal noise, yielding relatively low quality impedance estimates.

A handful of receivers were only processed in poorly performing groups. To ensure that all responses were of the highest caliber possible, I reshuffled such receivers into groups that displayed the best SNR. A few receivers deployed in the vicinity of the trench-axis consistently produced low quality responses relative to other receivers. A quick look at the FCs from trench site s32 in Figure 2.9 shows that it is riddled with high-power noise. The noise appears to be coherent in electric and magnetic channels and oscillates over a diurnal cycle. This is a clear indication that the source of the noise is tidal-driven ocean-bottom currents. Even with the significant level of noise in trench sites, the multi-station algorithm appears to effectively differentiate tidal driven signals from the plane-wave source when grouped with sites that did not suffer from tidal noise.

While the eigenvalues and eigenvectors provide us with the means to identify
Figure 2.10: Processed data from sites s03 and s32 shown as apparent resistivity and phase for the $Z_{xy}$ and $Z_{yx}$ components.
problematic impedance tensor estimates, the ultimate indicator of data quality is distinguished by how well-behaved the impedance estimates are as a function of frequency. We use equations (2.31)-(2.32) to calculate the more meaningful quantities of apparent resistivity and phase. The converted off-diagonal transfer functions of two SERPENT sites are shown in Figure 2.10.

2.4 Dimensional analysis

In order to apply the appropriate numerical tools to model MT data, we must decipher the dimensionality and geoelectric strike of the data. Using one-dimensional techniques to model the apparent resistivity of 2D or 3D data will produce spurious artifacts in the resistivity sounding, invalidating any interpretations of the result. Thus, a priori constraints on the data dimensionality are critical.

The impedance tensor can offer significant insight into the dimensionality of the data. For 1D electrical Earth structure, the diagonal components of the tensor are zeros and the non-diagonals are equal and opposite in sign:

\[
\begin{bmatrix}
0 & Z_{xy} \\
-Z_{xy} & 0
\end{bmatrix}
\]

(2.64)

In practice, our field measurements are always noisy and the real Earth is not perfectly 1D so that the diagonal and off-diagonal transfer functions of even simple geology almost never amount to be zeros or equal magnitude, respectively.

In a 2D Earth, where the resistivity varies with depth and one horizontal direction, the angle between the strike of the underlying resistivity structure and the strike of the field measurements must be considered, which I refer to as the relative strike angle (\(\zeta\)). When the relative strike is neither parallel (0°) nor perpendicular (90°), then the diagonal components of a noise-free 2D impedance tensor are equal and opposite sign and the off-diagonal components are different magnitudes. In the case where the relative strike is parallel or perpendicular, the diagonal components become zeros:
In 2D, the MT-source is comprised of two independent modes. Assuming the y-axis is aligned in the direction of conductivity gradients, then the $Z_{yx}$ off-diagonal component is the Transverse-Magnetic (TM) mode, which is driven by the magnetic field polarization that is perpendicular to the vertical plane (y- and z-axis). The TM mode is dependent on $H_x$, $E_y$, and $E_z$. The $Z_{xy}$ off-diagonal component is the Transverse-Electric (TE) mode, which is driven by the magnetic field polarization that is parallel to the vertical plane and is dependent on $H_y$, $H_z$, and $E_x$.

Lastly, in a 3D Earth, all components can take on any magnitude regardless of the relative strike angle:

$$Z = \begin{bmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & -Z_{xx} \end{bmatrix} \{ \zeta \neq 0^\circ, 90^\circ \}$$  \hspace{1cm} (2.65)

$$Z = \begin{bmatrix} 0 & Z_{xy} \\ Z_{yx} & 0 \end{bmatrix} \{ \zeta = 0^\circ, 90^\circ \}$$

Equations (2.64)-(2.66) show that the value each component of the impedance tensor takes on depends on dimensionality and the strike of the measured electric and magnetic fields relative to the geoelectric strike.

Since marine EM instruments are deployed from the deck of a ship and sink until landing on the seafloor, the receiver orientations are random both from site to site and with respect to the regional geoelectric strike. Therefore, we must rotate the tensors to the survey strike in order to compare the data between sites. We use the rotation matrix $R$:

$$R = \begin{bmatrix} \cos \theta & \sin \theta \\ -\sin \theta & \cos \theta \end{bmatrix}$$  \hspace{1cm} (2.67)

to rotate the reference frame of $Z$ by an arbitrary angle $\theta$. 

$$Z = \begin{bmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & -Z_{xx} \end{bmatrix} \{ \zeta \neq 0^\circ, 90^\circ \}$$  \hspace{1cm} (2.65)

$$Z = \begin{bmatrix} 0 & Z_{xy} \\ Z_{yx} & 0 \end{bmatrix} \{ \zeta = 0^\circ, 90^\circ \}$$

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$$Z = \begin{bmatrix} 0 & Z_{xy} \\ Z_{yx} & 0 \end{bmatrix} \{ \zeta = 0^\circ, 90^\circ \}$$

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$$R = \begin{bmatrix} \cos \theta & \sin \theta \\ -\sin \theta & \cos \theta \end{bmatrix}$$  \hspace{1cm} (2.67)

to rotate the reference frame of $Z$ by an arbitrary angle $\theta$: 

Figure 2.11: Matrix plot of the entire SERPENT dataset showing apparent resistivity as a function of frequency for the TM ($Z_{yx}$) and TE ($Z_{xy}$) modes.
Figure 2.12: Matrix plot of the entire SERPENT dataset showing phase as a function of frequency for the TM ($Z_{yx}$) and TE ($Z_{xy}$) modes.
\[ Z(\theta) = R(\theta) \, Z \, R(\theta)^{-1} \]  

(2.68)

Written explicitly, the rotated transfer functions are:

\[
Z_{xx}(\theta) = Z_2 + Z_3 \sin 2\theta + Z_4 \cos 2\theta \\
Z_{yy}(\theta) = Z_2 - Z_3 \sin 2\theta - Z_4 \cos 2\theta \\
Z_{xy}(\theta) = Z_1 + Z_3 \cos 2\theta - Z_4 \sin 2\theta \\
Z_{yx}(\theta) = -Z_1 + Z_3 \cos 2\theta - Z_4 \sin 2\theta
\]

(2.69)

where:

\[
Z_1 = \frac{Z_{xy} - Z_{yx}}{2}, \quad Z_2 = \frac{Z_{xx} + Z_{yy}}{2} \\
Z_3 = \frac{Z_{xy} + Z_{yx}}{2}, \quad Z_4 = \frac{Z_{xx} - Z_{yy}}{2}
\]

(2.70)

Figures 2.11 and 2.12 show the apparent resistivity and phase of the SERPENT dataset after rotating the impedance tensors to the survey strike. For typical oceanic plate, we expect the conductivity structure to vary as a function of plate age. The SERPENT survey strike was aligned perpendicular to the trench axis at a rotation of 51° clockwise from north. Conveniently, the strike of the trench also happens to be parallel to the regional strike of constant plate age. Therefore, having rotated the data to the survey strike, the \( Z_{yx} \) and \( Z_{xy} \) transfer functions form the TM and TE modes, respectively.

### 2.4.1 Amplitude-based dimensionality

We cannot determine the data dimensionality by inspecting unrotated impedance tensor estimates, at least not without \textit{a priori} knowledge of the geoelectric strike. The diagonal transfer functions of noise-free impedance tensor estimates are equal to zero for 1D and strike-aligned 2D data, as described in equations (2.64) and (2.65). Upon further inspection of equation (2.69), it is clear that the trace of the impedance tensor matrix (the sum of \( Z_{xx} \) and \( Z_{yy} \)) is rotationally invariant. Thus, we can detect 3D
data when the trace is non-zero. We can also distinguish the geoelectric strike by finding the angle that maximizes and minimizes the magnitude of the off-diagonal and diagonal transfer functions, respectively. Although this gives us an estimate of the data’s geoelectric strike, it does not effectively tell us the dimensionality because we rarely observe zero value diagonal transfer functions in noisy data. In fact, both approaches are not practical since all measured data suffer from noise.

A classic estimator that quantifies the dimensionality of measured data is the **Swift skew**:

\[
S = \left| \frac{Z_{xx} + Z_{yy}}{Z_{xy} - Z_{yx}} \right|
\]

which is equal to the quotient of the trace and anti-trace of the impedance tensor, both of which are rotationally invariants *Swift* (1967). The trace is normalized by the anti-trace in order to more meaningfully quantify the magnitude of the diagonal components. In practice, a skew of 0.2 has been defined as an appropriate threshold to distinguish between noisy 1D/2D data and outright 3D data.

Figure 2.13 shows a matrix plot of Swift skews for the SERPENT dataset. Most of the data fall below the subjectively defined 3D threshold, but sites on the trench and the outermost forearc slope exhibit extremely large skews, with some even exceeding a normalized value of 1! This is the only dataset to my knowledge that has Swift skew estimates greater than unity, and implies that the data are sensing highly 3D conductivity structure. Figure 2.14 shows the eigenvector polarization ellipses for trench site s32, which demonstrates that the polarization of the largest eigenvalue is consistent with 3D data.

Based on the anomalous Swift skews, I exclude all of the data at sites s31, s32, s33, and s34 from 2D inversion. It is important to note that a Swift skew of less than 0.2 does not guarantee 1D/2D data because symmetrical 3D data will yield zero trace. In this regard, the impedance polar diagrams are a more robust tool for detecting 3D data.

Impedance polar diagrams provide a visualization of dimensionality by taking advantage of the symmetrical properties inherent to tensors. Following *Berdichevsky and Dmitriev* (2008), we apply equation (2.69) to rotate the impedance tensor incrementally
Figure 2.13: Swift skews of the entire SERPENT dataset. Notice that data with skews greater than 0.4-0.5 are isolated to sites s31-s34. Otherwise, most of the data are not conclusively 3D.
**Figure 2.14:** Eigenvector polarizations at 356 s period for trench site s32, where the Swift skews are greater than unity. Continental slope sites s36 and s37 have low Swift skews and are shown for comparison. Notice that for s32, the largest eigenvalue polarization (eigenvector #1) has magnetic and electric ellipses that are close to parallel, which is consistent with 3D distortions.

over one full cycle (from 0 to $2\pi$), plotting the amplitudes as a function of rotation angle as we go along:

\[
|Z_{xx}(\theta)| = |Z_{yy}(\theta + \pi/2)| = |Z_2 + Z_3\sin2\theta + Z_4\cos2\theta| \\
|Z_{xy}(\theta)| = |Z_{yx}(\theta + \pi/2)| = |Z_1 + Z_3\cos2\theta - Z_4\sin2\theta| \\
|\arg Z_{xy}(\theta)| = |\arg Z_{yx}(\theta + \pi/2)| = \left| \frac{\text{Im}(Z_1 + Z_3\cos2\theta - Z_4\sin2\theta)}{\text{Re}(Z_1 + Z_3\cos2\theta - Z_4\sin2\theta)} \right| 
\] (2.72)

Figure 2.15 illustrates how different data dimensionalities are rendered in polar diagram form. The amplitude of an off-diagonal component is constant and plots as a circle for 1D, peaks when the tensor is rotated to the geoelectric strike angle and plots as an oval for 2D, and shows oblique shapes for 3D data. The amplitude of a diagonal component is always zero for 1D, is zero when rotated parallel or perpendicular to the geoelectric strike and plots as a four petal flower for 2D, is axially symmetric for symmetric 3D, and shows oblique shapes for asymmetric 3D data.
Figure 2.15: Rotated transfer functions of the diagonal and off-diagonal components according to data dimension. The Swift skew is zero for the 3D example with a fully symmetric diagonal polar diagram (third from the left), but the Bahr skew is non-zero. Modified from Berdichevsky & Dmitriev (2008).

The impedance polar diagrams for the full SERPENT dataset are shown in Figure 2.16. The data portray a smoothly evolving dimensionality that correlates with changes in seafloor geology. The sites furthest offshore on the abyssal plain produce mostly 1D diagrams that become increasingly 2D with proximity to the outer rise. Within the trench and the outermost forearc, the data suddenly shift to 3D. The data return to 1D/2D on the upper slope and shelf of the Caribbean plate.

The geoelectric strike of the data can be defined as the angle that maximizes the amplitude of the off-diagonal polar diagram. The amplitude strike directions are plotted in Figure 2.17. The majority of the data between periods of 20-1000 s are either 1D or 2D, and when 2D the peak amplitude often occurs at an angle that is nearly

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Figure 2.16: Impedance polar diagrams of the SERPENT data.
Figure 2.17: Strike angle that maximizes $Z_{xy}$ polarization of the impedance polar diagrams.
perpendicular or parallel to the strike of the trench-axis. This is expected since the electrical structure of oceanic plates is likely controlled by plate age and the relative direction of plate motion. At periods greater than 1000 s, an interesting and unexpected shift in the strike of 2D data occurs at the majority of Cocos plate sites. The strike appears to rotate such that the maxima is aligned in an east-west direction. Although most of the data remain 2D, the shift in strike implies that the deeper electrical structure is independent of the shallower structure. Inverting the longer period east-west striking data with the shorter period plate motion strike data is difficult to reconcile with the limitations of 2D modeling, and requires 3D inversion.

