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SUBDUCTION ZONE HYDROGEOLOGY: QUANTIFYING FLUID SOURCES, PATHWAYS, AND PRESSURE

A Dissertation in
Geosciences

by
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ABSTRACT

Subduction zone fluids and fluid pressure distribution exert fundamental controls on the transport of heat and volatiles, and their associated distribution throughout the forearc. Additionally, excess fluid pressures have been proposed as a mechanism governing fault slip behavior, including their association with slow-slip events (SSE), episodic tremor and slip (ETS) and very-low frequency earthquakes throughout the conditionally stable region of the outer forearc. Fluid pressures develop as subducted sediments compact during tectonic burial, and through the production of dehydration derived fluids at depths associated with higher temperatures, and low porosity. Due to their production in a low porosity environment, the dehydration sources have an enhanced potential for pressurization, and the location of these reactions is therefore critical to understanding the mechanical strength of the plate boundary décollement, as it is mediated by the fluid pressure distribution. In an effort to further our understanding of fluid distribution in the forearc, I first develop a fluid budget using a numerical model parameterized by a combination of data obtained through ocean drilling, and laboratory testing of recovered sediments, which estimate the compaction behavior during subduction, and associated permeability reduction. Previous efforts to quantify the fluid budget did not include permeable splay faults, despite geochemical and geophysical evidence that they represent regions of focused flow, and provide a hydraulic connection from the source of fluids at the plate boundary, to the seafloor. The model results suggest that faults capture up to 35% of the total flux, and modeled flow rates are highly consistent with studies of both seafloor seeps, and flow at the trench, suggesting a quantitative link between the underlying permeability architecture of the forearc and flow rates at the seafloor. The results highlight the importance of these features in efficiently channeling heat and solutes from depth.
The second study examines the importance of faults in the determining the distribution of fluid pressures and deeply sourced fluids by coupling the results of dehydration modeling with a flow and transport model. The dehydration derived freshwater effectively acts as a tracer to consider the role of faults in distributing deeply sourced fluids, and translating fluid pressures away from the plate boundary. Model results indicate that faults are efficient translators of fluids, and capture deeply sourced fluids before they reach the trench. Overpressures develop at the base of the slope sediments, regardless of the permeability architecture employed, suggesting a potential mechanism for the formation of mud volcanoes, spatially correlated with faults that penetrate the forearc.

The results of a comprehensive heat-flow campaign offshore Costa Rica suggests plate boundary temperatures are much lower than previously thought, with dehydration reactions shifted farther down-dip, into regions of further depleted porosity. The third project in Costa Rica uses this new thermal structure to evaluate the spatial distribution of dehydration reactions along a 500-km segment of the plate boundary, from North of Nicoya, to north of Osa peninsula. I find that peak dehydration sources are spatially correlated with the rupture patch of all major earthquakes since 1950, and LFEs and slow-slip occur immediately down-dip of the reaction midpoint where the mole fraction of Smectite-Illite in the mixed layer clays is 50% Smectite and 50% Illite. I find that the transition from aseismic to seismic behavior correlates well with the temperature range associated with the precipitation of silica produced through smectite transformation, which stiffens the sediment through chemical compaction, and suggests a potential chemical mechanism for the onset of seismicity.
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Chapter 1

Introduction

1.1 Subduction zone hydrogeology

In subduction zones, fluid sources in the outer forearc are primarily associated with compaction of the underthrust sediments, as they are progressively exposed to increasing stress during tectonic burial. The water released through compaction is essentially seawater, with the largest sources occurring near the trench, where initially high porosity (60-80%) sediments experience rapid compaction and thinning, releasing ~90% of the pore fluid within the first 5-km of burial. In Costa Rica this burial depth is achieved ~20-km from the trench, where heatflow models previously predicted temperatures between 50 and 80°C, sufficiently high to drive clay dehydration reactions which provide another pulse of fluid that is added to the pore space as freshwater. Together these processes drive flow through a commensurate increase in fluid pressure, and the resulting fluid and pressure distribution, are both determined by the permeability structure of the forearc. Flow paths are governed by permeable pathways that focus flow from the subducted sediments to the seafloor, either along the plate boundary décollement to the trench, or via permeable faults that connect the plate boundary to the seafloor. These structures also govern the maintenance and distribution of fluid pressures, and therefore exert control on the mechanical behavior of the plate boundary. The role of the décollement zone in transporting heat and geochemical signals from deep within the subduction zone to the trench has been widely studied [e.g., Kastner et al., 1991; Bekins et al., 1995; Henry, 2000; Spinelli et al., 2006], but the importance of additional major faults in these processes has only recently been
acknowledged [e.g., Hensen et al., 2004; Sahling et al., 2008; Ranero et al.). Major splay faults are observed globally in convergent margins, and significant flow along these structures has been inferred from mapping of seafloor seeps [e.g., McAdoo et al., 1996; Suess et al., 1998], and local shoaling of the BSR [e.g., Bohrmann et al., 2002]. Geochemical analysis of fluid seeps identify a signature of dehydration reactions and elevated B and $^{18}$O values, which together suggest a fluid source region at much greater depths, with temperatures between 85 and 130°C (Hensen et al., 2004). Together these observations highlight the need to quantitatively establish the role of faults in:

1.) the fluid budget of the forearc, as regions of focused flow that provide a hydraulic connection from the site of deep reactions at the plate boundary, to the seafloor;

2.) translating pressure away from the plate boundary by providing a permeable pathway for fluids that tends to decrease fluid pressure, and increase effective stress and strength of the fault;

3.) determining the distribution of deeply sourced fluids throughout the forearc.

1.2 Chapter 1: The impact of splay faults on the fluid budget

I use a two-dimensional numerical model to evaluate the role of faults in the overall fluid budget, and the pattern of fluid discharge at the seafloor in non-accretionary subduction zones, using Costa Rica as an example. I include permeable splay faults, and décollement in the model, and parameterize the fluid sources based on experimental results from testing of recovered sediment inputs. I find that for a range of permeability from $10^{-13}$ to $10^{-16}$ m², faults capture from 6 to 35% of the total flux. Modeled flow rates are highly consistent with studies of both seafloor seeps, and flow at the trench, and I propose that the results establish a quantitative link between the underlying permeability architecture of the forearc, flow rates at the seafloor, and highlight the importance of these features in efficiently channeling heat and solutes from depth.
1.3 Chapter 2: Investigating the role of faults in distributing dehydration derived fluids, and translating fluid pressure

I use a 5th-order kinetic expression to model the transformation of smectite group clays to illite, and constrain the fluid source parameters from XRD analyses of sediment inputs. Using the results of the geochemical model, I combine the freshwater sources from smectite transformation with the seawater sources from compaction, using a mixing model that accounts for the effect of porosity loss in the sediments during progressive subduction. The combined source volume and chloride signature are incorporated into a two-dimensional flow and transport model, with a forearc geometry that emulates the Cost Rica margin. I use chloride as a tracer to evaluate the distribution of dehydration fluids, and determine a budget for their exit at the seafloor, using both a “warm” and “cold” thermal structure. The results suggest that faults readily transmit fluids, with up to 60% of dehydration fluid sources exiting the forearc along a single fault. In cold margins, the freshening impact is enhanced, owing the reactions initiating deeper, where pore space is further reduced. For the range of permeability considered, I find that faults efficiently translate pressures updip along the fault to the base of the slope sediments, where near-lithostatic, or lithostatic pressures develop. The results establish a potential mechanism for the formation of mud volcanoes and mound structures spatially correlated with splay faults, in response to unsustainable pressures that drive their formation, and enable persistent focused flow as observed on the upper slope.
1.4 Chapter 3: Along strike variability in clay dehydration sources: implications for fault slip behavior in Costa Rica

The new thermal model for Costa Rica (Harris et al., 2010ab) incorporates an unprecedented number of heat flow measurements and constraints on the shallow thermal structure along a 500-km section along strike. The model results suggest the temperatures at the plate boundary are significantly cooler than previously considered, and kinetically controlled reactions will therefore initiate and proceed at greater depths and distances from the trench. The new models, informed by unprecedented coverage of the margin, provide an ideal opportunity to re-evaluate clay dehydration and consider along strike variations in greater detail, along 16-transects that traverse the plate boundary. Here, I evaluate smectite transformation to illite along a 500-km section of the plate boundary, to assess the implications of the new thermal models in shifting the reaction further down-dip, and into the seismogenic zone. To consider the 2-dimensional distribution of fluid sources at the plate boundary, I use the 16-transects of thermal data corresponding to the plate boundary, and model the distribution of clay-dehydration sources along a 500-km section from north of Nicoya to north of Osa peninsula. I find that peak fluid sources from illitization are located with the rupture patch of all major earthquakes since 1950, and LFEs and slow-slip occur immediately down-dip of the reaction midpoint where the mole fraction of Smectite-Illite in the mixed layer clays is 50% Smectite and 50% Illite. I find that the transition from aseismic to seismic behavior correlates well with the temperature range associated with the precipitation of silica produced through smectite transformation, which stiffens the sediment through chemical compaction, and suggests a potential chemical mechanism for the onset of seismicity.
Chapter 2

Fluid Budgets of Subduction Zone Forearcs: The Contribution of Splay Faults

Abstract

Geochemical and geophysical evidence indicate that splay faults cutting subduction zone forearcs are a key hydraulic connection between the plate boundary at depth and the seafloor. Existing modeling studies have generally not included these structures, and therefore a quantitative understanding of their role in overall fluid budgets, the distribution of fluid egress at the seafloor, and advection of heat and solutes has been lacking. Here, we use a two-dimensional numerical model to address these questions at non-accretionary subduction zones, using the well-studied Costa Rican margin as an example. We find that for a range of splay fault permeabilities from $10^{-16}$ m$^2$ to $10^{-13}$ m$^2$, they capture between 6 - 35% of the total dewatering flux. Simulated flow rates of 0.1-17 cm/yr are highly consistent with those reported at seafloor seeps and along the décollement near the trench. Our results provide a quantitative link between permeability architecture, fluid budgets, and flow rates, and illustrate that these features play a fundamental role in forearc dewatering, and in efficiently channeling heat and solutes from depth.