Both Swift skews and polar diagrams only consider the impedance amplitudes. Amplitude-only measures potentially suffer from galvanic distortion and thus may produce unreliable estimates of dimensionality and geoelectric strike (Chave and Jones, 2012). This is a considerable limitation for land-based MT data due to the inherent heterogeneity of near-surface electrical structure that can take on a wide-range of spatial scales. For marine-based MT, most regions of the seafloor are sheltered from galvanic distortions by a thick cover of conductive sediment. Compared with galvanic effects, distortions caused by 3D bathymetry and particularly 3D coastlines are a far more concerning (Wheelock, 2012). Nonetheless, caution is warranted.

2.4.2 Phase-based dimensionality

The phase-sensitive skew, or Bahr skew (Bahr, 1988), is a measure of dimensionality that considers the imaginary part of the impedance tensor:

\[
s_B = \sqrt{2} \frac{|\text{Im}(Z_{yx}Z_{xx}^* + Z_{yy}Z_{xy}^*)|}{|Z_{xy} - Z_{yx}|}
\]

where the asterisk denotes the complex conjugate. Like the Swift skew, the Bahr skew is also a quotient of rotational invariants and thus is independent of the data strike. Phase-based measures are insensitive to local galvanic distortions precisely because they consider the imaginary part of the impedance tensor. While the robustness of the Bahr skew to distortions eliminates a critical weakness inherent to the Swift skew and
other amplitude-based measures, it is still susceptible to bias from noise, a characteristic for all real data.

Figure 2.18 shows the Bahr skew for the SERPENT dataset. Like the Swift skew, the trench sites all have unusually large Bahr skews. However, unlike the Swift skew, the Bahr skew exceeds 0.3 for Caribbean plate sites s38-s49 at periods longer than 512 s. Overall, there is consistent agreement between the Bahr skew and the impedance polar diagrams.

The Swift skew and Bahr skew and their many offspring are limited by the fact that they are derived from the presumption of 2D regional structure. The phase tensor is the first measure to be proposed that makes no underlying assumption about subsurface dimensionality, nor does it assume that the H or E fields are undistorted. Like other phase-based measures, it is a robust indicator of galvanic distortions. Below, I give a short description of the phase tensor. For detailed derivation and discussion, consult Caldwell et al. (2004) and Booker (2013).

The phase tensor is defined as:

\[ \Phi = X^{-1}Y = \begin{bmatrix} \Phi_{xx} & \Phi_{xy} \\ \Phi_{yx} & \Phi_{yy} \end{bmatrix} \] (2.74)

where \( \Phi \) is the real-valued phase tensor and \( X \) and \( Y \) are the real and imaginary parts of the impedance tensor \( Z \). A distorted impedance tensor can be decomposed into undistorted and distorted parts:

\[ Z_D = DZ = DX + D(iY) \] (2.75)

where \( D \) is a real-valued non-singular distortion tensor. Although we cannot determine \( D \) without independent constraints, it is readily shown that the phase tensor is independent of \( D \):

\[ \Phi_D = DX^{-1}DY = X^{-1}D^{-1}DY = \Phi \] (2.76)
Figure 2.18: Phase-sensitive (Bahr) skew of the SERPENT MT data.
The realization of the phase tensor for a 2D impedance that is oriented parallel or perpendicular to the geoelectric strike is:

$$\Phi = \begin{bmatrix} \tan(\phi_{yx}) & 0 \\ 0 & \tan(\phi_{xy}) \end{bmatrix}$$  

(2.77)

where $\phi_{xy}$ and $\phi_{yx}$ are the phase angles of the transfer functions $Z_{xy}$ and $Z_{yx}$, respectively.

The skew angle measures the asymmetry of the phase tensor in the same way that the Swift skew and Bahr skew use rotational invariants to measure the asymmetry of the impedance tensor. The three relevant rotational invariants are the trace $tr(\Phi)$, the skew $sk(\Phi)$, and the determinant $det(\Phi)$:

$$tr(\Phi) = \Phi_{11} + \Phi_{22}$$  

(2.78)

$$sk(\Phi) = \Phi_{12} - \Phi_{21}$$  

(2.79)

$$det(\Phi) = \Phi_{11}\Phi_{22} - \Phi_{12}\Phi_{21}$$  

(2.80)

where the skew angle ($\psi$) is the skew normalized by the trace:

$$\psi = \tan^{-1}\left(\frac{sk(\Phi)}{tr(\Phi)}\right)$$  

(2.81)

From equation (2.77), we see that $\psi$ is zero for noise-free 2D data. Thus, any non-zero $\psi$ reflects 3D data.

The phase tensor can be represented graphically with elliptical diagrams. Since the phase tensor is real-valued, we can rotate it around a unit circle and plot the resulting ellipse. A simple way to express the phase tensor ellipse is with the length of the semi-major and semi-minor axes (Moorkamp, 2007), which are rotational invariants:
\[
\Phi_{\text{min}} = \left( \frac{\text{tr}(\Phi)^2 + \text{sk}(\Phi)^2}{4} \right)^{\frac{1}{2}} - \left( \frac{\text{tr}(\Phi)^2 + \text{sk}(\Phi)^2}{4} - \text{det}(\Phi) \right)^{\frac{1}{2}} \tag{2.82}
\]
\[
\Phi_{\text{max}} = \left( \frac{\text{tr}(\Phi)^2 + \text{sk}(\Phi)^2}{4} \right)^{\frac{1}{2}} + \left( \frac{\text{tr}(\Phi)^2 + \text{sk}(\Phi)^2}{4} - \text{det}(\Phi) \right)^{\frac{1}{2}} \tag{2.83}
\]

In addition to the length, we need the rotation of the axes:

\[
\alpha = \frac{1}{2} \tan^{-1} \left( \frac{\Phi_{12} + \Phi_{21}}{\Phi_{11} - \Phi_{22}} \right) \tag{2.84}
\]

\[
\theta = \alpha - \frac{\psi}{2} \tag{2.85}
\]

The angle \( \theta \) that maximizes \( \Phi_{\text{max}} \) is equal to the difference of the angle \( \alpha \) and half the skew angle \( \psi \). Thus, we can reconstruct the phase tensor from the variables that define the ellipse:

\[
\Phi = R(\theta)^{-1} \begin{bmatrix} \Phi_{\text{max}} & 0 \\ 0 & \Phi_{\text{min}} \end{bmatrix} R(\psi)R(\theta) \tag{2.86}
\]

In practice, skew angles \(|\psi| < 7^\circ\) are thought to reflect ‘quasi-2D’ impedance tensors.

Figure 2.19 shows the phase tensor ellipses for the SERPENT dataset, where the color fill is a measure of the absolute value of the skew angle. Caribbean plate sites with non-trivial Bahr skews (see Figure 2.18) show up as large phase skew angles as well.

The phase tensor offers many more useful constraints than just the skew angle. We can visualize changes in rotation and phase associated with crossing geological structures, as well as inspect the undistorted phases of \( \Phi_{\text{max}} \) and \( \Phi_{\text{min}} \) for signs of anisotropy (Heise et al., 2006). Figure 2.20 plots rectangles that are rotated to the angle \( \theta \) of equation (2.85) and colored to indicate the angle of the minor axis, \( \tan^{-1}(\Phi_{\text{min}}) \). We
Figure 2.19: Phase tensors of the SERPENT data with skew angle color fill.
Figure 2.20: Phase tensor rotation angles with $\Phi_{\text{min}}$ angle color fill.
Figure 2.21: Phase tensor rotation angles with the difference of $\Phi_{\text{max}} - \Phi_{\text{min}}$ color fill.
Figure 2.22: The subset of data excluded from inversion modeling based on dimensional analysis.
clearly see that the phase tensor variations are dominated by the trench. Figure 2.21 again plots the rotation angle $\theta$, but this time the color fill shows the absolute phase split between the major and minor axes, $\tan^{-1}(\Phi_{\text{max}}) - \tan^{-1}(\Phi_{\text{min}})$. The phase split increases with increasing period for sites located on the Cocos plate, possibly due to anisotropic structure at depth.

The phase tensors and Bahr skews are consistent with the impedance polar diagrams. The subset of data that I have excluded from 2D modeling is shown in Figure 2.22. We set an error floor of 10% for both TE and TM mode data.

### 2.5 Occam’s inversion

Now that we have defined a subset of data appropriate for 2D numerical computation, we must solve the inverse problem to determine the electrical resistivity structure that fits the data. In general, we seek to minimize the sum-squared misfit $\chi^2$ between the data and model. We make the inversion more stable by weighting the measured and predicted data by the errors:

$$\chi^2 = \sum_{i=1}^{N} \frac{1}{\sigma_i^2} \left[ d_i - \hat{d}_i \right]^2 = \|W(d - f(m))\|^2$$

(2.87)

where $d$ is the data vector, $m$ is the model parameterization, $\hat{d} = f(m)$ is the model response, and $W$ is a diagonal matrix of the data variance. A $\chi^2$ misfit that is equal to the number of data points reflects an inverse model that fits the data to within the confidence level of the errors. It is possible to drive the misfit lower but then the model would “overfit” the data, assuming the errors are perfectly realized. Hence, we refine our minimization to seek a reasonable target misfit ($\chi^2_*$) relative to the errors. We often quote the root mean square (RMS) misfit to describe how well the inverse model fits the data:

$$\text{RMS} = \sqrt{\frac{\chi^2}{N}}$$

(2.88)
since it is independent of \( N \), the number of data.

The MT inverse problem is complicated by the fact that it is non-linear, ill-posed (non-unique), and ill-conditioned (highly unstable). There are several approaches to deal with non-linear non-unique inverse problems, all of which apply some variation of Tikhonov regularization – a least squares regression with an additional constraint added in order to ‘regularize’ an ill-posed problem:

\[
\min \left[ \| Rm \|^2 + \lambda^{-1} (\| Wd - Wf(m) \|^2 - \chi^2) \right] \tag{2.89}
\]

where \( R \) is a user-defined regularization matrix that is a measure of the model ‘roughness’ and \( \lambda^{-1} \) is a smoothness term known as the Lagrange multiplier. The Lagrange multiplier is a trade-off parameter that modulates the balance between the data fit and model roughness. The objective of regularized inversion is to simultaneously minimize the sum of the data misfit and model roughness. There are any number of ways to define the measure of model roughness. For instance, a simple application of the regularization matrix is:

\[
R = \begin{bmatrix}
-1 & 1 & 0 & 0 & \cdots & 0 \\
0 & -1 & 1 & 0 & \cdots & 0 \\
0 & 0 & -1 & 1 & \cdots & 0 \\
\vdots & \ddots & \ddots & \ddots & \ddots & \ddots \\
0 & \cdots & 0 & -1 & 1
\end{bmatrix} \tag{2.90}
\]

where \( Rm \) yields a vector of the model gradient, the L_2 norm of which is the first-difference measure of roughness.

Our modeling tool of choice is Occam’s inversion. Occam was developed for 1D MT inversion and expanded to 2D inversion by Scripps principal investigator Prof. Steven Constable \( (\text{Constable et al., 1987; de Groot-Hedlin and Constable, 1990}) \). Occam searches for the best-fitting model iteratively following the Gauss-Newton approach, which linearizes around a starting model with a Taylor expansion to compute the next
iteration step, and then repeats until convergence:

\[ f(m_{i+1}) = f(m_0 + \Delta m) = f(m_0) + J\Delta m + O(\Delta m^2) \] (2.91)

where the Jacobian:

\[
J = \frac{d}{dm} f(m) = \begin{bmatrix}
\frac{\partial d_1}{\partial m_1} & \cdots & \frac{\partial d_1}{\partial m_M} \\
\vdots & \ddots & \vdots \\
\frac{\partial d_N}{\partial m_1} & \cdots & \frac{\partial d_N}{\partial m_M}
\end{bmatrix}
\] (2.92)

is the matrix of partial derivatives of model responses with respect to model parameters.

The Gauss-Newton method requires a user-defined Lagrange multiplier. The beauty of the Occam method is that it performs a search over the Lagrange multiplier space in each iteration. When the inversion converges on a pre-defined misfit, the Occam algorithm then begins to search for a model with the smallest Lagrange multiplier that still satisfies the target misfit. In effect, it is searching for the smoothest model that fits the data to \( \chi^2_* \), hence the reference to Occam’s razor. This is a significant advantage over Gauss-Newton because it removes the user’s subjectivity of what combination of Lagrange multiplier and converged \( \chi^2 \) misfit to deem as best.

There are various ways to format the data for inversion: the real and imaginary components of the impedance tensor, the linear apparent resistivity and linear phase, or the log apparent resistivity and linear phase. We invert for the log apparent resistivity and linear phase of the impedance tensor because its realization of the misfit space is significantly smoother and better behaved than any of the other data formats (Wheelock et al., 2015), maximizing the likelihood that the inversion converges on the global minima.

### 2.6 2D inversion with MARE2DEM

I employ the publicly available state-of-the-art MARE2DEM (Modeling with Adaptively Refined Elements for 2D EM) software package to invert the SERPENT
MARE2DEM was developed by Scripps principal investigator Prof. Kerry Key (Key and Ovall, 2011; Key, 2012). The model domain is discretized using unstructured finite element grids in order to accurately incorporate bathymetry, which is essential for 2D modeling since marine data are strongly dependent on seafloor relief. The discretization is implemented using dual grids, one each for the inversion solver and the forward solver. The user only need design the inversion grid, with the freedom to allocate arbitrary polygonal elements of any shape and size. MARE2DEM automatically populates the inversion grid with the minimum number of conforming triangular elements and proceeds to solve the forward problem iteratively, applying a goal-oriented error estimator with adaptive mesh refinement to ensure accurate response solutions (Key and Ovall, 2011). The forward solver is parallelized because the forward problem is considered ‘embarrassingly parallel’.

MARE2DEM adopts Occam’s inversion and solves for isotropic to triaxially anisotropic electrical conductivity. In addition to MT data, it can handle electric and magnetic dipole data for controlled-source applications, including joint inversion for any combination of the various data types. MARE2DEM is a monumental advancement over previous generations of 2D EM numerical modeling tools. On 64 processors, the isotropic inversion of 1.6k SERPENT MT data with a 12k element mesh reached a target RMS misfit of 1.0 in under one hour. The inversion converged on the smoothest model in 22 iterations and two hours time. This is at least two orders of magnitude faster than the older generation Occam2DMT solver, which is limited to structured grids, a single CPU, and isotropic inversions. For the same number of data and model parameters, Occam2DMT required 45 hours and 37 iterations to reach an RMS of 1.1, and was unable to converge on the pre-defines misfit of RMS 1.0.

In addition to the much needed increase in speed, MARE2DEM is able to handle bathymetry with ease. Figure 2.23 compares a segment of outer rise bathymetry as meshed by Occam2DMT and MARE2DEM. Although both meshes contain ~12k elements over the full model domain, MARE2DEM trumps Occam2DMT when it comes to accurately capturing bathymetry.

Due to the decreasing resolution of MT data as a function of depth, we typically design our model discretization to have larger element dimensions with increasing
Figure 2.23: Discretized bathymetry in an Occam2DMT structured grid (top panel) and a MARE2DEM unstructured grid (bottom panel).

depth. Therefore, implementing the unweighted first-difference regularization in equation (2.90) would have the unwanted consequence of penalizing the roughness of smaller elements more than larger elements. This has a tendency to bias the model such that conductivity variations are emplaced at greater depths, where the elements are larger and the corresponding roughness penalty is smaller.

Not only do we discretize our models with a wide range of element sizes, but the element aspect ratios may vary considerably. The unweighted first-difference will tend to smear isolated anomalous structures in the direction of the longest dimension. For instance, a model discretization that is populated with vertically exaggerated elements (many times taller than they are wide) will noticeably smear structures vertically. This is illustrated in Figure 2.24, which shows an isotropic SERPENT MT model inverted with Occam2DMT using a structured mesh; several conductive and resistive structures are significantly smeared.

To achieve isotropic first-difference smoothing, we simply scale $\mathbf{R}$ by the aspect
Figure 2.24: Smoothing with unweighted regularization generates smeared structure in the Occam2DMT inversion model (top panel). The smearing is due to large aspect ratios in the structured mesh (bottom panel).
Figure 2.25: Smoothing with weighted regularization to account for parameter aspect ratios. We ran the same inversion as in Figure 2.24 but with the weighted regularization to show that artificial smearing is no longer an issue.

ratio of individual elements of $\mathbf{m}$. For a vertically exaggerated rectangular grid, this translates to increasing the roughness penalty between vertical touching elements by the ratio of height over width and decreasing the roughness penalty between horizontal touching elements by the ratio of width over height. Figure 2.25 shows the inversion using the scaled roughness for the same exact data and mesh used in the inversion shown in Figure 2.24. The changes are significant, and while we may find the latter model more pleasing, this also demonstrates the subjectivity of assigning roughness since both models fit the data equally well.

For oceanic plates, it is reasonable to assume that the first-order conductivity structures will be horizontally layered. Thus, we often prefer to use rectangular elements with horizontally exaggerated aspect ratios near the surface that become more isotropic
Figure 2.26: The converged isotropic MARE2DEM inversion model. Fits the data to RMS 1.0.

with depth. More importantly, we use a regularization that triples the roughness value of horizontal conductivity gradients with respect to vertical gradients in order to drive the inversion to converge on a horizontally smooth model. Outside the area of interest and at greater depths, we use triangular elements to minimize the total number of parameters. The MARE2DEM converged isotropic inversion (RMS 1.0) is shown in Figure 2.26 and the mesh discretization is shown in Figure 2.27.

At the trench axis, the isotropic inverse model displays a vertical conductive channel stretching to depths of 50 km, where it connects with a separate horizontal conductive channel that extends seaward into the Cocos plate. The anomalous feature splits the otherwise continuous resistive oceanic plate in half. We do not believe the vertical channel to be real because there are no MT data within the trench to constrain it. In fact, there is a 30 km data gap centered on the trench axis. Thus, we conclude that
Figure 2.27: A closeup view of the MARE2DEM discretized model domain used to invert the data. The mesh extends several hundred kilometers more than shown to make sure the boundary conditions are not violated.
Figure 2.28: Converged anisotropic MARE2DEM inversion model. $\rho_y$ is the resistivity along the horizontal axis, $\rho_x$ is the resistivity along the axis going in and out of the page, and $\rho_z$ is the resistivity along the vertical axis. Fits the data to RMS 1.0.
we cannot confidently interpret the isotropic model, and that the data require either anisotropic or 3D inversion.

For triaxial anisotropic inversions, we must penalize the model roughness in all three axis directions as well as the roughness between them:

\[
\| \mathbf{Rm} \|^2 = \| \mathbf{Rm}_x \|^2 + \| \mathbf{Rm}_y \|^2 + \| \mathbf{Rm}_z \|^2 + \gamma \left( \| \mathbf{m}_x - \mathbf{m}_y \|^2 + \| \mathbf{m}_y - \mathbf{m}_z \|^2 + \| \mathbf{m}_z - \mathbf{m}_x \|^2 \right)
\]

(2.93)

where \( \gamma \) is a scaling parameter that weights the anisotropy penalty. MARE2DEM implements the anisotropic scaling so that \( \gamma \) can range from 0 to 1. For \( \gamma = 1 \) the anisotropic and isotropic inversions are equivalent and for \( \gamma = 0 \) the inversion is independently modeling the TE and TM modes. Hence, we allow progressively more anisotropy into our model as we decrease the user-defined \( \gamma \) value.

The three principal axes of our preferred anisotropic inverse model are shown in Figure 2.28. The inversion converged to RMS 1.0 with \( \gamma = 0.75 \). In addition to being smoother, this is the least anisotropic model that no longer requires the conductive artifact seen at the trench axis of the isotropic model.

The text of Chapter 2, in part, is a reprint of the material as it appears in Naif, S., Key, K., Constable, S., and Evans, R. L. (2013). Melt-rich channel observed at the lithosphere-asthenosphere boundary. Nature, 495(7441), 356-359. The dissertation author was the primary investigator and author of the published work.
2.7 References


Chapter 3

Melt-rich lithosphere-asthenosphere boundary revealed with 2D MT inversion

The SERPENT survey was specifically designed to extend more than 150 km seaward of the trench axis, far from the significant bathymetric relief of the outer rise. Our goal was to quantify the impact of outer rise faulting on crustal porosity and fluid migration pathways. We extended the instrument array far onto the abyssal plain to image what we expected to be unremarkable conductivity beneath the oceanic plate, which would provide a baseline for comparisons with conductivity features found beneath the trench-outer rise and continental margin.

On the contrary, the abyssal plain data led us to make an intriguing discovery.

3.1 Anisotropic inversion results

Figure 3.1a shows the resistivity in the direction parallel to the survey strike ($\rho_y$) and Figure 3.1b the anisotropy for the survey parallel over trench parallel resistivity ratio ($\rho_y/\rho_x$). All three tensor components were shown in Figure 2.28. As previously detailed, the survey strike is perpendicular to the trench axis and plate-age isolines. Note that although I will refer to the $\rho_y$ axis as the plate-motion parallel resistivity, it is slightly oblique ($\sim15^\circ$) to the plate-motion direction.
Figure 3.1: Resistivity model obtained from anisotropic inversion of the seafloor magnetotelluric data. At the top is the surface view; arrows show the direction of north and $\rho_y$ (see below), and triangles denote seafloor magnetotelluric station locations.  

**a**, The electrical resistivity in the direction parallel to the survey profile ($\rho_y$). The resistivity color scale is logarithmic, with blue and red colors corresponding to resistive and conductive (less resistive) features, respectively. The dashed red line is the plate interface from Slab 1.0 (Hayes et al., 2012). Earthquake hypocenters from up to 50 km off-axis are shown as black circles (from the USGS/NEIC catalogue). The region enclosed by the dashed black line is where the model is at least 1.5 times more conductive in the direction parallel to plate motion.  

**b**, Horizontal anisotropy ratio. The color scale is in $\log[\rho_y/\rho_x]$ and shows the moderate anisotropy of the conductive layer at 45-70 km depth (red regions beyond 150 km offshore). Although the lithosphere overlying the conductive layer shows a strong anisotropy, we warn that this is not well constrained because the magnetotelluric method is primarily sensitive to conductive rather than resistive features (Evans et al., 2005). The deeper mantle beneath the conductive layer is isotropic.
Landward of the trench the MT data reveal a resistive subducting slab, with a notable correlation between the location of earthquakes and high resistivity. The widespread distribution of earthquakes indicates that the brittle slab is rupturing over a wide depth range rather than being concentrated along the plate interface, consistent with high resistivity indicating a relatively cold, fluid-free mantle.

Seaward of the trench, a veneer of low resistivity sediment and extrusive volcanics overlie a highly resistive lithosphere, typical of oceanic plate structure (Cox et al., 1986). Underneath the resistive oceanic lithosphere, the MT data reveal an anisotropic high conductivity layer confined to depths of 45-70 km. It extends at least to the western edge of the survey profile where our MT array ends, while its eastern edge lies beneath the trench-outer rise wall. The anomalous conductor is 4-6 Ωm in the plate motion parallel direction ($\rho_y$) and 8-10 Ωm in the perpendicular direction ($\rho_x$). The deeper mantle is isotropic and approximately 10 Ωm.

Figure 3.2 shows three model perturbation tests that were performed to investigate the sensitivity of the converged model in the region beneath the Cocos plate. We perturbed the resistivity of the conductive channel to half (2-4 Ωm) and twice (8-12 Ωm) its original value. This increased the RMS misfit by 10% and 5%, respectively. However, the increased misfit is dampened by data from the Caribbean plate that have little sensitivity to structures beneath the Cocos plate, including the conductive channel. By only considering data from instruments located on the Cocos plate, the misfit increased by 17% and 12%. The deeper isotropic mantle was perturbed to 5 and 30 Ωm, which increased the misfit by 5% in each case. The data have limited sensitivity below the conductor, which is not unexpected since the channel should attenuate the source signal. Last, we extended the conductive channel into the margin so as to mimic following the subducting slab to 100 km and 130 km depths. The misfit increased by 19% and 20%, respectively. Our tests confirm that the data have strong sensitivity to the first-order structural features we seek to interpret and moderate sensitivity to the mantle below the conductor.

Since MT data are primarily sensitive to conductance (also known as the conductivity-thickness product), the response of the observed 25 km channel is equivalent to that of a 2-3 Ωm 12.5 km thick channel. However, we favor the layer thickness
Figure 3.2: Model sensitivity tests. We altered the resistivity by an equal magnitude for all three axes in order to conserve the anisotropy of the original model. Tests were performed to investigate the data sensitivity to (A) the conductive channel, (B) the underlying isotropic mantle, and (C) the possibility of a subducting channel.

observed in the converged inverse model since the larger resistivity allows a more realistic interpretation. It is unlikely that the layer is thicker since the inversion machinery is designed to yield the smoothest possible model, and thus would smear structure where the data lack sensitivity. The fact that we are able to resolve the base of the conductor is evidence that it is required by the data.

As will be shown in the following sections, the resistivity and anisotropy signature of the conductive channel leads us to interpret it as a sheared partially molten layer trapped at the base of the lithosphere. We infer that the channel is likely to exhibit low viscosity, decoupling the overlying lithosphere from the deeper convecting mantle. Because such a boundary layer has the potential to behave as a lubricant to plate motion, its proximity to the trench may have new implications for subduction dynamics.

3.2 Electrical properties of the oceanic upper mantle

In the mantle, conduction is dominated by thermally-activated diffusion of ions through point defects in the crystal lattice structure (Yoshino, 2009). Most laboratory studies consider olivine as a proxy for the electrical properties of the mantle. Since
conduction in the mantle is thermally-activated, the resistivity of dry unaltered olivine varies as a function of temperature and oxygen fugacity (Schock et al., 1989). Dry olivine resistivity decreases from $\sim 10^6$ $\Omega$m at 700°C to 10$^2$ $\Omega$m at 1400°C for a quartz-fayalite-magnetite oxygen fugacity buffer (Constable, 2006). There are two ways to further reduce mantle resistivity: the addition of partial melt, and hydration.

Volatile-free basaltic melts are typically $10^{-1}$ $\Omega$m, which is two to three orders of magnitude less resistive than melt-free dry olivine at the same temperature. Melts become significantly more conductive as a function of water content, with resistivity as low as $10^{-1.5}$ $\Omega$m (Ni et al., 2011). Carbon dioxide also acts to enhance conduction in melts; at $10^{-2.5}$ $\Omega$m, carbonatitic melts are by far the most conductive yet known (Gaillard et al., 2008).