2.1 Introduction

In subduction zones, permeable structural and stratigraphic pathways exert a primary control on fluid flow and associated chemical and heat transport (e.g., Moore and Vrolijk, 1992). These flow pathways also govern the maintenance and distribution of fluid pressure, and therefore impact the mechanical behavior of faults. The role of the décollement zone in transporting heat and geochemical signals from deep within the subduction zone to the trench has been widely studied (e.g., Kastner et al., 1991; Bekins et al, 1995; Henry, 2000; Spinelli et al.,
2006), but the importance of additional major faults in these processes has only recently been recognized (e.g., Hensen et al., 2004; Sahling et al., 2008; Ranero et al., 2008).

Major splay faults that cut the forearc are ubiquitous at convergent margins, yet a systematic, quantitative understanding of their impact on overall fluid budgets, rates and distribution of seafloor venting, and transport of heat and volatiles has been lacking. Significant focused fluid flow along such faults has been inferred from mapping of seafloor seeps (e.g., McAdoo et al., 1996; Suess et al., 1998), and local shoaling of the BSR (e.g., Bohrmann et al., 2002). Geochemical observations at seep sites, including low-Cl, elevated B, and $\delta^{18}$O values, imply that these faults provide a hydraulic connection from deep within the subduction zone to the seafloor, and therefore play an important role in the flux of volatiles from the subducting slab and sediments at or beneath the plate boundary décollement (e.g., Ranero et al., 2008; Hensen et al., 2004; Sahling et al., 2008). However, previous modeling studies of subduction zone hydrogeology have generally considered only the décollement zone as a permeable pathway (e.g., Bekins et al., 1995; Spinelli et al., 2006). Some studies have considered the effects of splay faults in local sediment dewatering (Shi et al., 1989) and in transporting heat (Cutillo et al., 2003), but have not systematically explored their role in forearc hydrogeology.

Here, we quantitatively evaluate the role of permeable splay faults on forearc fluid budgets and flow rates at non-accretionary subduction zones. We focus on a case study of the Costa Rican margin, where geochemical and thermal data and seafloor mapping show that these structures play a fundamental role in transporting fluids (e.g., Hensen et al., 2004; Sahling et al., 2008), and where the relevant rock and sediment physical properties, thermal regime, and geology are well-constrained by drilling and geophysical studies (Kimura et al., 1997; Ranero et al., 2008). We use a 2D model of fluid flow in a cross-section oriented perpendicular to the trench, in order to (1) quantify the budget and partitioning of fluid escape between permeable splay faults, diffuse flow through the overriding plate, and the décollement, (2) calculate flow
rates at the seafloor for comparison with values inferred from field measurements of geochemistry and heat flow, and (3) explore the sensitivity of these fluxes to fault and décollement permeability ($k_f$ and $k_d$) in order to provide a robust and quantitative framework for future interpretation of seafloor and subsurface data.

2.2 Geologic and hydrogeologic setting

The Middle America Trench (MAT) is formed by subduction of the Cocos Plate beneath the Caribbean Plate at ~8.5 km Myr$^{-1}$ (Figure 2.1A; DeMets, 2001). In our study area offshore the Nicoya Peninsula, the age of the subducting Cocos plate is 24 Ma. Anomalously low heat flow of 20-40 mW m$^{-2}$ has been documented in this region, and is attributed to vigorous low-temperature hydrothermal circulation in the ocean crust (e.g., Langseth and Silver, 1996). The sediments on the Cocos plate include ~250 m of pelagic carbonate ooze overlain by ~150 m of clay-rich diatomaceous mud (Kimura et al., 1997). The margin is non-accretionary; the incoming sediment section is completely subducted, and the overriding margin wedge is composed of the Nicoya Complex, an extension of the onshore basement (Kimura et al., 1997). The margin wedge is overlain by up to ~2 km of slope sediment, and a narrow (5-10 km) deforming prism at the toe of the margin wedge is formed by offscraping and reworking of these sediments (Figure 2.1).

The Costa Rican margin has been the focus of numerous geological, geophysical, and hydrologic investigations, including seismic reflection surveys, heat flow campaigns, ODP and IODP drilling, and mapping and sampling studies of vent sites on the continental slope (e.g., Shipley et al., 1992; Kimura et al., 1997; Hensen et al., 2004). Seafloor mapping, shallow heat flow, and geochemical data from seep sites all show that splay faults form an important flow pathway for fluids derived from subducting and dehydrating sediments at depth (e.g., Hensen et al., 2004).
Figure 2.1: (A and B) Map of the study area and model transect (dashed line). (C) Interpreted geologic structure from seismic reflection data, showing ODP drill sites and schematic locations of mud mounds (after Hensen et al., 2004 and Ranero et al., 2008).
Figure 2.2: Diagram of model domain showing subducting sediments (dark gray), margin wedge (light gray), frontal prism (stippled area), slope sediments, and splay faults and décollement (lines).

2.3 Modeling Methods

We use the two-dimensional finite element code SUTRA (Voss, 1984) to simulate steady state fluid flow along an 80-km long transect oriented perpendicular to the trench and approximately parallel to the plate convergence direction (e.g. Spinelli et al., 2006; Figures 2.1-2.2). These steady-state models represent the time-averaged fluid pressures and flow rates relevant to the first-order fluid budget of the forearc [e.g., Saffer & Bekins, 1998]. In our models, we prescribe fluid sources to emulate tectonic loading and compaction of the sediments on the incoming plate (e.g., Bekins & Dreiss, 1992). These fluid sources drive pore pressure and fluid flow in the model, mediated by the spatial distribution of permeability. Although our model does not explicitly couple pore fluid pressure, effective stress, and sediment compaction, the prescribed fluid source terms are consistent with overpressured conditions (e.g., Spinelli et al., 2006). Moreover, previous studies have shown that to first order, sediment dewatering and the resulting fluid pressures and flow patterns are not highly sensitive to the detailed distribution of fluid source terms when varied over a realistic range (Bekins & Dreiss, 1992; Saffer & Bekins, 1998).

Our model domain extends from 20-km seaward to 60-km landward of the trench, and includes the upper plate, décollement zone, and subducting sediments. In order to explore the role of splay faults on flow rates and the partitioning of fluid expulsion, we include seven faults that cut the upper plate margin wedge and connect the décollement to the seafloor (Figure 2.2). Based on constraints from seismic reflection data and drilling, we include a 2.5 km-thick drape of slope sediments overlying the margin wedge, and a 5 km-wide accretionary prism near the trench.
composed of reworked slope sediment (Figures 2.1-2.2). We define the model geometry and a fault spacing of 5-km based on seismic reflection data (Ranero et al., 2008; Shipley et al., 1992). The top and seaward boundaries of the model domain are prescribed as hydrostatic constant head. The bottom and landward boundaries of the model are set to a no-flow condition (Figure 2.2), under the assumptions that: (1) there is little hydrologic communication between the sediments and oceanic crust, as documented by inferred pore fluid pressure distribution within the subducting sediment section (e.g., Saffer, 2003); and (2) fluid sources and permeabilities > 60-km from the trench are sufficiently low that fluid flow there is negligible (e.g., Bekins et al., 1995). Previous modeling studies have shown that the inclusion of a permeable upper oceanic crust has little effect on fluid fluxes and pore pressures unless the sediments themselves are highly permeable (e.g., Spinelli et al., 2006), so relaxing the first assumption should not significantly impact our results. The primary model inputs are the fluid source terms and hydraulic conductivity structure; each of these is described in detail below.

2.3.1 Porosity and Fluid Sources

We calculate fluid source terms from porosity loss as sediments are transported into the subduction zone and buried (e.g., Bekins & Dreiss, 1992). We define porosity ($\phi$) as a function of depth ($z$, km below seafloor), constrained by data from both drilling and exhumed subduction zone sediments (Bray & Karig, 1985; Spinelli et al., 2006):

\[
\phi(z) = 0.7136 e^{-3.714 z} \quad 0 < z < 2 \text{ km} \tag{1}
\]

\[
\phi(z) = 0.4 e^{-2.1z} \quad z \geq 2 \text{ km} \tag{2}
\]
Porosities range from 0.7 at the trench to ~0.03 at 12 km depth. For the case of subducting sediments subjected to vertical (uniaxial) consolidation, the dewatering rate ($\Gamma$, units of $V_{\text{fluid}}/V_{\text{sed}} \text{s}^{-1}$) is then defined by:

$$\Gamma = \frac{\partial \phi}{\partial z} \tan(\alpha + \beta)v_p$$

(3)

where $v_p$ is the plate convergence rate ($\text{m} \text{s}^{-1}$), and $\alpha$ and $\beta$ are the surface slope and décollement dip angle, respectively. We use a total taper angle of 12-degrees ($\alpha=3.8^\circ$; $\beta=7.6^\circ$) (Shipley et al., 1992). In computing fluid sources, we do not include the contribution of fluids from mineral dehydration, because they are generally 10-100 times smaller than those derived from sediment compaction (e.g., Bekins et al., 1995; Spinelli et al., 2006). On the basis of porosity data at ODP drill sites and as described by equation (2), the total inventory of pore fluids entering the subduction zone is 21.7 m$^3$yr$^{-1}$ per m along-strike.

2.3.2 Permeability and Hydraulic Conductivity

We define the permeability of the incoming and subducted sediments using a permeability-porosity relationship defined by laboratory measurements on core samples from ODP Sites 1039, 1040, 1254, and 1255 (Saffer & McKiernan, 2005). We assign a permeability of $2 \times 10^{-17}$ m$^2$ to the slope sediments, which represents the effective vertical permeability for a 2.5 km-thick section, using a permeability-porosity relationship for similar slope sediments offshore Peru (Marsters & Christian, 1990) . In a subset of models, we do not include the slope sediments, in order to simulate a scenario in which the slope apron is breached by faulting or mud volcanoes (e.g., Hensen et al., 2004). Based on lithologic similarities identified through drilling (Kimura et al., 1997), we assign the narrow frontal prism the same permeability as the slope sediments.
To evaluate the impact of splay faults on the fluid budget and flow, we establish a baseline model with Nicoya Complex \( (k_n) \) and décollement \( (k_d) \) permeabilities of \( 10^{-19} \text{ m}^2 \) and \( 10^{-14} \text{ m}^2 \), respectively, and without including any other faults. In subsequent model runs, we include permeable splay faults, and consider a range of splay fault \( (k_f) \) and décollement permeabilities from \( 10^{-16} \) to \( 10^{-12} \text{ m}^2 \). These values are consistent with the range of subduction fault zone permeabilities measured by drill stem testing and inferred from flow rates, geochemical signals, estimated pore pressures, and numerical modeling studies (Saffer & Tobin, 2011 and references therein). A previous study of this margin (Spinelli et al., 2006) established that the fluid budget is insensitive to the permeability of the intact upper plate; therefore we assign the Nicoya Complex a permeability of \( 10^{-19} \text{ m}^2 \) for all simulations. In defining hydraulic conductivity, we account for the effects of temperature on fluid properties, using temperatures from the thermal model of Spinelli et al. (2006).

### 2.4 Results

#### 2.4.1 Forearc Fluid Budget

Our results show that splay faults play a major role in the fluid budget of convergent margins that has, until recently, been overlooked (e.g., Ranero et al., 2008). Their effects on the partitioning of fluid escape scale with their permeability and that of the décollement zone. In our baseline model (no permeable splay faults), 64% of the incoming fluids exit the forearc along the décollement, 36% via diffuse flow through the overriding plate, and <2% are subducted beyond the arcward model boundary with the downgoing plate. This is consistent with results of previous modeling studies at this and other subduction margins, which indicate that 60-70% of fluids exit along a permeable décollement and most of the remaining fluids escape by diffuse flow across the
seafloor of the forearc (Spinelli et al., 2006, Screaton et al., 1990, Saffer & Bekins, 1998). Although not the focus of our work, it is important to note that pore pressures in all of our simulations are modestly elevated, but do not exceed lithostatic values, consistent with both large-scale mechanical constraints and pore pressures inferred from drilling data (e.g., Saffer, 2003; Spinelli et al., 2006).

In models that include permeable splay faults, the percentage of fluids captured by the splay faults varies systematically with both $k_d$ and $k_f$ (Figure 2.3). For simulations with $k_d=10^{-14}$ m$^2$, the fraction of incoming fluid exiting along splay faults ranges from 6% to 30% as $k_f$ is increased from $10^{-16}$ m$^2$ to $10^{-13}$ m$^2$ (Figure 2.3). For simulations with lower décollement permeability ($k_d=10^{-15}$ m$^2$), the splay faults carry between 17% and 35% of the incoming fluid. The budget is generally not sensitive to further increases in splay fault permeability for values of $k_f > k_d$ (Figure 2.3), because for sufficiently high conduit permeabilities, the flux of fluids along the décollement and splay faults is limited by upward flow of fluids from the low-permeability subducting sediments (Figure 2.2A). We anticipate that for higher underthrusting sediment permeability, the same patterns would persist, but the faults could tap a larger proportion of the total fluids.
Figure 2.3: Partitioning of fluid escape between splay faults (dark shaded area) and along the décollement at the trench (light grey shaded area) as a function of fault permeability, for two different values of \( k_d \) as noted. The remaining fluids escape the forearc via diffuse flow.

In general, with increased splay fault permeability, the proportion of fluid exiting the forearc along the décollement and via diffuse flow both decrease, because the splay faults capture a larger fraction of fluid from the dewatering subducting sediments. This effect is also partly mediated by the décollement permeability; with higher \( k_d \), splay faults access and intercept more fluid from the décollement. This decreases flow rates along the décollement, but results in only a small impact on diffuse flow (Figure 2.3). In contrast, for lower values of \( k_d \), the splay faults access less fluid directly from the décollement, because flow to the splays along the décollement is the limiting factor (Figure 2.3, blue curves). In this case, the splay faults tap fluids that would otherwise exit the forearc via diffuse flow.
2.4.2 Fluid Flow Rates

Modeled seafloor flow rates demonstrate the efficiency of faults in channeling fluid, heat and solutes away from the plate boundary. Focused flow rates are up to 600 times higher than diffuse flow, which further documents their important role in the regional hydrogeology of the forearc (Figure 2.4). For cases with $k_d = 10^{-14} \text{m}^2$, simulated seafloor seepage rates along splay faults range from $0.08 \text{ cm yr}^{-1} \ (k_{\text{faults}}=10^{-16} \text{ m}^2)$ to $2.6 \text{ cm yr}^{-1} \ (k_{\text{faults}}=10^{-12} \text{ m}^2)$. For lower décollement permeability ($k_d = 10^{-15} \text{ m}^2$) the flow rates range from $~0.2$ to $4.4 \text{ cm yr}^{-1}$. In models that do not include the slope sediment, simulated peak flow rates are considerably higher, (up to $25 \text{ cm yr}^{-1}$), because flow at the seafloor remains focused along the structures rather than being diffused by the less permeable slope sediment. The simulated flow rates are in good agreement with rates of $0.3 - 20 \text{ cm yr}^{-1}$ inferred from geochemical profiles and heat flow measurements at seep sites (cf. Figure 2.1C; Hensen et al., 2004; Ranero et al., 2008); only the most active site of fluid expulsion exhibited a local flow rate significantly higher than our simulations ($300 \text{ cm yr}^{-1}$; Hensen et al., 2004). It is important to note that the flow rates reported at seep sites reflect significant focusing of fluid flow along-strike (i.e. these rates are not sustained at all locations along a given fault outcrop) (e.g., Ranero et al., 2008). As such, the reported rates should be considered maxima, whereas those we simulate in our 2-D model, which does not account for any out-of-plane effects, represent mean flow rates per unit length along-strike (Figure 2.4). Modeled flow rates within the décollement near the trench are also consistent with rates of $1.4 - 45 \text{ cm yr}^{-1}$ and $0.02-5 \text{ cm yr}^{-1}$ at Sites 1043/1255 estimated from drilling data (Saffer and Screaton, 2003), and measured by a borehole flowmeter (Solomon et al., 2009), respectively (Figure 2.4). Taken together, our results suggest that splay fault permeabilities are likely in the range $\sim 10^{-12}$ to $10^{-14} \text{ m}^2$. In all of our models, rates of diffuse flow at the seafloor are $<0.01 \text{ cm/yr}$ over the Nicoya Complex, and range from $0.02 – 0.2 \text{ cm/yr}$ over the frontal prism.
Figure 2.4: Simulated fluid flow rates at the seafloor for a range of scenarios (as labeled), and with a décollement permeability set to $k_d = 10^{-14}$ m$^2$. For comparison, we also show flow rates inferred from borehole observatory data (1; Solomon et al., 2009); drilling near the trench (2; Saffer & Screaton, 2003); and coring at seep sites, excluding only a local datum from the most active site (3; Hensen et al., 2004; Ranero et al., 2008).

Simulated flow rates along the splay faults are sufficiently high to advect solutes from the plate boundary at depth to the seafloor, and are up to ~600 times higher than in adjacent areas (Figure 2.4). Geochemical signals associated with deeply sourced reactions should therefore be effectively channelized, with strong and highly localized expression where splay faults intersect the seafloor, or near structures with an underlying hydraulic connection to the plate boundary (e.g. mud mounds, mud volcanoes). Our results are consistent with patterns of focused flow through the upper plate inferred from geochemical data (e.g., Teichert et al., 2005; Ranero et al., 2008), which document an increase in the proportion of chemically distinct deeply sourced fluids.
exiting the forearc with distance landward from the trench. This pattern is readily explained by splay faults that capture fluids originating at or below the plate interface.

2.5 Summary

The fate of fluids released by dewatering of subducted sediments has been assessed previously using numerical models, but generally assuming a simplified permeability architecture in which fluids reach the seafloor via diffuse flow through the overriding plate or along a highly permeable décollement. Recent observations of geochemical and thermal anomalies on the continental slope have underscored the importance of splay faults that cut the upper plate as key dewatering pathways that transport deeply sourced fluids to the seafloor across the forearc, especially at non-accretionary margins (e.g., Hensen et al., 2004; Ranero et al., 2008).