The conductivity of hydrated olivine increases with increasing water content due to hydrogen diffusion across lattice point defects, which operates via an ionic mechanism known as proton hopping (Karato, 1990). The first two laboratory studies to quantify the effect of water on olivine conductivity were published together in the same issue of Nature by independent groups and give conflicting measurements that differ by an order of magnitude (Wang et al., 2006; Yoshino et al., 2006)! Wang et al. (2006) measured polycrystalline olivine samples while Yoshino et al. (2006) measured single crystal olivine samples.

A following study from an independent research group brought up a critical point that is often overlooked. Poe et al. (2010) note that both Wang et al. (2006) and Yoshino et al. (2006) applied unpolarized Fourier Transform Infrared Spectroscopy (FTIR) with a generic calibration for water in glass (Paterson, 1982) to measure the water content in their olivine samples. First, using unpolarized radiation to measure an anisotropic medium such as olivine leads to inaccuracies; reliable measurements require polarized FTIR or better still Secondary Ion Mass Spectrometry (SIMS). Second, an olivine specific calibration for polarized FTIR determined by Bell (2003) demonstrates that the generic Paterson calibration underestimates the water content by a factor of 2.3. Bell (2003) also shows that unpolarized FTIR underestimates water content by a factor of 3.5, but is careful to point out that retrospective correction is more complex for unpolarized spectra than the simple factorization implies. These results were replicated by Mosenfelder (2006).
The latest greatest calibration is that of Withers et al. (2012), which concludes that the Bell calibration overestimates water content by a factor of 1.5.

Most recently, Gardès et al. (2014) compiled all data from the various laboratory groups (Wang et al., 2006; Yoshino et al., 2006; Poe et al., 2010; Dai and Karato, 2014a; Yang, 2012; Dai and Karato, 2014b) and used regression analysis to estimate a “unified” hydrous olivine conductivity law. The study attempts to compensate for inaccuracies in water content measurements by applying data errors based on the corrections of Bell (2003) and Withers et al. (2012).

### 3.3 A partial-melt channel at the LAB

The lithosphere-asthenosphere boundary (LAB) separates rigid oceanic plates from the underlying warm ductile asthenosphere. A viscosity decrease beneath this boundary is essential for plate tectonics. Three end-member models invoke either temperature (Stixrude and Lithgow Bertelloni, 2005), partial melting (Anderson and Sammis, 1970), or mantle hydration (Karato and Jung, 1998) as the mechanism responsible for rheological weakening in the Low Velocity Zone (LVZ).

Seismic studies intermittently identify a sharp velocity reduction – known as the Gutenberg (G) discontinuity – below oceanic plates at depths thought to coincide with the LAB (Beghein et al., 2014). The average sharpness (<30 km) and shear wave velocity reduction (~6.5%) would require unrealistic temperature gradients that are inconsistent with thermal models (Fischer et al., 2010). Consensus on the origin of the G-discontinuity remains elusive, with various studies generally invoking either partial melting (Kawakatsu et al., 2009) or a mantle dehydration boundary (Gaherty et al., 1999) as explanations. Additionally, there is ongoing disagreement as to whether its depth of onset is plate-age dependent (Rychert and Shearer, 2009; Rychert et al., 2012).

The anomalous conductor imaged with the SERPENT MT data is too conductive to be consistent with localized temperature variations, but could be explained by either a hydrated mantle or a small degree of partial melt. I estimate the hydration and partial melt fraction that is required to explain the geometric mean resistivity of the conductor (7.5 Ωm) for three representative mantle temperatures (1350, 1400 and 1450°C).
The conductivity of both hydrous olivine and melt vary as a function of water content. Since water does not partition equally between different minerals or between the solid and melt phase, the water concentration in the bulk mantle is not the same concentration found in olivine or basaltic melt. I use the mineral/mineral partition coefficients of Aubaud et al. (2004) to calculate the water contents in olivine and pyroxene phases. For 250 ppm H$_2$O dissolved in a bulk mantle composed of 60% olivine and 40% pyroxene, a total of 54 ppm H$_2$O will be partitioned into olivine and 540 ppm H$_2$O in pyroxene. I calculate the resistivity of hydrated mantle using the Hashin-Shtrikman upper bound mixing model and the resistivity of each mineral phase as a function of water content (Gardés et al., 2014; Dai and Karato, 2009). Figure 3.3a shows the mantle resistivity as a function of bulk water content. The red line is the geometric mean resistivity and the green shaded box is the resistivity of the anomalous conductor along the $\rho_y$ axis. We require 480 ppm H$_2$O at 1450°C or 840 ppm H$_2$O at 1350°C to explain the 7.5 $\Omega$m conductor with mantle hydration.

I use the mineral/melt partition coefficients of Hirschmann et al. (2009) to calculate the water content in basaltic melts, which depends on the H$_2$O concentration in the pre-melt mantle as well as the fraction of host rock that melts. For 50-250 ppm H$_2$O mantle melted to 0.5-5% levels, the basaltic melt water content ranges from 0.1-2.0 wt% (1 wt% = $10^4$ ppm). I use the basaltic melt resistivity model of Ni et al. (2011) to calculate mantle resistivity as a function of melt water content with the parallel mixing model. Figure 3.3b shows the bulk mantle resistivity as a function of melt porosity. At 1450°C, 0.8% partial melt is needed to reach 7.5 $\Omega$m in 250 ppm H$_2$O mantle, or 2.2% partial melt in 50 ppm H$_2$O mantle. At 1350°C, 1.6% partial melt is needed to reach 7.5 $\Omega$m in 250 ppm H$_2$O mantle, or 4.0% in 50 ppm H$_2$O mantle.

We must investigate the validity of interpreting our observed conductor as either partial melt or hydrated mantle. The necessary thermodynamic conditions that sustain partially molten peridotite depend on temperature, pressure, and bulk mantle volatile concentrations. I use the Stein & Stein plate cooling model, which assumes a 95 km plate thickness and 1450°C mantle potential temperature (MPT), and the Hasterok model, which assumes a 90 km plate thickness and 1364°C, to calculate the geotherm for 23 Ma oceanic lithosphere (Stein and Stein, 1992; Hasterok, 2013).
Figure 3.3: Evidence for the stability of melt. a, Hydrous mantle resistivity is estimated as a function of H$_2$O content for different temperatures. The green shaded region represents the observed resistivity of the conductive channel along the $\rho_y$ axis (4-6 $\Omega$m), while the red line represents its geometric mean resistivity (7.5 $\Omega$m). In the absence of melt, the bulk mantle must contain at least 480 ppm H$_2$O to account for our observations. b, Bulk resistivity of partial melt shown as a function of melt fraction for different temperatures and bulk mantle water contents, assuming a calculated mineral/melt partition coefficient of 0.008.
We need to quantify the solidus of mantle peridotite to determine the depth at which melting begins. For dry mantle, I use the solidus of Hirschmann (2000). For damp mantle, I use the solidus of Hirschmann et al. (2009), which applies the cryoscopic approximation using the molar entropy of fusion formulation. I assume an oxide molar unit (59 g/mole). I update the pyroxene partition coefficients in Hirschmann et al. (2009) with results from O’Leary et al. (2010), which account for effect of aluminum.

Figure 3.4 shows that the warmer geotherm exceeds the 250 ppm H$_2$O solidus at about 50 km depths, signifying the minimum water content required for the stability of melt to roughly coincide with our observed high conductivity layer. The colder Hasterok geotherm requires twice as much water. Recall that to explain the origin of the conductor with mantle hydration, we require at least 480 ppm H$_2$O at 1450°C. Figure 3.4 illustrates that such a large concentration of water would lead to significant partial melting in the 45-70 km depth range. This leads us to conclude that the conductive layer must contain partial melt arising from a mantle with a lower (more reasonable) degree of hydration.

It is more informative to precisely quantify the amount of water required to induce melting. To do so, first I vary the Stein & Stein MPT from 1300-1500°C and record the mantle temperature between 45-60 km depths. Then, I calculate the amount of water needed so that the peridotite solidus and the mantle temperature match. The results are shown in Figure 3.5. It is clear that increasing the depth to the top of the melting region from 45 to 50 km significantly reduces the amount of water required. The resolution of MT data at 50 km depths is on the order of 5-10 km. Furthermore, our inversion algorithm converged on the smoothest model that fits the data to RMS 1.0, in which the conductor is brightest near 55 km depths and gradually becomes resistive above and below. Therefore, within the uncertainty of our data and possible smearing of the conductor, it is plausible that the depth to the top of the conductor is ∼50 km, requiring 230 ppm H$_2$O.

Estimates derived from mid-ocean ridge basalt (MORB) samples constrain the upper mantle to 50-200 ppm H$_2$O. Geochemical analysis of melt inclusions can reliably constrain mantle volatile contents; melt inclusion samples from basaltic glass erupted at the EPR were used to infer 142 ± 85 ppm H$_2$O in the upper mantle (Saal et al., 2002),
Figure 3.4: High asthenosphere conductivity explained by a thin partially molten layer. Solid lines depict the solidus of dry and wet peridotite for various mantle H$_2$O contents. The dashed lines show the geotherm derived from two plate cooling models for 23 Ma oceanic lithosphere. Melt is stable at depths greater than 47 km for peridotite with 275 ± 85 ppm H$_2$O and a 1450°C mantle potential temperature. A colder 1364°C MPT geotherm requires 500 ± 155 ppm H$_2$O. The solidus of wet peridotite is calculated with the cryoscopic approximation assuming an oxide molar mass (Hirschmann et al., 2009).

just shy of the 230 ppm H$_2$O needed to sustain partial melts at 50 km depths. However, uncertainties in partition coefficient measurements may accommodate this discrepancy, where the statistical lower bound of 160 ppm H$_2$O is well within the range of EPR-MORB estimates. We do not consider the effect that other volatiles have on melting, specifically CO$_2$, which is known to significantly reduce the solidus (Hirschmann, 2010; Dasgupta et al., 2013). In fact, a recent study interprets our observed conductor as incipient carbonatite melts (Sifré et al., 2014), but the authors must assume the mantle contains 400+ppm CO$_2$, significantly more than the 100 ppm CO$_2$ found in EPR-MORB samples (Saal et al., 2002).
Figure 3.5: Dependence of melt stability on water concentration and mantle temperature. The four curves define the amount of water and the MPT required for the dehydration melting to coincide with 45-65 km depths. The depth to the top of the stable melt zone is highly sensitive to water concentration and temperature.

In addition to bulk mantle water content, the onset of melting is highly sensitive to mantle temperature. Current consensus places the MPT at \( \sim 1350^\circ \text{C} \). However, this is a global average and is not necessarily applicable to the mantle beneath the Cocos plate. In fact, a gravity and bathymetry study located on the Cocos plate at the northern EPR documented anomalous subsidence rates and a high density of seamounts, indicating that some combination of thinner lithosphere, warmer mantle, and off-axis partial melts is required (Cormier et al., 2011). Figure 3.6 illustrates that in the vicinity of the SERPENT survey the bathymetry is anomalously shallow compared to the ideal seafloor depth predicted with plate cooling models (Crosby and McKenzie, 2009). On average, the seafloor is 500 m too shallow, and over 1 km too shallow at its peak! Thus, a warmer MPT is likely justified. Figure 3.7 shows that the region is also cluttered with a high density of seamounts, which is evidence for intraplate volcanism possibly fed by partial melt from the LAB. Notice that further to the north, the seafloor is smooth. This hints at
Figure 3.6: Anomalous bathymetry in the vicinity of our survey region is determined by subtracting the seafloor depth predicted with a plate cooling model from the observed depth. On 20 Ma Cocos plate, the seafloor depth is equivalent to the predicted depth for 6 Ma plate.

The possibility that the Galapagos hotspot may have something to do with the unusual bathymetry in the region.

The intersection of the geotherm with the solidus may culminate in a freezing front where melt solidifies due to the colder temperatures above. We infer that this freezing front forms a permeability barrier that traps buoyant melt rather than allowing it to percolate to shallower depths. This barrier is further reinforced by the higher viscosity of the uppermost mantle above since it has been depleted of volatiles during upwelling and melting at the mid-ocean ridge (Hirth and Kohlstedt, 1996). This upper boundary may also be caused by a sharp increase in the water solubility of mantle minerals, and thus a corresponding sharp decrease in the stability of partial melt (Mierdel et al., 2007). The depth-dependent viscosity of basaltic melt may also play a
Figure 3.7: Bathymetric map of the Cocos plate. Taken from recently updated ship-track and satellite data (Sandwell et al., 2014). In addition to being anomalously shallow, the seafloor in the vicinity of our survey region is littered with numerous seamounts. A plausible culprit is partial melt sourced at the LAB.

role in allowing ponding at particular depths (Sakamaki et al., 2013).

The electrical anisotropy we observe indicates the melt is made up of a network of tubes or elongated spheroids aligned in the direction of plate motion (Caricchi et al., 2011). The estimated conductivity of partial melt using a parallel mixing model is more appropriate than the Hashin-Shtrikman upper bound for anisotropic conductivity. Further reductions in melt estimates are possible, but require water concentrations that may be unrealistically high. For instance, 0.3% melt is attained for a 1450°C ambient mantle with 600 ppm H$_2$O. Likewise, carbonatite melt is significantly more conductive and thus yields even smaller fractions, but is an unlikely explanation since it is unstable at these depths (Hirschmann, 2010). Bulk conductivity in the deeper isotropic
asthenosphere is best described by the Hashin-Shtrikman upper bound, which predicts well-connected melt fractions of 0.2-1.0% for the observed 10-20 Ωm mantle.