We use a numerical model of fluid flow at the Costa Rica convergent margin to systematically investigate the role of splay faults in forearc hydrogeology. Our results show quantitatively that splay faults exert a primary control on the partitioning of fluid expulsion. Simulated flow rates are consistent with flow rates determined from field measurements. The simulated flow rates are sufficiently high that these faults should efficiently advect heat and solutes through the forearc to sites of focused expulsion at the seafloor. One key implication is that these structures play a fundamental role in fluid, heat, and chemical transport through the forearc, with implications for non-accretionary convergent margins globally. Our results also provide a quantitative framework for interpreting observed flow rates and geochemical anomalies, by linking surface observations with the underlying physical hydrogeology.
2.6 Acknowledgements

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2.7 References


Chapter 3

Role of splay faults in chemical transport and pressure translation in subduction zones

Abstract

Observations at seep sites on the continental slope document that splay faults are primary hydraulic connections to the plate boundary at depth, carry deeply sourced geochemically distinct fluids to the seafloor, and are spatially associated with mud volcanoes. However, the role of these structures in forearc hydrogeology and solute transport remains poorly quantified. We use a 2-D numerical model that simulates fluid flow and chemical transport resulting from tectonically driven compaction and fluid release from clay dehydration to examine the physical and chemical hydrogeology of the forearc. We focus on the well-studied Costa Rica margin as an example, and consider the effect of thermal state on the distribution of fluid sources, and associated chloride concentration by simulating both a “cold” and a “hot” margin. Dehydration derived sources are calculated from a kinetic expression for smectite transformation and combined with compaction sources to determine the chlorinity distribution of fluids exiting the underthrust sediments. Our results suggest that faults readily transmit fluids to the seafloor and capture dehydration-derived fluids before they reach the trench via the décollement. In “cold” margins like Costa Rica, dehydration reactions initiate ~25-km from the trench, where porosity is reduced by up to 85%, and the subsequent impact on freshening is pronounced. In contrast, results for a “hot” margin suggest that fluids from dehydration will exit via the trench and frontal prism, and the effect on freshening is greatly diminished, owing to the high porosity of the sediments where dehydration occurs. We find that for a wide range of splay fault permeabilities, these structures effectively translate pore pressure updip, where faults intersect the base of low-permeability slope sediments, promoting the development of near-lithostatic fluid pressures. The results establish a plausible
mechanism for the formation of mud volcanoes and mound structures spatially correlated with splay faults, in response to unsustainable pressures that drive their formation, and enable persistent focused flow as observed on the upper slope.

3.1 Introduction

In subduction zones, porous sediments on the incoming plate are subjected to a progressive increase in tectonic and overburden stress that is initially borne by the fluid, creating a pressure gradient that drives flow, and leads to rapid dewatering and compaction of the subducted sediment (Shipley et al., 1990; Saffer et al., 2000). Compaction derived fluid sources are greatest near the trench, where high porosity sediments are initially subjected to tectonic loading (e.g., Bray & Karig, 1985; Moore & Vrolijk, 1992; Bekins & Dreiss, 1992; Saffer & Tobin, 2011). Excess pressures drive fluid flow that results in dewatering, as well as associated heat and solute transport throughout the forearc. Pore pressure also exerts a primary control on the mechanical behavior of faults and sediments by mediating effective stress (e.g., Byerlee, 1990; Moore and Vrolijk, 1992; Scholz, 1998).

Sediments follow a temperature and pressure trajectory that varies between subduction zones, and dehydration reactions tied to specific temperature windows shift accordingly (Kastner et al., 1991; Underwood, 2007). Metamorphic fluid release has been hypothesized to affect the type of slip on faults over a wide range of depths, for example by promoting episodic tremor (Peacock, 2009), both deep and shallow low-frequency earthquakes (Shelley, et al., 2006; Kitajima & Saffer, 2012), and silent-slip (Kodaira et al., 2004). Exhaustion of fluid production from clay dehydration has been suggested to delineate the up-dip limit of the seismogenic zone (e.g., Moore and Saffer, 2001; Spinelli and Saffer, 2004; Ranero et al., 2008), because the associated increase in effective stress would facilitate stick-slip behavior (e.g., Scholz, 1998).
Fluids sampled at seafloor seep sites and in boreholes at convergent margins are commonly geochemically distinct from seawater, and characterized by reaction products derived from deeper in the subduction zone (e.g., Kastner et al., 1991; Hensen et al., 2004; Sahling et al., 2008; Ranero et al., 2008) In many cases, these observations also document that faults cutting the upper plate play a key role in transporting fluids and solutes from the slab to the seafloor (Hensen et al., 2004, Tiechert et al., 2005; Ranero et al., 2008, Sahling et al., 2008; Reynes and Eberhart-Phillips, 2009) The primary source of dehydration derived fluid in the outer forearc is the transformation of smectite to illite (Bethke, 1986; Kastner and Elderfield, 1993; Bekins et al., 1994), which occurs over a temperature range of ~60-150° C and is strongly controlled by reaction kinetics. Consequently, the locus of smectite dehydration is intimately tied to the subducted sediment composition, the thermal structure of the subduction zone, and the plate convergence rate, all of which vary between margins and along-strike of individual subduction zones (Spinelli and Saffer, 2004; Underwood, 2007).

Previous modeling studies aimed at quantifying pore pressure and fluid flow patterns in subduction zones have generally assumed a simplified permeability structure that includes only the main plate boundary as a permeable conduit (Bekins et al., 1995; Saffer & Bekins, 1998; Matmon & Bekins, 2005; Spinelli et al., 2006). Splay faults are common features at non-accreting margins, and observations suggest that splay faults act as important hydraulic connections from the plate interface that breach the upper plate, although few modeling studies have considered their effects (e.g., Cutillo et al., 2006; Shi & Wang, 1988; Lauer & Saffer, 2012). As a result, the role of these structures in overall margin hydrogeology, including channeling deeply sourced fluids and translating fluid pressures, remains largely unquantified. Here, we address this problem using a 2-D numerical model of coupled fluid flow and transport, using Costa Rica as an example, along a 70-km transect oriented perpendicular to the Middle America trench, offshore the Nicoya peninsula (Figure 3.1). Specifically, we (1) investigate the role of splay faults in
controlling the patterns of fluid egress and fluid geochemistry at the seafloor; (2) assess the relationship between the thermal structure of the forearc and the impact of dehydration reactions on fluid geochemical transport patterns; and (3) discuss the implications of these findings in terms of flow and pressure translation.

3.2 Geologic Setting of the Nicoya Margin

3.2.1 Tectonic and Thermal Setting

The Middle America trench is formed by northeastward subduction of the Cocos plate beneath the Carribean plate at a rate of ~85 mm/yr (DeMets, 2001; Figure 3.1A). Offshore of the Nicoya Peninsula, the subducted crust is uniform in age (~20-25Ma), but is divided between crust formed at the Cocos-Nazca Spreading center (CNS) to the southeast, and crust formed at the East Pacific Rise (EPR) to the northwest (Barkhausen et al., 2001). This region experiences regular seismicity, with approximately 30 earthquakes ≥ M 6.0 documented in the past 40-years, in the 250-km region surrounding the Nicoya peninsula. The seismogenic zone for this margin lies beneath the Nicoya peninsula, making it an ideal site for multiple GPS campaigns and ocean-bottom seismometer deployments designed to investigate the seismic portion of the forearc, and identify regions associated with slow-slip, episodic tremor, and microseismicity (Newman et al., 2002; Norabuena et al., 2004; Ghosh et al., 2008; Outerbridge et al., 2010). Geodetic studies of the Nicoya region have identified locked portions of the plate boundary, 55-km and 80-km from the trench (11–18 km depth), by amounts that exceed 50% of plate rate (Norabuena et al., 2004), and a slow-slip region down-dip of the seismogenic zone (~30-km depth) that is correlated with observations of seismic tremor (Outerbridge et al., 2010)
CNS derived crust exhibits surface heat flow of 105-115 mW/m², consistent with lithospheric cooling models. In contrast, the EPR crust is characterized by surface heat flow of only 20-40 mW/m² (Langseth and Silver, 1996; Fisher et al., 2003; Harris et al., 2010). The sharp thermal transition and anomalously low heat flow in EPR crust is attributed to vigorous hydrothermal circulation through basaltic basement outcrops that provide a hydraulic connection between the igneous oceanic crust and the seafloor, resulting in efficient mining of heat by large-scale lateral flow between outcrops (Fisher et al., 2003; Hutnak et al., 2007; Harris et al., 2010).

The thermal structure of the incoming plate has been well constrained by >300 heat flow measurements conducted on both CNS and EPR derived crust (Fisher et al., 2003; Hutnak et al., 2006; Harris et al., 2010a; Harris et al., 2010b), and these data have been used to constrain thermal models of the subduction zone (e.g., Harris & Wang, 2002; Spinelli and Saffer, 2004). These thermal modeling studies show that the thermal variability observed in the incoming plate impacts the thermal structure of the entire subduction zone, and may partly control pore pressure distribution and fluid flow patterns by affecting both the spatial distribution of diagenetic fluid sources, and fluid viscosity (Spinelli and Saffer, 2004).

The sediment on the incoming Cocos plate is approximately 400-m thick, and consists of ~250-m of carbonate pelagic ooze underling ~150 m of diatomaceous clay-rich mud (Shipboard Scientific Party, 1997a; Spinelli and Underwood, 2004). On average, the diatomaceous section contains ~60 wt% smectite, with no significant variation along strike (Spinelli & Underwood, 2004). The combined fluid source from compaction and dehydration is ~23.8 m³/yr per m along strike, with 21.7 m³/yr⁻¹ derived from compaction, and 2.1 m³/yr⁻¹ through smectite transformation to illite.