3.4 Conclusions

Deep off-axis melt emplacement can occur during lithosphere formation at the ridge or from accumulation of a small degree of intraplate melting (Hirschmann, 2010). An earlier MT experiment at the southern EPR observed an off-axis conductive asthenosphere at 60-120 km depths (Evans et al., 2005; Baba et al., 2006). We interpret this conductive layer as requiring partial melt since its conductivity and anisotropy are similar to observations beneath our profile. That portion of the ridge generates the Nazca plate; if we assume equivalent ridge processes are occurring for the Cocos plate, then most of the hydrous melt has been emplaced at or near the ridge axis. Additional melt may accumulate through deeper intraplate melting caused by small-scale convection (Ballmer et al., 2007). Grain boundary migration of deeper partial melt (Zhu et al., 2011) carries with it increased water content as a result of the preferential partitioning of water into silicate melts. As water-enriched melt rises, it collects beneath the colder, less permeable lithosphere (Katz et al., 2006) and shears into a network of interconnected horizontally aligned melt bands (Kohlstedt and Holtzman, 2009), possibly driven by large scale asthenospheric flow (Höink et al., 2011). Stress gradients perpendicular to the shear direction have been shown to reduce melt connectivity in the shear perpendicular axis (Caricchi et al., 2011), offering a possible mechanism for our observed anisotropy. Thus, the depth extent over which the asthenosphere is being sheared can be inferred from the anisotropy and increased conductivity of the melt layer, which the MT data constrain to a maximum thickness of 30 km.

The existence of a horizontally extensive melt layer that is being sheared over a confined depth interval indicates the LAB is a thin low viscosity channel. Theoretical studies suggest that even small melt fractions (<1%) can lower viscosity by up to two orders of magnitude, effectively decoupling the lithosphere from the asthenosphere (Takei and Holtzman, 2009). The depth extent over which this decoupling occurs as well as its proximity to the trench axis has potentially new implications for plate dynamics, but
little is known since previous studies have focused on the effects of melt on mid-ocean ridge and mantle wedge dynamics. A pervasive feature that exists beneath oceanic plates at large distances from spreading centers requires the stability of a partially molten layer, as dictated by the solidus of wet peridotite and a warm mantle geotherm. A previous MT study of 140-150 Ma lithosphere in the Pacific Ocean Basin did not find a conductive melt layer (Matsuno et al., 2010). This is consistent with a thicker cold plate containing too little water to sustain hydrous melts at deeper LAB depths. However, the discovery of "petit-spot" seafloor volcanism at the outer rise of 135 Ma Pacific plate suggests the possibility that partial melt is a ubiquitous feature that exists beneath all oceanic plates (Hirano, 2006; Yamamoto et al., 2014). Furthermore, a recent active-source seismic study images a 10 km thin anomalous layer at the LAB of the 100 Ma Pacific plate subducting beneath New Zealand’s North Island that is strikingly similar to our conductive channel (Stern et al., 2015).

In our profile the conductive layer ends near the trench, while the layer anisotropy persists in a trajectory that follows the subducting slab, albeit with a decreasing anisotropic factor. This signifies that as the plate subducts, some melt is likely to remain in place beneath the shallower LAB of the trench outer rise due to its buoyancy. This offers an explanation in which the observed melt-rich LAB is caused by the concentration of low fraction melts at the trench over time as the plate continues to be subducted; if this mechanism is responsible for locally enriching the asthenosphere with melt, a prediction is that melt would be most concentrated near the trench and decrease towards younger regions of the plate.

The text of Chapter 3, in part, is a reprint of the material as it appears in Naif, S., Key, K., Constable, S., and Evans, R. L. (2013). Melt-rich channel observed at the lithosphere-asthenosphere boundary. Nature, 495(7441), 356-359. The dissertation author was the primary investigator and author of the published work.
3.5 References


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Chapter 4

Water-rich faults revealed at the Middle America Trench with 2D CSEM inversion

Subduction of hydrated oceanic lithosphere is the primary process by which water is transported to the interior of the Earth (Thompson and Connolly, 1992). The magnitude of water subducted modulates many fundamental tectonic processes, including seismic coupling at the megathrust plate interface (Saffer and Tobin, 2011) and arc magma genesis (Gaetani and Grove, 1998). Quantifying the flux of water transported by oceanic plates and the distribution of fluids released during subduction is critical to understanding the pattern of seismic coupling at the plate interface and the cycling of water between the solid and fluid Earth.

We present a comprehensive electrical resistivity image from 2D inversion of the marine CSEM data that reveals fluid-rich zones in the incoming oceanic crust, the plate interface, and the overlying crust at the MAT offshore of Nicaragua. Our image provides new constraints on three important processes: (1) the migration of seawater into the incoming oceanic crust along bending-induced normal faults at the outer rise, (2) the complete subduction of the incoming sediment layer into the forearc margin, and (3) sediment underplating and the migration of fluids from the plate-interface to fluid seeps and mud mounds on the outer forearc seafloor. In addition to the descriptive elements above, we estimate the porosity of the incoming oceanic crust and subducted
sediments from our resistivity model, allowing us to quantify to flux of pore water being subducted. We find that significantly more crustal pore water enters the margin than previously thought (up to a factor of two). In the margin, our estimates show sediment porosity that decays exponentially as it is subducted along the plate-interface. Existing constraints on subducted sediment porosity are derived from laboratory compaction studies and a limited number of boreholes drilled into the margin, whereas our study quantifies the porosity structure in the outermost 30 km of the forearc slope.

4.1 Fluid-tectonic processes in a subduction zone

Subducting tectonic plates transport interstitial and mineral-bound water that is partitioned into sediment, crust, and upper mantle reservoirs. Pore water trapped in benthic sediments can account for close to half of the total influx depending on the fraction that is underthrust into the margin with a subducting slab (Jarrard, 2003). The magnitude of volatiles in the uppermost portion of oceanic crust tend to increase with time, where older extrusive crust contains relatively more mineral-bound and less interstitial water than younger crust (Carlson and Herrick, 1990; Staudigel, 2014). Because the volatile content of the intrusive oceanic crust is generally assumed to remain constant with age, the total crustal water that is subducted to depth is estimated to vary by only 3% between 5 to 145 Ma plate. This suggests that the regional flux of crustal water is nearly uniform at all convergent margins (Jarrard, 2003).

Mineral-bound water in the oceanic upper mantle could rival or significantly exceed the combined flux of crustal and sedimentary sources in the global subduction zone water budget (Rüpke et al., 2004; Ulmer and Trommsdorff, 1995), but the degree of hydration is poorly constrained (Hacker, 2008). The upper mantle of an oceanic plate formed at a fast spreading ridge is expected to be relatively dry until it nears the trench, where flexural bending at the trench-outer rise generates Moho crossing normal faults that may provide a permeable pathway for seawater to percolate into the mantle (Peacock, 2001; Ranero et al., 2003; Faccenda et al., 2009). The addition of water to mantle peridotite triggers a series of alteration reactions that consume H₂O and form hydrous serpentinite minerals (Moody, 1976). To date, regional active-source seismic studies
provide the only in-situ estimate of mantle hydration at the outer rise (Ivandic et al., 2008; van Avendonk et al., 2011). By comparison, fluid fluxes in the sediments and extrusive crust are better constrained by the relatively abundant number of observations, but the large inter-study variability of crustal H$_2$O estimated to subduct suggests important uncertainties remain (Kastner et al., 2014).

The outermost forearc of subduction zones are marked by either accreting or non-accreting margins (von Huene and Scholl, 1991). In accreting margins, a fraction of the sedimentary layer overlying oceanic plates is scraped off at the trench axis while the remaining sediment is underthrust with the subducting oceanic plate. The mechanics of an accretionary wedge are described by the critical taper theory (Davis et al., 1983; Wang and Hu, 2006), which quantifies the slope failure criterion of sediments with a force balance for the effective compressive stresses. The critical taper angle is the sum of the seafloor slope angle and the subducting plate dip angle. Over time, the accreted sediments pile up and internally deform until reaching the critical taper, at which point they fail and slide at a steady state as sediments continues to be accreted. The process forms a large network of fold-and-thrust belts typically with a 6° taper angle and a 2° gently sloping seafloor (Clift and Vannucchi, 2004).

For non-accreting margins, most if not all of the incoming sedimentary layer is fully subducted. The lack of accretion translates to an overlying forearc composed of crystalline basement or ancient lithified accretionary wedge whose internal strength is far greater than in actively accreting margins. Stronger margins sustain greater internal deformation. Consequently, the taper and seafloor slope angle are steeper at 10° and 6° on average, respectively (Clift and Vannucchi, 2004). In some margins, including Nicaragua, the taper angle exceeds 15°.

Another noteworthy distinction between the two margin styles is the subducting flux of H$_2$O. Since sediment porosity drops off exponentially with depth, the accreted sediments happen to be the most rich in pore fluids. With most, if not all, of the sediments underthrust at non-accreting margins, the magnitude of H$_2$O subducted is nearly one order of magnitude larger than for accreting margins (Pichon et al., 1993). Interestingly, a key characteristic of non-accreting margins is their rapid convergence rate, which further adds to the net input flux of H$_2$O (Clift and Vannucchi, 2004). How
the fluids behave upon subduction is another matter.

Water released from the downgoing slab affects several processes depending on its source and migratory pathway. Water carried to depth with the slab generates arc magmas by lowering the solidus of mantle rocks (Kushiro et al., 1968; Gaetani and Grove, 1998) and may contribute to deep mantle hydration (Hacker, 2008). Water fluxed back into the ocean contributes to the long-term evolution of oceanic elemental compositions by driving land-ocean geochemical cycling and may also sustain seafloor and crustal microbial communities (Kastner et al., 1991; Staudigel, 2014; Santelli et al., 2008). Additionally, pore water may remain trapped within relatively impermeable sediments subducted with the downgoing plate (Screaton et al., 2002). Trapped water is increasingly loaded by the overlying forearc crust, leading to excess pore pressures that decrease the effective normal stress at the plate interface and therefore impact the megathrust fault stability (Scholz, 1998). An overpressured plate interface has been proposed to induce tremor and episodic slip (Beroza and Ide, 2011), as well as to facilitate rupture to the trench during large to giant earthquakes (Dean et al., 2010; Kimura et al., 2012; Noda and Lapusta, 2013). Geophysical observations confirm that some megathrust earthquakes indeed penetrate the up-dip aseismic zone and propagate to the trench seafloor, efficiently generating tsunamis (Lay et al., 2012).

With the onset of subduction, both mechanical and thermal forces expel water from the downgoing slab. Rapid burial and tectonic loading mechanically release pore water from sediments via compaction, generating the primary source of fluid from the trench to 7 km into the outermost forearc margin (Moore and Vrolijk, 1992). The steeper taper angle combined with faster convergence at non-accreting margins means that the sediments are loaded at a drastically faster rate than accreting margins. Add to that the considerably larger H\textsubscript{2}O flux and it is clear that non-accreting margins release vastly more fluids into the upper plate while sustaining larger pore pressures along a wider swath of the megathrust plate interface (Pichon et al., 1993).

Further down-dip, the subducting plate is progressively heated and loaded. When temperatures exceed 50°C, hydrated minerals begin dewatering due to diagenetic and metamorphic dehydration reactions. Smectite clay and biogenic opal in sediments dewater between 60-150°C, providing fluids that promote excess pore pressures (Saffer
and Tobin, 2011). One hypothesis is that the dewatering of sediments at non-accreting margins leads to such extreme overpressures that the base of the upper plate is hydraulically fractured, transferring material to the plate interface where it is subducted with the slab (von Huene et al., 2004a). This process is thought to be ongoing in Nicaragua (Ranero et al., 2008).

4.2 The SERPENT CSEM survey

In order to investigate sub-seafloor fluid migration and the pattern of plate hydration and dehydration, we collected CSEM data to image electrical resistivity along a 220 km profile offshore of Nicaragua that spans the abyssal plain and trench-outer rise of the Cocos plate and the forearc slope of the Caribbean plate, shown in Figure 4.1. The CSEM method uses electromagnetic induction to map crustal resistivity variations, providing a powerful tool that can image fluid pathways and quantify porosity. The electrical resistivity of the crust, to first order, is dependent on porosity and the temperature and salinity of pore fluids (Evans, 1994). Since saline water is up to six orders of magnitude less resistive than crystalline rock, slight variations in pore fluid content will change bulk resistivity above the detection threshold of our broadband electromagnetic instruments (Constable, 2013). Porosity is then estimated from resistivity by applying multi-phase mixing models given assumptions about pore geometry and interconnection (Becker et al., 1982).

Our instruments continuously record horizontal electric and magnetic field time-series in two orthogonal directions. While it is possible to utilize naturally occurring electromagnetic fields to image electrical resistivity with the MT method, attenuation of the high-frequency source energy by the highly conductive ocean limits the ability of marine MT data to constrain shallow structure. We have shown that the SERPENT MT results resolve upper mantle electrical structure at depths corresponding to the LAB (Naif et al., 2013). Considering that our primary objective is to image the crust, we augment the low frequency MT method with CSEM data collected at higher frequencies.

Acquiring CSEM data entails deep-towing an electric dipole antenna close to the seabed so that the transmitted energy couples to the seafloor, allowing it to propagate
Figure 4.1: Map of the CSEM survey. The faulted seafloor fabric is clearly seen in the high-resolution bathymetric map. White squares outlined in black show the location of ocean bottom electromagnetic receivers whose recorded CSEM data will be used here. The red line shows the CSEM transmitter towpath. The black dashed line is the region inferred to have high seismic coupling based on seismic slip inversions of the 1992 tsunamigenic earthquake (Ye et al., 2013). The magenta line shows the location of a nearby active source seismic survey that detected mantle penetrating bending faults (Ranero et al., 2003).

Transmitting large electrical currents down a cable that is several kilometers in length leads to significant energy loss due to Joule heating. It is far more efficient to supply a high voltage source. We step-up transform the ship’s power supply to a high voltage low current source and feed it down a 0.680” coaxial cable. A transmitter attached to the end of a coaxial cable step-down transforms the power to a high current source and feeds it to a dipole antenna that extends from the back of the transmitter. An array of receivers measure the attenuation of this propagating energy, which depends on the seafloor electrical resistivity structure (Cox et al., 1986).
We deep-towed the Scripps Undersea Electromagnetic Source Instrument (SUESI) approximately 100 m above the seafloor while outputting 300 amps of alternating current from the dipole of a 250 m antenna. We used a complex binary waveform with a 4 s fundamental period to ensure a wide frequency spectrum suitable for constraining crustal structure (Myer et al., 2011). The transmissions were recorded by the array of seafloor EM receivers sampling at 62.5 Hz. A cartoon diagram in Figure 4.2 describes a typical CSEM survey.

Seafloor EM instruments were navigated with a Long-Baseline (LBL) system using transducers installed on the hull of the vessel. Most receivers were navigated to better than 5 m accuracy, but we struggled to navigate a handful of sites located in the trench region with the LBL system. For the problematic sites, we used an analog system to range on the receivers. We determined the absolute position with the closest point of approach of two perpendicular ship tracks centered on the recorded instrument drop point.

SUESI was navigated with an inverted long-baseline (iLBL) system (Key et al., 2012). Two acoustic transponders were outfitted with GPS hardware to record their
absolute position and were deployed from the back of the vessel and towed on the sea surface. The transponders ping on SUESI at 12 kHz and SUESI pings back at 15.5 kHz. The GPS positions, travel times, and SUESI depths (recorded with a pressure gauge) were used to triangulate SUESI’s absolute position. In deep waters, the navigation is accurate to 10 m along the ship-tow direction and 100 m in the cross-line direction. The large cross-line uncertainty is problematic at receiver-transmitter offsets of less than ~2 km since it leads to modeling errors that exceed 10%.

The dipole altitude and dip angle are additional parameters needed for accurate CSEM modeling. SUESI is equipped with an altimeter to record altitude and the dipole antenna is equipped with a tail-end transponder (TET) and pressure gauge to record position and depth. We calculated dip using the TET/SUESI transponders to get the antenna separation length and the TET/SUESI pressure gauges to get the depth offset. Figure 4.3 shows SUESI’s altitude and dip along with the SUESI-derived seafloor bathymetry for the profile transect.

### 4.3 CSEM data processing and 2D inversion

Following the methodology outlined in *Myer et al.* (2011), we obtained CSEM response functions by transforming the time-series data into the frequency domain. We performed FFT on 4 s non-overlapping windows (equivalent to a single transmitted waveform) after applying a first-difference pre-whitening filter to remove unwanted low frequency signals. The FCs were then post-darkened and normalized by the transmitter dipole moment to yield the complex-valued CSEM data. Figure 4.4 shows the amplitude and phase of the 0.25 Hz inline electric field data from site s04. Notice the large scatter at lower amplitudes. We stack the resulting Fourier coefficients into 120 s intervals (30 FFT windows) to increase signal-to-noise. Stacking the data has the added benefit of providing data variance estimates.

The acoustic transponders at a moderate number of receivers were pinging as SUESI was being towed across them. An acoustic ping generates a large delta-like spike in the time-series that significantly distorts the FCs of that window. Figure 4.5 shows one such example from site s14. The SERPENT CSEM data are unusual in this regard;
the pinging distortions are far more frequent compared with other survey datasets. This was a side effect of simultaneously collecting multi-beam swath bathymetry data, which uses a 12 kHz signal that happens to be a common listening frequency for some of the transponders on our EM instruments.

We manually discarded any data infected by acoustic pings prior to stacking, a fairly simple task considering the amplitudes were off by at least one order of magnitude. The remaining data were stacked with a robust algorithm that iteratively flagged and removed data with residuals greater than three times the median absolute deviation until convergence. Responses with signal-to-noise ratios of less than two were discarded from further analysis. The cleaned stacked responses from two receivers are shown in Figure 4.6.
Figure 4.4: High-quality CSEM data at site s04. Both amplitude and phase evolve smoothly with few outliers. The data are shown in time. Navigation is used to determine the receiver-transmitter offsets, which are required to model the data.
Figure 4.5: Distorted data at site s14 due to accidental triggering of the receiver’s transponder, which produces an acoustic ping that strongly affects both amplitude and phase. Prior to stacking, we manually filtered outliers for SERPENT sites that suffered from acoustic noise.

The subset of data determined suitable for 2D modeling consist of inline electric field responses (amplitude and phase) at the 1st, 3rd, and 7th waveform harmonics (0.25 Hz, 0.75 Hz, and 1.75 Hz), where most of the energy is concentrated. Inverting multiple frequencies of data provides exponentially better model resolution/sensitivity compared with single frequency inversions (Key, 2009). We limited the data to the electric field component because magnetometers are susceptible to noise associated with instrument shaking and using both electric and magnetic field data adds redundant constraints (Key, 2009). The only exception is site s30, which had a malfunctioning electrode sensor but functional magnetometers. Lastly, we discarded all data at offsets of less than 2 km due to the increased error associated with navigation uncertainties. The short range data are mostly sensing the conductive sedimentary layer and so do not reduce the sensitivity to structure at crustal depths.

Accurate bathymetry is critical for CSEM modeling since the data are highly sensitive to seafloor relief, as exemplified in the amplitudes and phases at site s27, where SUESI traversed large fault scarps (see Figure 4.6). We utilized the bathymetry derived from SUESI (shown in Figure 4.3a) to finely discretize our model mesh. SUESI is equipped with an altimeter and pressure gauge that can be combined to map the seafloor relief with higher accuracy than even multi-beam swath bathymetry can offer.
Figure 4.6: Inline electric field amplitude and phase data at 0.25, 0.75 and 1.75 Hz from sites s09 and s27. Vertical bars are data uncertainties. The 2D model responses (black lines) fit the data remarkably well.
We inverted for a two-dimensional isotropic resistivity model from the multi-frequency CSEM responses with MARE2DEM (Key and Ovall, 2011; Key, 2012). Figure 4.7 shows the converged electrical resistivity model, which fits the data to RMS 1.0 relative to a 2% error floor. Figures 4.8 and 4.9 show matrix plots of the amplitude and phase data and model responses. The inversion of the high-frequency CSEM data captures the electrical structure of the Cocos crust with significantly higher resolution than previous efforts with marine MT data (Naif et al., 2013; Worzewski et al., 2010).

4.4 Water-rich bending faults of the outer rise

The seafloor spreading fabric of the incoming Cocos plates is oriented parallel to the trench such that flexural bending at the trench-outer rise optimally reactivates a dense network of normal faults that have been imaged to extend several kilometers into the upper mantle (Ranero et al., 2003). Bending faults are thought to provide fluid pathways that lead to hydration of the upper mantle (Peacock, 2001; Faccenda et al., 2009). While geophysical anomalies detected beneath the outer rise have been interpreted as broad crustal and upper mantle hydration (Ivandic et al., 2008; van Avendonk et al., 2011; Key et al., 2012), no observational evidence exists to confirm that bending faults behave as fluid pathways.

Our model depicts spatially evolving electrical resistivity that correlates with the location and intensity of outer rise bending faults. Beyond 60 km seaward of the trench both the bathymetry and resistivity are relatively one-dimensional. The electrical structure reflects compositionally distinct layers; a veneer of conductive high porosity sediment is underlain by an increasingly resistive and less porous crust and upper mantle. At the onset of dense faulting approximately 60 km from the trench, a heterogeneous electrical structure emerges. Seafloor fault scarps correlate with sub-vertical conductive channels that extend into the lower crust, readily observed in Figure 4.10. Because these conductive channels occur in step with normal faults and a porosity increase is the simplest explanation for the enhanced conductivity, we infer that seawater penetrates into the crust along the fault planes. Fluidized faults can also account for fault-parallel electrical anisotropy that was previously detected at the outer
Figure 4.7: The electrical structure of the incoming Cocos plate from nonlinear inversion of deep-towed CSEM data. The vertical cross-section shows the sub-seafloor electrical resistivity structure and the stitched upper panel shows seafloor bathymetry. The dark blue cubes show the location of EM receivers. The region of the seafloor marked by steeply dipping relief correlates with sub-vertical conductive channels, which we interpret as evidence for the migration of seawater along bending faults.
Figure 4.8: Amplitudes from the inverted data and the model responses in the left and right columns, respectively.
Figure 4.9: Phases from the inverted data and the model responses in the left and right columns, respectively.
Figure 4.10: Close up of outer rise electrical structure. Low resistivity channels associated with fault scarps require significant crustal hydration. The black filled circles show EM receiver locations, the gray dashed line is the Moho, and the white dashed lines shows contours of seismic P-wave velocity reductions interpreted to be serpentinized mantle (Ivandic et al., 2008).

rise (Key et al., 2012). As the plate approaches the trench axis, the anomalous channels deepen and become more conductive, suggesting that additional fluids percolate to greater depths as the fault throw grows and the crust is further damaged.

In order to investigate the sensitivity of CSEM data to conductive fault channels at the outer rise, we construct two synthetic resistivity models that contain the same bathymetry and data density as our inverted data set. Figure 4.11a shows a model of a simple 1D layered half-space with no conductive fault channels. The layer depths and resistivity are computed from the expected porosity of 24 Ma oceanic crust with 400 m of sediments (Jarrard, 2003). The second model, shown in Figure 4.12a, adds three conductive channels which represent porous faults that penetrate the upper crust, the lower crust, and the upper mantle. We calculate forward CSEM responses for the synthetic models, add 2% random noise, and invert the resulting data with a 2% error floor. The two synthetic inversions converge to a final RMS of 1.0. The model results
in Figures 4.11b and 4.12b show that the CSEM data are sensitive to the presence of conductive channels to upper mantle depths, but progressively lose resolution below mid crust depths.

Although we image resistivity reductions in the outer rise crust, we do not detect reduced resistivity in the upper mantle. Indeed, there is a slight increase in resistivity in the mantle beneath the outer rise, which is consistent with a feature in our previously published MT data (Naif et al., 2013). Since bending faults have been imaged to extend to mantle depths (Ranero et al., 2003), we would expect the triggering of serpentinization reactions to consume pore H$_2$O and reduce porosity by the ensuing volumetric expansion (Moody, 1976). This leads us to conclude that a lack of reduced resistivity in the outer rise mantle is not incompatible with serpentinization, and the small increase in resistivity may, perhaps, suggest it. However, since serpentinization weakens the mantle, it could lead to enhanced deformation. Additional deformation can expose fresh mantle surface area to promote serpentinization. Thus, with the onset of faulting at the outer rise, it is possible that a positive feedback is triggered to promote continuous serpentinization of the upper mantle (Macdonald and Fyfe, 1985). The resistive mantle in our model suggests this is unlikely, favoring episodic and spatially isolated mantle hydration not detectable by our data.

While our CSEM inversion model demonstrates that electrical resistivity can qualitatively map fluid pathways and fault damage zones, we must quantify porosity to constrain the volume of fluids in the crust. We apply Archie’s law, a robust empirical relationship (Archie, 1942), to estimate bulk porosity from our electrical resistivity model:

$$\phi = \left( \frac{\rho_w}{\rho} \right)^{1/m}$$ (4.1)

where $\rho$ is the bulk resistivity, $\rho_w$ is the pore fluid resistivity, $\phi$ is the porosity, and $m$ is the cementation exponent. Since the resistivity of saline fluid varies as a function of temperature (Quist and Marshall, 1968), we apply the cubic relationship of Constable et al. (2009) to estimate the pore fluid resistivity:

$$\rho_w(T) = 2.903916 \left( 1 + 2.97175 \times 10^{-2}T + 1.5551 \times 10^{-4}T^2 - 6.7 \times 10^{-7}T^3 \right)^{-1}$$ (4.2)
Figure 4.11: Synthetic model study of a layered Earth with no faults. Synthetic CSEM data generated from the layered model shown in panel a were inverted to RMS 1.0, yielding the smooth inversion model shown in panel b.
Figure 4.12: Synthetic model study of a layered Earth with faults. Synthetic CSEM data generated from the fault model shown in panel a were inverted to RMS 1.0, yielding the smooth inversion model shown in panel b.
where $T$ is temperature in degrees Celsius. We prescribe temperature using a plate cooling model for 24 Ma crust (Hasterok, 2013). However, anomalously low heat flux observed on the Cocos plate offshore of Nicaragua indicates vigorous hydrothermal circulation and cooler temperatures (Fisher et al., 2003; Harris et al., 2010). Conservatively prescribing temperature with a plate cooling model that does not account for thermal advection by hydrothermal circulation will underestimate pore fluid resistivity and as a consequence underestimate bulk porosity. Hence, we also use a colder geotherm suitable for well-ventilated crust with $\sim 10 \text{ mW/m}^2$ heat flux (Harris et al., 2010). In addition to temperature, bulk resistivity is sensitive to the degree to which pores are interconnected, which is incorporated into the cementation exponent in Archie’s law. We assign a cementation exponent of $m = 2$ based on constraints from laboratory and in-situ seafloor logging measurements of the Cocos crust (Becker et al., 1982; Brace and Orange, 1968; Becker, 1985).