The Costa Rican margin is non-accretionary, meaning that all of the incoming sediment is subducted at the trench (e.g., Shipley et al., 1992; Shipboard Scientific Party, 1997b). Apart from a small (~5-km wide) frontal prism composed of reworked slope sediments, the overriding
margin wedge consists primarily of basement rock, thought to be an extension of the onshore Nicoya Complex (Kimura et al., Shipboard Scientific Party, 1997b; Vannuchi et al., 2001). The margin wedge is draped by ~2-km of slope sediments, which are compositionally similar to the frontal prism (Kimura, et al., 1997). Seismic reflection studies (Shipley et al., 1990; Shipley et al., 1992) and drilling (DSDP Leg 84; ODP Legs 170 and 205) provide important constraints on the forearc geometry, and illustrate the initial phases of sediment dewatering.

### 3.2.2 Forearc Hydrogeology

The Costa Rica margin has been the focus of numerous studies investigating the hydrogeology of the forearc, through a combination of ocean drilling, direct sampling of pore fluids in areas of focused flow (Hensen et al., 2004), and numerical modeling studies designed to quantify patterns of fluid flow, and the forearc fluid budget (Silver et al., 2000; Screaton and Saffer, 2005; Lauer and Saffer, 2012). Together these studies afford a basic understanding of the outer forearc hydrogeology, where flow is primarily attributed to rapid sediment compaction and thinning. However, sampled pore fluids at the toe of the prism indicate a fluid component with source temperatures of 60-160°C (Chan and Kastner, 2000), indicating transport from ~20-25-km down-dip, and implying that the plate boundary is sufficiently permeable to transmit fluids from the site of these reactions to the trench. Low chloride anomalies have also been observed within the décollement at IODP Sites 1040, 1043, and 1254, and in the overlying wedge at Sites 1043, and 1040 (Shipboard Scientific Party, 1997a; Silver et al., 2000; Chan and Kastner, 2000; Shipboard Scientific Party, 2003; Spinelli et al., 2006) Chloride depletion in the overlying wedge is attributed to focused flow along faults or a permeable structure that occurs at ~ 75-mbsf in Site 1043, and ~200-mbsf in Site 1040.
In addition, low chloride fluids that are enriched in Boron and thermogenic hydrocarbons have been documented at fault scarps and mound structures on the upper slope (Hensen et al., 2004). These data, combined with isotopic analyses of the fluids, suggest that they originate at temperatures of 85-150º C. Thus, splay faults act as important conduits for fluid flow and the transport of deeply sourced reaction products from the plate boundary to the seafloor (Hensen et al., 2004; Sahling, et al., 2008). Here, we build on previous studies by considering the role of faults in transporting solutes, and in translating fluids and fluid pressures away from their source at the plate boundary.

3.3 Methods

We use the finite element code SUTRA (Voss, 1984) to simulate steady state fluid flow and solute transport along a 70-km transect oriented perpendicular to the trench (e.g. Spinelli et al., 2006; Lauer and Saffer, 2012; Figures 3.1, 3.2). A steady state simulation was chosen to represent the time-averaged, long-term flow rates and pathways, and the fluid pressure distribution that develops in response to subduction. In our models, fluid pressures and flow are driven by fluid sources driven by two processes: (1) tectonic loading and subsequent compaction (porosity loss) of the sediments on the incoming plate, and (2) smectite dehydration (e.g., Bekins & Dreiss, 1992; Bekins et al., 1995; Spinelli et al., 2006). The dehydration-derived component of the fluid is fresh, which enables us to track the fate of these fluids (e.g., Bekins et al., 1995).

Steady state fluid flow in 2-dimensions is described by:

\[
\nabla \left[ \frac{k \rho_w}{\mu} \right] \cdot (\nabla P_f - \rho_w g) - \rho \Gamma = 0
\]

where \(k\) is permeability [m\(^2\)], \(\rho_w\) is fluid density [kg m\(^{-3}\)], \(\mu\) is dynamic fluid viscosity(Pa s), \(P_f\) is fluid pressure [Pa], and \(\Gamma\) is a fluid source term that includes sources from...
both compaction and clay dehydration. The dehydration-derived sources introduce fresh water to
the system; we use Chloride (Cl\(^-\)) as a conservative tracer to track the fate of these fluids as they
are transported from the underthrust sediments to their egress at the seafloor (e.g., Kastner &
Elderfield, 1993; Bekins et al., 1995). The equation governing steady state 2D non-reactive solute
transport is:

\[ \nabla \cdot (v \rho_w c) - \nabla \cdot (D \nabla \rho_w c) = 0 \tag{2} \]

where the first term describes the advective component of transport: \(v\) is the fluid velocity derived
from Darcy’s Law and \(c\) is the molal concentration, and the second term describes the diffusive
component of transport, where \(D\) is the coefficient of hydrodynamic dispersion.

Our model domain extends from 10-km seaward to 60-km landward of the trench, and
includes the upper plate and subducting sediments (e.g., Spinelli et al., 2006). Based on
constraints from seismic reflection data and drilling, we include a 2.5 km-thick drape of slope
sediments overlying the margin wedge, and a small (5 km-wide) accretionary prism near the
trench composed of reworked slope sediment (Kimura et al., 1997; Silver et al., 2000; Figures
3.1B, 3.2). To explore the role of splay faults in partitioning of deeply sourced fluids and
translating fluid pressures, we include seven faults that connect the décollement to the seafloor
and cut across the upper plate (Lauer & Saffer, 2012) (Figure 3.2). We define the model geometry
using a taper angle of 11.4° (décollement dip, \(\beta=7.6°\); surface slope, \(\alpha=3.8°\)), and a fault spacing
of 5-km, both based on seismic reflection studies of the margin (Ranero et al., 2008; Shipley et
al., 1992; Christeson et al., 1999).

The top and seaward boundaries of the model are assigned a hydrostatic constant head,
and seawater concentration, or 550 mM Cl\(^-\). The bottom and landward boundaries of the model
are set to a no-flow condition (Figure 3.2), assuming: (1) there is minimal hydraulic
communication between the sediments and oceanic crust, as suggested by the pore fluid pressure
distribution within the subducting sediment section (e.g., Saffer, 2003); and (2) the fact that fluid sources and permeabilities > 60-km from the trench are low enough to consider fluid flow as negligible (e.g., Bekins et al., 1995). Additionally, previous studies have shown that including permeable upper oceanic crust in flow models has a small effect on simulated fluid fluxes and pore pressures except when the sediments themselves are highly permeable (e.g., Saffer and Bekins, 1998; Matmon and Bekins, 2005; Spinelli et al., 2006), so relaxing the first assumption above would not appreciably impact our results.

Figure 3.1A/B/C: Map of the study area and model transect (dashed line). (C) Interpreted geologic structure from seismic reflection data, showing ODP drill sites and schematic locations of mud mounds (after Hensen et al., 2004 and Ranero et al., 2008).
Figure 3.2: Diagram of model domain delineating subducting sediments (dark gray), margin wedge (light gray), frontal prism (stippled area), slope sediments, and splay faults and décollement (lines).

Figure 3.3: Porosity distribution used in numerical model, based on shallow IODP drillholes and exhumed subduction complex porosity (Bray and Karig, 1995; Stewart and Peselnick, 1977).
3.3.1 Porosity and compaction fluid sources

We compute the compaction derived fluid sources from assumed porosity loss as sediments are transported into the subduction zone and progressively buried (e.g., Bekins & Dreiss, 1992; Screaton et al., 1990). In prescribing porosity reduction and associated fluid sources, we pin our reference frame to the trench, and move sediments through a steady-state porosity field that accounts for compaction and porosity loss arcward of the deformation front, as the sediments consolidate through progressive burial. The volume of water lost from a parcel of sediment moving through the porosity field is equivalent to the amount of pore volume lost through compaction, as the parcel follows a path relative to the deformation front. The rate of fluid production is therefore tied to the sediment velocity, which is in turn determined by the plate velocity.

We describe porosity ($\phi$) using an exponential porosity-depth relationship that honors data from drilling at the Costa Rican margin (IODP Leg(s) 170, 205; Sites 1039, 1040, 1043; Figure 3.1B), and from exhumed subduction zone sediments (e.g., Bray & Karig, 1985) (Figure 3.3):

$$\phi(z) = 0.7136 e^{-37.14z} \quad 0 < z < 2 \text{ km}$$

$$\phi(z) = 0.4 e^{-21z} \quad z \geq 2 \text{ km}$$

where $z$ is the depth below seafloor in meters. Porosity values range from ~0.7 at the trench, to 0.03 at a burial depth of ~12.5-km (Figure 3.3).

Because the Costa Rican margin is non-accretionary, the compaction-driven fluid source terms arise only in subducted sediments, which are subjected to primarily vertical consolidation. In this case, the computation of fluid sources ($\Gamma$, units of $V_{\text{fluid}}/V_{\text{sed}} \text{ s}^{-1}$) is greatly simplified in comparison to the case of accreted sediments. The sources are defined by:
\[
\Gamma = \frac{\partial \phi}{\partial z} \tan(\alpha + \beta) \cdot v_p \ [s^{-1}]
\]  \hspace{1cm} (5)

where \( v_p \) is the plate convergence rate (\( m \ s^{-1} \)), \( \phi \) is porosity (unitless), \( z \) is the depth below the seafloor (\( m \)), and \( \alpha \) and \( \beta \) are the surface slope and décollement dip, respectively.