In order to illustrate the impact of outer rise faulting on the magnitude of pore water in the crust, we split our resistivity model into five 20 km wide sections beginning at the trench and extending 100 km seaward. The section furthest from the trench (-100 to -80 km) represents normal oceanic lithosphere unaffected by faulting and each successive section features a rising intensity of faulting as expressed on the seafloor. For each section, we horizontally average the resistivity per unit depth below the seafloor and then estimate the porosity of the average with Archie’s law. The five bulk resistivity and porosity curves are plotted as a function of depth in Figure 4.13a and 4.13b, respectively. Figure 4.13 shows progressively less resistive and more porous crust with proximity to the trench, which correlates with fault density.

Oceanic crust typically contains three compositionally distinct layers of extrusive volcanics, sheeted diabase dikes, and plutonic gabbros that each exhibit contrasting porosities. We assume the Cocos crust contains 600 m of extrusives, 1 km of dikes, and 4 km of gabbros based on a combination of seismic and drilling constraints (Jarrard, 2003; Ivandic et al., 2008). To show the lateral changes in porosity within each crustal layer in greater detail, we use Archie’s law to calculate the porosity of our resistivity model without averaging. We again assume a cementation exponent of $m = 2$, but only
Figure 4.13: Broad-scale increase of crustal porosity with proximity to the trench. 

**a**, Lines show the averaged bulk resistivity as a function of depth of five 20 km wide sections from the trench to 100 km seaward. **b**, Porosity estimated from the averaged resistivity sections of panel **a** using Archie’s law with a cementation exponent of $m = 2$. Lower bound estimates are based on temperatures from the warmer plate cooling thermal model (Hasterok, 2013) and upper bound from the colder thermal model that considers hydrothermal circulation (Harris et al., 2010).

We use the more representative ventilated thermal model to assign pore fluid temperatures. Then, we vertically average the porosity for each layer in Figure 4.14b-d. Figure 4.14a shows the bathymetric slope angle along our transect, which we use to identify the location of seafloor faults according to their steepened slopes. Taken together, Figures 4.14 shows that faulted regions correlate with zones of enhanced porosity.

We find that the fault damage zones are typically less than 2 km wide in the extrusive volcanics and 2-6 km wide in the intrusive crust. The larger widths in the lower crust are likely due to decreasing resolution with greater depth. The resistivity model is generated with the Occam inversion scheme, which searches for the smoothest possible structure that fits the data to an RMS of 1.0 (de Groot-Hedlin and Constable,
Figure 4.14: Faults correlate with enhanced crustal porosity. a, The bathymetric slope of the seafloor along our survey transect clearly identifies the location of faults, which express suddenly steepened gradients. We highlight the location and width of fault zones at the seafloor as gray shaded regions, and shift them laterally to mark their respective location at depth assuming a 60°dipping fault. The vertically averaged porosity of the b, extrusive, c, dike, d, and gabbro crustal layers. The slight offsets between the seafloor fault location and the peak porosity at depth reflects the dip of the faults.
Therefore, our damage zone width estimates are upper bounds; the true *in-situ* widths of enhanced porosity may be sharper. Since our data are most sensitive to the conductivity-thickness product, reducing the width of the conductive fault channels imaged in Figure 4.10 requires an equivalent resistivity reduction. As a result, when keeping the cementation exponent constant, the porosity of a fault zone will be larger the more we reduce its width, but the laterally averaged bulk porosity estimates are unaffected. Hence, our laterally averaged porosity shown in Figure 4.13 and reported in Table 4.1 are more robust as they are less sensitive to our choice of regularization. Lateral averaging also dampens the effects of spatially varying permeability and thermal structure, which are strongest near faults that drive fluid circulation and advect heat.

In order to compare our porosity estimates with those of other studies, we take each crustal layer porosity shown by Figure 4.14 and calculate the average in 20 km intervals. The estimates are reported in Table 4.1. Bending faults effectively double the pore water content of the intrusive crust. The estimated porosities from the seaward end of our transect are in good agreement with the global average of 24 Ma crust (Jarrard, 2003), corroborating the significant enhancement we observe at the trench-outer rise. The extrusive crust is only slightly affected by faulting since the average porosity increases by a factor of 1.2 as the Cocos plate traverses the outer rise. Conversely, the porosity of the intrusive crust is considerably enhanced by faulting, increasing by a factor of 1.8 and 2.5 for the sheeted dikes and gabbros, respectively. Integrated over the full crustal thickness, we estimate that a 200 m thick column of water is subducted at the trench.

The low heat flux measurements demonstrate that the EPR-Cocos extrusive crust is both colder and more permeable than typical oceanic crust. Higher permeability warrants a lower cementation exponent for the upper crust. Matters are further complicated by surface charge conduction in clay minerals. Since clay is a prominent secondary mineral in the upper crust, ignoring its contribution to conduction yields overestimated porosity (Evans, 1994). We acknowledge these uncertainties and elect to simply use $m = 1.6$ as lower bound for the extrusive crust. The enhanced porosity and damage associated with faults will likely enhance permeability as well. It is difficult to assess the effect of faulting on the cementation exponent without *in-situ* borehole logs and
Table 4.1: Porosity of the incoming oceanic crust in fraction percent. Our most seaward crustal porosity average matches well with the independently estimated porosity of 24 Ma oceanic crust (Jarrard, 2003). The crust becomes progressively more porous with proximity to the trench, which correlates with increasing fault density.

<table>
<thead>
<tr>
<th>Segment distance from trench</th>
<th>Extrusives</th>
<th>Dikes</th>
<th>Gabbros</th>
</tr>
</thead>
<tbody>
<tr>
<td>24 Ma crust</td>
<td>m = 1.6, 2</td>
<td>10.4</td>
<td>3.0</td>
</tr>
<tr>
<td>$\rho \rightarrow \phi \rightarrow \text{mean}(\phi)$</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>80-100 km</td>
<td>7.5, 12.2</td>
<td>2.7</td>
<td>0.7</td>
</tr>
<tr>
<td>60-80 km</td>
<td>8.5, 13.5</td>
<td>3.7</td>
<td>0.8</td>
</tr>
<tr>
<td>40-60 km</td>
<td>8.5, 13.5</td>
<td>4.7</td>
<td>1.3</td>
</tr>
<tr>
<td>20-40 km</td>
<td>8.9, 14.1</td>
<td>4.8</td>
<td>1.3</td>
</tr>
<tr>
<td>5-20 km</td>
<td>9.1, 14.3</td>
<td>4.8</td>
<td>1.7</td>
</tr>
<tr>
<td>$\rho \rightarrow \text{mean}(\rho) \rightarrow \text{mean}(\phi)$</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>80-100 km</td>
<td>7.2, 11.9</td>
<td>2.6</td>
<td>0.7</td>
</tr>
<tr>
<td>60-80 km</td>
<td>8.1, 13.1</td>
<td>3.6</td>
<td>0.8</td>
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<tr>
<td>40-60 km</td>
<td>8.1, 13.1</td>
<td>4.5</td>
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<td>20-40 km</td>
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<tr>
<td>5-20 km</td>
<td>8.5, 13.6</td>
<td>4.5</td>
<td>1.5</td>
</tr>
</tbody>
</table>

rock samples. We prefer to interpret the porosity calculated with an exponent of $m = 2$ since it is unlikely that the fault damage results solely in enhanced permeability, the enhanced permeability is spatially isolated, and it is otherwise impossible to determine the value of the cementation exponent for the faulted outer rise crust without additional constraints.

For the values reported in the top half of Table 4.1, we elected to estimate porosity first and then proceed to calculate averages. The bottom half of Table 4.1 reports the porosity calculated by averaging the resistivity first and then using Archie’s law to estimate porosity. The two methodologies yield slightly different estimates that have minimal effects on our results.

The addition of water to unaltered basaltic crust will inevitably lead to alteration and hydrous mineral formation (Staudigel, 2014). We have shown that enhanced porosity is concentrated along the fault trajectories, which indicates the development of a heterogeneously hydrated crust. This has been demonstrated to impact the magnitude of water released by, as well as timing of, metamorphic dehydration reactions as the plate is subjected to continuously greater pressures and temperatures during subduction.
Furthermore, the average crustal porosity progressively grows until it peaks at the trench, which suggests that the hydration is continuous and lasts for the full extent of the outer rise.

The highly porous oceanic crust that we observe in the outer rise is consistent with geophysical and geochemical indications of a wet margin wedge and slab beneath Nicaragua (Abers et al., 2003; Patino et al., 2000; Worzewski et al., 2010). Fluidized bending faults – in conjunction with cold crustal temperatures, a highly permeable crustal aquifer, and hydrous alteration of the lower crust and upper mantle – likely have important implications for fluid cycling the state of the Nicaraguan seismogenic zone (Spinelli and Wang, 2008). Such conditions may also be optimal to harbor microbial colonies at the extreme limit of life in the deep crustal biosphere (Takai et al., 2014). A significant fraction of the total length of subduction zones develop bending faults at the outer rise (Masson, 1991). In light of our results, we conclude that the magnitude of water in subducting oceanic crust exceeds existing estimates and warrants consideration in the context of fluid cycling at subduction zones.

4.5 Water-rich megathrust of an erosive margin

Much of the Middle America Trench is a non-accreting margin where the entire layer of incoming sediment is underthrust with the subducting slab (Ranero et al., 2000). Seafloor drilling on the Guatemala forearc slope a few hundred kilometers northwest of our survey dictate that the upper plate is mafic and ultramafic igneous basement of oceanic origin (Aubouin et al., 1984). The structure off Guatemala has been interpreted to persist to the southwest off Nicaragua and northern Costa Rica (Vannucchi et al., 2001; Ranero et al., 2008).

In our electrical resistivity model (Figure 4.7), the forearc has a veneer of conductive sediments that covers the seafloor (approximately 500 m thick from the trench to 20 km landward and more than 1 km thick further landward). Underlying the seafloor sediments is a region up to two orders of magnitude more resistive, indicative of low porosities that are consistent with a non-accreting margin composed of crystalline rock or old accreted sediments that have since lithified. We observe the sediments on
the Cocos plate seafloor to be a highly conductive layer that is 400-500 m thick. At the trench, the conductive layer extends from the seafloor into the interior of the forearc margin.

Taking a closer look, Figure 4.15 shows a relatively thin underthrusting conductive channel that is coherent to ~25 km landward of the trench and has remarkable congruence to the geometry of the plate interface (from Hayes et al. (2012) Slab 1.0). We interpret the channel to be subducted sediments that maintain much of their electrical resistivity signature, and therefore some fraction of their initial pore fluid content. At 17-24 km, the conductor emanates from the plate interface and extends 2-3 km into the overriding plate. It terminates 1-2 km below a high concentration of active fluid seeps and mud diapers mapped on the forearc seafloor (Sahling et al., 2008). Further landward, the conductor rapidly dissipates to a resistive background.

We constructed a representative model with the same bathymetry and data composition as our inversion result to investigate the sensitivity and resolution of the data to the conductive channel and its particular geometry. The model, shown in Figure 4.16, includes a 600 m tall conductive channel at the plate interface and a resistive upper plate. We calculate forward CSEM responses for the synthetic model, add 2% random noise, and invert the resulting data with a 2% error floor. The synthetic inversion converged to RMS 1.0. The recovered model in Figure 4.16b shows that the CSEM data are sensitive to the presence of conductive sediments, but begin to lose resolution below 5 km depths, which is coincident with the plate interface at 25 km from the trench. Thus, the recovered model retains resolution to structure in the upper plate, confirming that the resistive signature and upward migration of the conductor are robust features in the data and not artifacts of inversion smoothing or modeling errors.

The particular location where the conductive channel transitions into the upper plate is quite intriguing. Looking back at Figure 4.7, we see that there exists a high density cluster of active seeps venting methane-enriched fluids on the seafloor (Sahling et al., 2008). The conductive channel deviates from the plate interface into the upper plate and peaks precisely beneath the region where seeps populate the seafloor. This suggests that the venting fluids originate from subducted sediments and migrate along steep seaward dipping pathways, which runs counter to previous observations of active
Figure 4.15: Close up view of forearc electrical structure. The channel of low resistivity is caused by subducted sediments that are congruent the predicted geometry of the plate interface. The 1992 tsunami earthquake ruptured this section of the megathrust.

seeps interpreted to vent fluids that migrate along shallow landward dipping splay fault pathways and thus originate from subducted sediments at greater depth and temperature (Hensen et al., 2004; Lauer and Saffer, 2012). Figure 4.17 is a schematic diagram that illustrates the two distinct pathways. Our observation does not negate splay fault activated seeps. Rather, it shows that the particular seeps here are venting locally-sourced fluids migrating along steeply dipping pathways. Ranero et al. (2008) show steep seaward dipping normal faults cutting through the forearc, surmised to be fluid pathways.