### 3.3.2 Clay Dehydration Fluid Sources and Fluid Source Composition

The largest fluid source from mineral dehydration results from the transformation of smectite-group clay minerals to illite (Moore and Vrolijk, 1992), a reaction that occurs in a series of individual steps that are described well by a single “lumped” kinetic expression (e.g., Pytte and Reynolds, 1988; Huang et al., 1993). The reaction rate is determined primarily by the exposure time and temperature of the sediments, which are in turn controlled by the plate convergence rate and thermal structure. In Costa Rica, the thermal state of the incoming plate varies, but the sediments are thin, and as a result, temperatures are not high enough to initiate reaction progress until \( \sim 20 \)-km landward of the trench. In our simulations, we use the Pytte and Reynolds (1988) kinetic expression, because it provides the best fit to the available data that cover the large range of temporal and thermal conditions experienced by subducted sediments, and matches observed reaction progress at other subduction zones well (Bekins et al., 1994; Saffer et al., 2008). The reaction rate is given by:

\[
\frac{\partial S}{\partial t} = -A e^{-E/RT} \left( \frac{[K^+] \text{ } [Na^+]}{S^n} \right) S^n
\]  \hspace{1cm} (6)

where \( S \) is the mole fraction of smectite; \( A \) is a scaling factor of \( 5.2 \times 10^{-7} \) \( s^{-1} \); \( E \) is the reaction activation energy (\( 1.38 \times 10^{-5} \) J/mol); \( R \) is the ideal gas constant; \( T \) is temperature (Kelvin); \( n \) represents the reaction order with respect to smectite (\( n=5 \)). The prescribed temperature field was based on the results of thermal modeling for Costa Rica (Spinelli et al.,...
For our models, we consider an end-member scenario, in which all of the interlayer clay is unreacted smectite, or $S_{\text{initial}}=1.0$. Analysis of sediment inputs indicate a mole fraction of ~90% smectite in the mixed layer clays (Spinelli and Underwood, 2004), slightly less than our end member scenario. In applying equation (7), we assume that the ratio of $K^+$ to $Na^+$ is in equilibrium with K-feldspar, and varies with temperature (Pytte and Reynolds, 1988):

$$\left(\frac{[K^+]}{[Na^+]}\right) = 74.2 \ e^{-\frac{2400}{T}}$$

Fluid production associated with illitization is given by:

$$\Gamma_{\text{illitization}} = \frac{\partial S}{\partial t} \cdot H \cdot C \cdot (1 - \phi)$$

where $\Gamma$ is the source term ($V_{\text{fluid}}/V_{\text{sediment}}$/s), $H$ is the volumetric water content of smectite, and $C$ is the volume fraction of smectite in the bulk sediment. We assume there are two layers of bound water, corresponding to a volumetric water content ($H$) of 0.4 (Colten Bradley, 1987; Bird, 1984), and the initial smectite content of the bulk material is 60 wt%, smectite, or $C=0.6$ (Spinelli and Underwood, 2004).

Implicit in equation (7) is the assumption that $K^+$ is available, and the reaction is therefore not limited. If we were to consider a case in which $K^+$ is limited (Perry and Hower, 1970), our simulated fluid production rates would be lower, and the reaction would extend to greater depths. In our calculation of the source terms, we also assume that: 1.) any porosity increase caused by a decrease in the volume of solid phases during illitization in the unlithified sediments is short-lived, and the pore space collapses to recover the previous value (e.g., Bangs et al., 1990; Bekins et al., 1995); and 2.) a 5% increase in fluid volume occurs through expulsion of the interlayer water (Bethke 1986; Bekins et al., 1995). Taking both sets of assumptions into account, the dehydration fluid source terms we report should be considered maxima.
We consider two scenarios for the margin thermal structure, to explore the range of dehydration-driven fluid sources and solute transport behaviors associated with hot and cold end members for the margin thermal state. We define these two cases based on recent heat flow studies and thermal modeling efforts along the Middle America Trench (Harris et al., 2010 a-b). For the cold model, we incorporate a thermal structure consistent with low heat flow on the incoming plate (Q =30 mW/m²; equivalent to a geothermal gradient on the incoming plate of 12.5°C/km). For our hot model, we use a thermal structure consistent with an incoming plate heat flow of 120 mW/m² (50° C/km). To compute the chemistry (chlorinity) of fluid sources in our model domain, we track the composition of pore fluids as they undergo consolidation and illitization. We accomplish this using a mixing model that accounts for porosity loss and progressive freshening from dehydration-derived sources (Figure 3.4):

\[ \text{Cl}^-_{i+1} = \frac{\text{Cl}^-_i \cdot \phi_i}{\phi_i + \Gamma_{\text{illitization}} \cdot v_{\text{plate}}} \]  

where \( \phi \) is given by equations (3) and (4), and the initial chloride concentration is set to a seawater value ([Cl] =550 mM). In equation 9, the subscripts refer to the current timestep (ii) and the next time step in the calculation (ii+1).
3.3.3 Permeability and Fluid Properties

Porosity loss through compaction leads to a corresponding decrease in permeability, with the highest rates of reduction occurring within the first 5-km of burial (Shipley et al., 1990; Saffer et al., 2000; Saito and Goldberg, 2001). We define the permeability of the incoming and subducted sediments using permeability-porosity relationships derived from laboratory measurements over a wide range of effective stresses on core samples from ODP Sites 1039, 1040, 1254, and 1255 (Saffer and McKiernan, 2005), representing both pelagic and hemipelagic sediment inputs (Figure 3.5). We assign a permeability of $2 \times 10^{-17} \text{ m}^2$ to the slope sediments draping the upper plate, based on laboratory-derived permeability-porosity relationships for slope sediments offshore Peru (Marsters and Christian, 1990). We obtain this value by calculating an effective vertical permeability for a 2.5-km section, using the porosity depth distribution attained...
through drilling. Based on lithologic similarities identified during drilling (Kimura, et al., 1997), we assign the small frontal prism the same permeability as the slope sediments.

To evaluate the role of faults in transporting solutes, and in translating fluids and fluid pressures away from their source at the plate boundary, we first establish a baseline model with a permeable décollement \(10^{-14} \text{m}^2\), and Nicoya complex \(10^{-19} \text{m}^2\), without including faults other than the décollement. We then conducted a suite of simulations to quantify the effects of splay fault permeability. In these models, we consider a range of fault permeability from \(10^{-16}\) to \(10^{-12}\) \(\text{m}^2\), and a décollement permeability of \(10^{-14}\) and \(10^{-15}\) \(\text{m}^2\), a range that is consistent with those identified for these structures through comprehensive sensitivity studies that are calibrated by numerous observations (Lauer and Saffer 2012, Spinelli et al., 2006)

We define the temperature distribution in the forearc from the thermal models of Spinelli et al. (2006), and account for the effect of temperature on fluid viscosity \(\mu\) by (Voss, 1984):

\[
\nu = 2.4 \times 10^{-5} \left[ 10^{24.837/\left(T^{1.163} + 0.155\right)} \right],
\]

(10)

where temperature \(T\) is in degrees Celsius. The total inventory of pore fluids entering the subduction zone is 23.8 \(\text{m}^3\text{yr}^{-1}\) per meter along-strike; consisting of 21.7 \(\text{m}^3\text{yr}^{-1}\) from compaction, and 2.1 \(\text{m}^3\text{yr}^{-1}\) derived through smectite transformation to illite.
Figure 3.5: Experimentally defined porosity-permeability used to parameterize the model.

3.4 Results and Discussion

3.4.1 Fluid source concentration

In a “hot” margin, reactions occur much closer to the trench, where porosity has not been significantly reduced, and fresh fluid is added to a larger volume of salt water entrained in the sediment. The subsequent impact on dehydration derived freshening is significantly diminished (Figure 3.6), and modeled chloride concentrations in fluid sources never decrease below 300 mM,
or 55% that of seawater (550mM). In contrast, results from a cold margin suggest a much greater freshening potential, owing to the reduced pore space at the depth of peak freshening. In this case, chloride concentration in fluid sources are reduced to ~50 mM (9% that of seawater) within 50-km of the trench (Figure 3.6C).

The mole fraction of smectite remains constant until 20-km landward of the trench, where temperatures reach ~60°C and the reaction initiates. The reaction progresses rapidly between ~30 and 50-km landward of the trench as Illite increases from 2 to 84% of mixed-layer clays (Figure 3.6A). The reaction rate slows further down dip as the reactant is depleted, but the impact of the freshwater source on chloride concentration is pronounced due to the greatly reduced pore volume (Figure 3.6C). Dehydration sources peak about 45-km from the trench, which corresponds to a temperature of ~100°C for this margin (Spinelli et al., 2006), and a porosity of 0.06, which reflects a >90% reduction in the original pore volume. The impact of clay dehydration on freshening is intimately tied to the porosity evolution, which determines the amount of fluid present before mixing. In a warmer margin, dehydration will initiate much closer to the trench, where porosities are still relatively high, and the freshening signature will be significantly reduced.
Figure 3.6 (A/B/C): Results of geochemical modeling illustrate the downdip progression of smectite transformation, results of mixing model combining compaction and dehydration derived sources, and freshwater fraction of the total source. The solid line represents the warm model and the dashed line is the cool model.
3.4.2 Fluid Budget and spatial distribution of deeply sourced fluids

The thermal structure of the forearc has a significant effect on the patterns of freshening and simulated fluid geochemistry, both in the subsurface and at points of fluid egress. In a cold margin like Costa Rica, the pattern of freshening is controlled by faults that efficiently transport deeply sourced fluids away from their source at the plate boundary, to the seafloor, before they reach the trench (Figure 3.7). To further the impact of the forearc thermal structure, we calculate the total fluid budget, and use the chloride concentration of fluids exiting the forearc at the seafloor to map the distribution/budget of fluids derived from illitization (3.8). In the cold simulation, 38% of the total fluid source exits along the décollement, and faults transmit 33% of the fluid, with the remaining 29% transmitted to the seafloor diffusely through the prism (25%) and Nicoya complex (4%) (Figure 3.8B). Deeply sourced fluids from illitization exit the forearc primarily along faults (64%), with the remaining 36% exiting the forearc diffusely through the Nicoya complex (Figure 3.8C). For the hot simulation, 50% of the total fluid source exits along the décollement, while faults transmit 21% of the fluid, and the remaining 29% exits diffusely through the prism (25%) and Nicoya complex (4%) (Figure 3.8B). Sources from illitization shift closer to the trench in the hot scenario, with 52.5% exiting the forearc via the décollement, and only 18.5% transmitted along faults. The remaining deeply sourced fluid exits diffusely through the prism (26%) and Nicoya complex (4%) (Figure 3.8C).