A resistive layer located 1-2.5 km below the seafloor separates the anomalous conductor from the surface sediments where the seeps reside. If the conductor is caused by fluids migrating along seaward dipping normal faults, we would expect it to extend
Figure 4.16: Conductive channel model study. Synthetic CSEM data generated from the model in panel a were inverted to RMS 1.0, yielding the smooth inversion model shown in panel b.
Figure 4.17: Forearc fluid migration pathways. Our CSEM resistivity model is consistent with (a) the seaward dipping pathway. Numerous seismic observations map splay faults below seafloor seeps that are consistent with (b) the landward dipping pathway.

from the plate interface all the way to the forearc seafloor. Furthermore, the conductor would require multiple closely spaced faults to account for its relatively wide spatial extent. If the conductor is smeared by the inversion smoothing, then the sharper it is the more conductive it needs to be to conserve the resistivity-thickness product.

Alternatively, the upper plate anomaly may be caused by hydraulic fracturing or a subducted seamount that damages the base of the overlying plate. A seismic reflection survey 100 km to the southeast imaged large negative polarity reflection amplitudes along the plate interface (Ranero et al., 2008). Negative polarity reflectors provide evidence for trapped fluids. Ranero et al. (2008) interpreted the MCS data as trapped water in the subducting slab generated by sediment dewatering that becomes highly overpressured and hydraulically fractures the base of the overriding plate, consistent with an erosional margin (von Huene et al., 2004a). The seafloor directly above the region of hydraulic fracturing would subside as a result.

The regional forearc bathymetry shown in Figure 4.18 helps to distinguish the cause of our upper plate anomaly. We loosely separate the forearc slope into lower, middle, and upper sections. Normally, we would expect the forearc to be most steep in the lower slope, gradually shallowing with increasing distance from the trench until becoming flat at the start of the forearc shelf. Instead, we see a coherent steepening in the middle slope along a wide swath of the Nicaragua forearc. von Huene et al. (2004b) made the same observation and attributed the steepening to either subsidence
by erosion or uplift by sediment underplating and preferred an erosional mechanism since it is more consistent with independent observations of long term subsidence offshore Guatemala (Vannucchi et al., 2004). We rule out subducting seamounts as the origin of the conductor since we would expect to find a spatially confined section of the seafloor that has been uplifted, which is not the case along our CSEM transect.

Let us consider a hydrofracturing origin for our upper plate conductor. In that case, the seafloor subsidence will peak at ~22-24 km from the trench to produce a steepened slope landward of the conductor. Alternatively, if we infer underplated sediments, the seafloor will be uplifted directly above the conductor, steepening the forearc seaward of ~22-24 km. We inspect the bathymetric gradient to test if either mechanism is consistent with the bathymetry. The seafloor gradient is estimated with a linear polynomial fit to 5 km wide overlapping windows of the SUESI-derived seafloor depth. The windows overlap 4.5 km so that the bathymetry slope is determined in 500 m increments in order to filter out small scale features such as fault scarps. Figure 4.19 shows the resulting seafloor angle as a function of distance from the trench. The seafloor steepens seaward of our upper plate conductor and begins sharply tapering off further landward. We conclude that the spatial extent of the conductor is consistent with sediment underplating. Furthermore, recent thermal modeling finds that the plate interface may be too cold to undergo extensive sediment dehydration in the vicinity of our anomalous conductor (Harris et al., 2010).

A large 3D seismic reflection survey performed during the early ’90s imaged landward dipping reflectors in the upper plate of the Nicoya forearc. The reflections were interpreted as evidence for underplated sediments that generate uplift and explain the extensional stresses observed in the region (McIntosh et al., 1993). Many workers have reinterpreted the MAT as an erosional non-accreting margin with a crystalline overriding plate since uplift is inconsistent with long term subsidence (Vannucchi et al., 2001). New results from an industry quality 3D active-source seismic survey reclassifies the MAT off Osa Peninsula (southeast of Nicoya) as an old severely deformed accretionary prism, not crystalline basement (Bangs et al., 2015). The 3D data tease out the internal structure of the margin, imaging fold-and-thrust belts throughout the forearc. Intriguingly, Figure 10 in Bangs et al. (2015) shows evidence for the emergence
Figure 4.18: A 3D view of the forearc slope offshore of Nicaragua. The steepened middle slope can be tracked for more than 100 km along strike. A subducted seamount on the right side of the map shows a unique bathymetric expression that is not seen in the immediate vicinity of our EM profile.
Figure 4.19: Forearc seafloor slope angle. The Nicaraguan forearc is drastically steeper than typical accreting margins. Between 15-25 km landward of the trench the seafloor slope peaks at 9°, which exceeds the average steepness of the lower slope (seaward of 15 km). The location of the anomalous seafloor relative to the conductive anomaly in our CSEM model favors an underplated sediment interpretation.
of a dome-like structure at the base of the upper plate, which loosely resembles our conductive anomaly. The authors briefly speculate on whether its origin is seamount subduction, concentrated faulting, or sediment underplating. The dome also marks a structural transition at the plate interface where high amplitude reflections cease to exist landward of the dome’s onset, which were inferred to reflect extensive fluid escape along steep seaward dipping pathways.

The seismic anomaly may very well be related to our conductive anomaly, but it is impossible to say without co-located seismic and EM data. Fortunately, the forearc segment of our profile is coincident with a past seismic survey. We perform a minimum gradient inversion of the forearc data to simplify the comparison with the seismic reflection. The minimum gradient inversion (MGS) mimics an $L_1$ minimization within an $L_2$ numerical framework (Portniaguine and Zhdanov, 1999). Figure 4.20 shows our MGS forearc resistivity model overlaid by the depth-migrated MCS section from profile NIC-50 (McIntosh et al., 2007). The geometry of the plate interface is steeper in the MCS data compared with the Slab1.0 model, so much so that the dip angles are unrealistic, exceeding 20° less than 25 km from the trench. The source of the discrepancy is likely the 1D velocity model used to depth migrate the two-way travel times. Reducing the upper plate velocity would shallow the plate interface and correct for the unrealistic dip. Sediment underplating justifies velocity reductions in the upper plate, lending further credence to our preferred interpretation.

The forearc porosity structure is also difficult to reconcile with hydrofracturing. As before, we estimate porosity with Archie’s law for cementation $m = 2$ and pore fluid resistivity from equation (4.2). However, plate cooling models are not applicable to subduction margins. Instead, we define the temperature in the forearc with the 2D thermal model of Harris et al. (2010). The model used forearc heat flux data collected along seismic transect BGR-99 line 41 offshore southern Nicaragua (magenta line in Figure 4.1) and considered the effect of increased advection in the porous/permeable subducting oceanic crust. Figure 4.21 shows the forearc porosity inferred from our resistivity model combined with the thermal model discussed above. The upper plate porosity is predominantly small, averaging less than 10%. The upper plate anomaly exceeds 15% porosity in a 10 km² region. Hydrofracturing would damage the rock and
allow fluid to escape, briefly sustaining excess porosity that collapse shortly after via compaction and hence subsidence. Larger permeability may persist to account for the enlarged conductivity, but this would have to correlate with seafloor subsidence. A more reasonable scenario is the addition of sediments to the base of the upper plate via underplating. It is a simple explanation for the higher porosity, while also being consistent with the bathymetry.

The physical circumstances that lead to underplating are not well understood. In a number of subduction margins outside the MAT, geologic and seismic observations identify anomalies interpreted as sediment underplating (Clift and Hartley, 2007; Collot et al., 2008; Kimura and Mukai, 1991; Platt et al., 1985; Park et al., 2002). Underplating where the subducting slab meets the crust-mantle boundary of the overriding plate can be attributed to sediments that are too buoyant to be transported any deeper (Bassett
Figure 4.21: Porosity cross-section estimated from the final converged resistivity model.

et al., 2010). Here, the proposed underplating is occurring at shallow depths. One possibility is that the edge of our conductor abuts with denser upper plate material. Walther et al. (2000) analyzed wide-angle seismic data from a transect approximate 50 km northwest of our CSEM survey. The velocity model superimposed on co-located seismic reflection data shows a boundary 30 km from the trench at the base of the upper plate that separates the seaward 4.5 km/s material from landward 5.8 km/s (see their Figures 3-4). In addition, they forward model gravity data with a coincident boundary separating 2.6 g/cm$^3$ seaward density from 2.9 g/cm$^3$ landward density (see their Figure 11). I propose that this boundary acts as a density filter and is responsible for the underplating inferred along our CSEM transect.

Regardless of the underlying cause, the proximity of our observed conductor to a high concentration of venting seeps suggests that some fraction of the escaping fluid is derived from subducted sediment sources. Fluids likely migrate vertically by diffuse flow or concentrated flow in permeable pathways (e.g., faults) and exit
the system via seeps on the seafloor, not necessarily by lateral fluid flow along the weakly dipping décollement to escape at the trench axis. The spatial extent of water escaping from the plate interface may be informed by the magnitude and density of venting fluids along the forearc slope. We cannot rule out deeply sourced contributions; dehydration of subducted crust and mantle at greater depths releases water that may migrate up dip along the décollement (Spinelli and Wang, 2009). If a pathway from the overlying continental crust penetrates the underthrust sediment barrier and taps into the subducting extrusive crust, deeper fluids could exit through the forearc slope (Tryon et al., 2010), possibly enhanced by underplated sediments that disrupt the décollement.

The incoming plate sediment cover contains ~150 m of hemipelagic mud (with 60 wt% smectite clay and 10 wt% opal) underlain by pelagic carbonate ooze (Spinelli and Underwood, 2004). Clays complicate electrical resistivity measurements since they produce a surface conductance that is non-linearly related to pore fluid salinity (Waxman and Smits, 1968). Thus, we estimate porosity along the conductive channel with an empirical relationship appropriate for shaly clay-rich sediments (Sen and Goode, 1992):

\[
\phi = \left( \frac{\rho_{\text{bulk}}^{-1}}{\rho_{w}^{-1} + R} \right)^{1/m},
\]

\[
R = 1.3\mu_{T}Q_{v} + \frac{1.93m\mu_{T}Q_{v}}{1 + 0.7\mu_{T}/\rho_{w}^{-1}}
\]

(4.3) (4.4)

where \(\mu_{T}\) is the mobility of counter-ions and \(Q_{v}\) is the clay counter-ion surface charge per unit volume. The mobility varies with temperature:

\[
\mu_{T} = 1 + 0.0414(T - 22)
\]

(4.5)

where \(T\) is temperature in degrees Celsius. We use two end-member parameter sets that are representative of clay-rich sediments to bound our estimated porosity (Conin et al., 2011). The sediment resistivity is defined to be the minimum value in the conductive channel as a function of distance from the trench. Figure 4.22 compares our estimated
Figure 4.22: Estimated sediment porosity compared with the independent porosity trend of hemipelagic sediment compaction. The dashed line denotes the region of our model that is not well constrained by the data at the deeper plate interface depths.

Porosity \( \phi \) with independent constraints. The predicted porosity trend is determined from laboratory based studies. Samples from regional sediment cores are used to measure the porosity to 3 km burial depths and a generic trend quantifies the porosity at greater depths (Spinelli et al., 2006). Overall, our estimates are in line with independent porosity trend.

During the initial stages of subduction, compacting sediments are expected to be the primary source of expelled fluids, followed by mineral dewatering at greater depths and temperatures (Pichon et al., 1993; Saffer and Tobin, 2011). In addition to the large influx of sediment water, we have shown that incoming oceanic crust is cooled by upper crust hydrothermal circulation and enriched with pore water along bending faults. The combined flux of sediment and crustal fluids in combination with cooler temperatures ultimately impact plate coupling by lowering the effective stress at the plate interface megathrust through pore pressure development. Lower temperatures
increase fluid viscosity, which reduce the magnitude of water needed to generate overpressures (Spinelli et al., 2006).

The onset of dewatering in smectite and biogenic opal sediment and the highly altered extrusive portion of the uppermost crust is a significant fluid source capable of generate extreme overpressures (Kameda et al., 2011; Saffer and Tobin, 2011). A colder margin will delay fluid release from dehydration reactions until further down dip and therefore alters the spatial variation of seismic coupling at the plate interface. The abundant availability of water at the plate interface suggests that the subducted sediments form a continuous albeit variably overpressured zone, lowering the effective normal stress along a large swath of the megathrust extending from the trench to ~20 km landward. At that point, sediment underplating changes the fluid structure of the megathrust further down dip and concurrently disrupts the décollement. In this scenario, the fluid structure that regulates effective normal stress and hence fault stability and plate coupling along the megathrust can be split into three distinct regions: an up dip zone dominated by sediment compaction, an intermediate zone where some fraction of the fluid-rich sediments are underplated, and a down dip zone where smectite-illite dehydration in the subducted sediment and crust is occurring.

The text of Chapter 4, in part, has been submitted for publication of the material as it may appear in Naif, S., Key, K., Constable, S., and Evans, R. L., Geochemistry Geophysics Geosystems, 2015. The dissertation author was the primary investigator and author of this paper. The text of Chapter 4, in part, is currently being prepared for submission for publication of the material. The dissertation author was the primary investigator and author of this material.
4.6 References


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