Calculated chloride concentration from seafloor fluxes further highlight the difference between thermal structures, and the freshening potential of reactions occurring deep in the forearc (Figure 3.8D). In a cold margin, simulated seafloor fluxes are significantly depleted in chloride, with chloride values at individual faults ~10% of seawater, or 60 mM [Cl\(^-\)], while results from a warm margin have a much smaller impact on the chloride concentration of exiting fluids, and chloride calculations never get below 300 mM (Figure 3.9), or ~55% of seawater (Figure 3.8D).
Previous modeling studies of Costa Rica predict high flow rates along faults, sufficient to advect solutes from the plate boundary at depth to the seafloor (Lauer and Saffer, 2012). Geochemical signals associated with deeply sourced reactions should therefore be effectively concentrated with strong and highly localized expression where splay faults intersect the seafloor, or near structures hydraulically connected to the plate boundary (e.g. mud mounds, mud volcanoes; Henry et al., 1996, Morris and Villinger, 2006).

The results presented here are consistent with other (erosive and accretionary) margins characterized by low heat flow. In northeast Japan, progressive freshening with depth has been attributed to the mobilization of mineral bound water, expelled along faults and fractures, (Kopf, et al., 2003), the same mechanism modeled by Bekins et al. (2004) for the Barbados margin. Additionally, in Cascadia—characterized by high heat flow, the pattern of freshening is consistent with progressive illitization landward of the trench (Torres et al., 2004), suggesting that porosity reduction plays a central role in determining the freshening potential of dehydration. In these previous studies, it was shown that fresh water sources from dehydration increase arcward, as the sediments are exposed to higher temperatures at longer exposure times. The current study complements previous work, by showing that the observed decrease in chlorinity must also coincide with a sufficiently depleted pore space.
Figure 3.7: Results from flow and transport model using $k_{decollement}=k_{faults}=10^{-14}$ m$^2$. Geochemically distinct fluid from illitization is effectively drained by faults prior to reaching the trench. Simulated borehole profiles (1,2) illustrate chloride depletion where profile crosses the fault zone.
Figure 3.8: Fluid fluxes at the seafloor for a “hot” and “cold” margin using $k_{	ext{decollement}} = k_{	ext{faults}} = 10^{14} \text{ m}^2$, presented in terms of: A) percentage of the total fluid source exiting at each of the seven faults, also shown as volume flux per year (RHS) B) fluid budget partitioning for the total fluid source, C.) fluid budget for sources derived from illitization, D) calculated chloride concentration of fluids exiting the seafloor along faults (mM), also shown as percentage of flux volume (RHS).

3.4.3 Pore Pressure Distribution

We report simulated pore pressures in terms of the overpressure ratio ($\lambda$):

$$\lambda = \frac{P_f}{P_i} = \frac{P_f}{\rho_b g z}$$ (11)
where $P_f$ is the fluid pressure, $P_l$ is the lithostatic pressure, and $g$ is gravitational acceleration. We find that the highest overpressure ratio occurs at the base of the slope sediments, where splay faults intersect the overlying less permeable slope sediments, effectively “trapping” fluid pressure (Figure 3.9). Previous studies have shown that dipping permeable layers tend to focus overpressures at their updip ends through a phenomenon known as the “centroid effect” (e.g. Flemings et al., 2002). In our case, pressures are translated along the faults to the base of the slope sediments. The fluid pressure exceeds hydrostatic values in all of our simulations, and the magnitude of the overpressure is sensitive to the splay fault permeability (Figure 3.9). In simulations where $k_{\text{fault}} \geq k_{\text{decollement}}$, fluid pressures are efficiently translated to the base of the sediments; in these scenarios, simulated pore pressures at the upper edge of the splay faults is near lithostatic values across the model domain (Figure 3.9). In cases where $k_{\text{fault}} \leq k_{\text{decollement}}$, fluid pressures at the base of the sediments are lower, and decrease systematically with decreased splay fault permeability. We attribute this behavior to the ability of more permeable faults to both access and transport fluids from depth, and their capacity to capture fluids from the décollement, as they represent a more permeable pathway to the seafloor.

The systematic variations in pressure translation with splay fault permeability are consistent with previous work showing that increased splay fault $k$ results in greater capture of fluids at depth that would otherwise be channeled along the décollement or seep out of the system diffusely (Lauer & Saffer, 2012). The presence of lithostatic pressures identifies a potential mechanism for the generation of mound structures or mud volcanoes that form as a result of unsustainable pressures at the base of the sediment pile, generated through the efficient transmission of fluids along faults (Milkov, 2000). Geochemical sampling of fluids venting from mound structures offshore Costa Rica document a strong signature of dehydration reactions (Hensen et al., 2004), which supports the assertion that these structures tend to form above faults that connect the plate boundary to the seafloor (Sahling et al., 2008).
Model fluid pressures in the underthrust sediments are highest about 12-km from the trench and do not exceed $P_L$, except in simulations with both fault and décollement permeabilities $< 10^{-15}$ m$^2$. The highest pressures are localized in the pelagic sediment section throughout the model domain (Figure 3.10). A second region of anomalous pressure occurs ~ 45-km from the trench, corresponding to the peak in simulated dehydration sources, where they exceed fluid sources from compaction (Figure 3.6). The impact of these reactions on geochemistry and fluid pressure is augmented by the reactions occurring in sediments with greatly diminished porosity and permeability, where fluid pressures cannot dissipate at a rate that exceeds their progressive compaction.

We evaluate the impact of fluid pressure on splay fault strength by calculating excess fluid pressure along each of the faults, or the magnitude of fluid pressure above the hydrostatic value. The pore pressure in all of the faults lie between hydrostatic and lithostatic values and steadily decline as a percentage of the lithostat, or lambda (Figure 3.11).
Figure 3.9: Overpressure ratio ($\lambda$) calculated at the top of the seven faults, where they intersect the base of the slope sediments. Fault permeability was varied over the range indicated while keeping $k_{decollement}$ constant, at $10^{-14}$ m$^2$. Values greater than 1.0 signify lithostatic pressures. Lower plot shows the volume and concentration of fluid sources in the underthrust sediments, following geochemical modeling of smectite transformation, and a mixing model that accounts for progressive compaction.
Figure 3.10: Results from flow and transport model using $k_{\text{decollement}} = k_{\text{faults}} = 10^{-14}$ m$^2$. Geochemically distinct fluid from illitization is effectively drained by faults prior to reaching the trench. Simulated borehole profiles (1,2) illustrate chloride depletion where profile crosses the fault zone.
Figure 3.11: Fault strength in terms of overpressure is plotted for each of the seven faults as a function of the overpressure ratio, or lambda.

3.5 Summary/Conclusions

We present the first comprehensive results that combine geochemical modeling of clay dehydration, with fluid flow and transport models of the Costa Rica forearc, using a permeability architecture that acknowledges the hydraulic connection represented by splay faults. The results are consistent with observations at fluid seeps and regions of focused flow on the upper slope that suggest a hydraulic connection from the plate boundary to the seafloor, through faults that efficiently channel fluids and solutes from the plate boundary to the seafloor. Where faults intersect the slope sediments, near-lithostatic pressures develop at the base of the sediments, as fluid pressures are translated upward along the dipping faults (Figure 3.9, Figure 3.10). We suggest that the extreme pressures provide a potential mechanism for the formation of mud volcanoes that form above splay faults through the development of a piercement structure that initiates their formation. This interpretation is further supported by the hypothesis that fluid
pressures provide the main driving force for mud volcano formation (Milkov, 2000). The modeled distribution of deeply sources fluid is also consistent with observations of depleted chloride with increasing depth in this and other subduction zones (eg. Barbados, NE Japan, Cascadia).

Our results indicate that splay faults exert a primary control on the distribution of deeply derived fluids and fluid pressures. Faults effectively tap into the zone of peak freshening, especially in the region where the drainage path length to the seafloor, or hydraulic impedance is less than the path via the décollement to the trench. Faults near the trench transmit a larger percentage of the total fluid source, but the fluids they access are not as depleted in Cl⁻, and exhibit a weaker signal of freshening. Excess fluid pressures in the underthrust sediments peak 12-km from the trench, with a second peak occurring at 40-km, corresponding to the peak in mineral dehydration. At their peak, fluid sources derived from illitization exceed the volume of compaction-derived fluids, in both hot and cold margins.

The fluid distribution presented here suggests that efforts to sample deeply derived fluids should focus on the mid to upper slope and not near the trench. Additionally, the model results provide a useful context for considering the distribution of fluid release at other margins characterized by higher heat flow, and dehydration reactions that occur closer to the trench.

3.6 References


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Chapter 4
Along strike variability in clay dehydration sources: implications for fault slip behavior in Costa Rica

Abstract
A large-scale heat-flow mapping campaign supplied constraints for new thermal models of the Costa Rica margin, which predict much lower temperatures at the plate boundary than previously considered. The new models, informed by unprecedented coverage of the margin, provide an ideal opportunity to re-evaluate clay dehydration and consider along strike variations in greater detail, along 16-transects that traverse the plate boundary. Here, I evaluate smectite transformation to illite along a 500-km section of the plate boundary, to assess the implications of the new thermal models in shifting the reaction further down-dip, and into the seismogenic zone. Smectite transformation is primarily a kinetically controlled reaction that results in the expulsion of fresh water between 50 and 150°C. This reaction has been associated with the up-dip limit of seismicity, as fluid sources are exhausted, and the fault recovers strength through a decline in fluid pressure. We use the results of the newly modeled thermal structure to simulate the transformation of smectite along the plate boundary, and consider reaction progress with the distribution of dehydration sources and their spatial correlation with a range of fault-slip behavior. We find that the largest dehydration sources track the 120°C isotherm around Nicoya, and to the southeast, with peak dehydration occurring at depths of ~11-km in warm crust, and ~18-km in cold crust, and are spatially correlated with the rupture patch of all major earthquakes since 1950. We calculate the pressurization potential from illitization sources, and determine overpressures are likely to develop ~35-50-km from the trench, at depths ~8-12-km below seafloor. LFEs and slow-slip occur immediately down-dip of the reaction midpoint, where the
mole fraction of Smectite-Illite in the mixed layer clays is 50% Smectite and 50% Illite. We find that the transition from aseismic to seismic behavior correlates well with the temperature range associated with the precipitation of silica produced through smectite transformation, which stiffens the sediment through chemical compaction. Although not implicitly modeled here, this process may help explain the onset of seismicity, either through changes in stiffness, or by further enhancing pressurization through precipitation that reduces the matrix permeability.

4.0 Introduction

Approximately half of Earth’s subduction zones are non-accretionary, in that the sediments on the incoming oceanic plate are entirely subducted rather than accreted to the overriding plate. It is generally accepted that the plate boundary will localize within these sediments that separate the overriding plate from the subducting oceanic crust, which emphasizes their importance in earthquake mechanics through their mechanical and hydraulic influence on the plate boundary fault (Scholz, 1998; Scholz 2000). Fluid pressures at the plate boundary have been implicated in numerous fault-related processes, from slow-slip and tremor (Kodaira et al., 2004; Ito and Ibara, 2006; Shelly et al., 2006; Ito et al., 2007) to hydrofracturing of the upper plate: the proposed mechanism for subduction erosion (von Huene et al., 2000; Ranero et al., 2008). Excess fluid pressures are driven by rapid tectonic loading combined with mineral dehydration reactions. Of the latter, clay transformation is the largest by volume and occurs first (Bethke, 1986; Kastner and Elderfield, 1993; Bekins et al., 1994).

Costa Rica represents an ideal location for investigating the relationship between seismicity and the distribution of fluid sources and potential fluid pressurization at an “erosive” margin, as a site with ocean drilling and observatory installations (CORKs) (Morris et al., 2003; Davis et al., 2006; Kastner), an OBS deployment site (Newman et al., 2002; Norabuena et al.,
2004; DeShon et al., 2006) the site of a large-scale heat-flow campaign (Fisher et al., 2001; Hutnak et al., 2007) as well as an extensive land based GPS-network (Newman et al., 2002; Norabuena et al., 2004; Outerbridge et al., 2010). Submarine mapping with high-resolution bathymetry complements the dataset by identifying regions of focused flow, or seeps at the seafloor (Hensen et al, 2004; Sahling et al., 2008; Kluesner et al, 2013), thought to represent a hydraulic connection from much greater depths, where fluid sources are primarily associated with clay transformation in the incoming sediments (e.g., Lauer & Saffer, 2012).

Figure 4.1: Map of Costa Rica showing the location of the 16 heat flow transects used to model the thermal structure of the margin (Harris et al., 2010 a,b). We use the modeled temperatures along these transects as inputs for our clay transformation model.
A large-scale heat-flow campaign along 16-transects, complemented by shallow seismic data was used to develop a series of well-constrained 2-D thermal models of the forearc, spanning a 500 km length of the margin along strike (Figure 4.1) [Harris et al., 2010]. The new models, informed by unprecedented coverage of the margin, provide an ideal opportunity to investigate the detailed distribution of clay transformation both down-dip and along-strike the Costa Rica margin, and quantify associated fluid release in the sediments as a function of plate rate (exposure time), forearc geometry, and plate-boundary temperature. Here, we use a kinetic expression representative of subduction zone temperatures and time scales (Pytte and Reynolds, 1988) to 1.) Assess the spatial distribution of fluid sources at the plate boundary 2.) Consider the potential contribution of clay transformation to the generation of excess pore fluid pressures along the base of the forearc, and 3.) Investigate the relationship between clay transformation, fluid release, and the distribution of seismicity and slow slip over this region (e.g., Spinelli & Saffer, 2004; Spinelli et al., 2006).

4.1 Geologic Setting

4.1.1 Tectonic and Thermal Structure

The Middle America Trench (MAT) offshore Costa Rica is formed by oblique subduction of the Cocos plate beneath the Caribbean plate, with measured plate velocities ~8-9-cm yr⁻¹ (DeMets, 2001). The overriding plate is an extension of the onshore basement Nicoya complex, which is covered by up to 2-km of slope sediments (Kimura et al., 1997). The margin is generally considered non-accretionary, in that all of the incoming sediment section is subducted rather than accreted to the upper plate (Von Huene, 1991). The total sediment thickness varies from 350-450-m along-strike, with ~250-m of pelagic carbonate ooze at the base of ~150-m of clay-rich
diatomaceous mud (Kimura et al., 1997).

A triple junction trace offshore Nicoya peninsula delineates similar age (~24 Ma, Barckhausen, 2001) crust originating at the Coco Nazca spreading center (CNS), and at the East Pacific Rise (EPR) (Figure 4.1). Despite a similar age across the suture, the EPR crust is “cold” for 24-Ma crust, attributed to hydraulic mining of heat from the oceanic basement to the seafloor through seamounts that extend above the low permeability sediments blanketing the incoming plate (Fisher et al., 2002, Fisher et al., 2006). The geometry of the forearc is well characterized by multiple seismic reflection and refraction studies of the region (e.g. Ye et al., 1996; Christesen et al., 1999; Sallares et al., 2000), which demonstrate the along strike variability in dip, with shallower dip observed in southern Costa Rica (Figure 4.2). This region has exhibited a wide range of fault behavior at the plate boundary, from repeated large magnitude earthquakes to regions of recorded episodic tremor and slip (ETS, Brown et al., 2009; Outerbridge et al., 2010) and slow-slip events (SSEs, Outerbridge et al., 2010; Feng et al., 2012), with detection potential of smaller events amplified in the region surrounding the well-instrumented Nicoya peninsula. This margin is further characterized by a complex seafloor morphology and heat flow regime, dominated by hydrothermal circulation at seamounts in the cooler crust, which provide a surface expression of the basement that extends above the low-permeability slope sediments.
4.1.2 Previous investigations of thermal structure and relationship to seismicity

Previous studies observed a thermal dependence on the depth of seismicity for interplate earthquakes (Newman, et al., 2002), based on the ~10-km shift in the up-dip limit of seismicity across the suture dividing CNS and EPR derived crust. To the north, the up-dip limit of the seismogenic zone lies 20-km below sea floor (bsf) and shifts to 10-km bsf south of the transition, in warmer CNS crust. Subducted sediments in this region contain large diagenetic fluid sources, the bulk of which are bound in the mineral smectite (Moore and Vrolijk, 1992) which releases water through its transformation to illite, a kinetically controlled process that varies according to the thermal history of the sediment (Colton-Bradley, 1987). In low permeability marine sediments, the release of bound water promotes pressurization of the pore fluids, and a
commensurate decrease in the effective stress. Moore and Saffer (2001) proposed that the
exhaustion of these fluid sources leads to a decline in fluid pressure, and an increase in effective
stress that may be instrumental in mechanically determining the updip-limit of the seismogenic
zone (Scholz, 1998). To investigate the potential relationship between mineral dehydration and
the up-dip limit of the seismogenic zone, Spinelli and Saffer (2004) used heat flow data to model
the thermal structure of the forear along transects representative of both CNS and EPR crust. The
thermal models, together with XRD analysis of the incoming sediment (Spinelli and Underwood,
2004) were used as inputs for simulation of clay transformation and associated fluid release, and
to calculate fluid pressures at the plate boundary from the flow models. Their results show that
the onset of microseismicity correlates well with the decay of illitization fluid sources, and the
associated decrease in fluid pressures. (e.g., Moore and Saffer, 2001; Spinelli and Saffer, 2004).

The along-strike variability in thermal structure was further characterized by an extensive
heat-flow campaign complemented by constraints on the base of gas hydrate stability, as defined
by the bottom simulating reflector (BSR) as a proxy for the shallow geothermal gradient
(Yamano, et al., 1982). These efforts resulted in coverage along 16-transects that traverse ~500-
km N-S along-strike (Harris et al., 2010ab), from the northern edge of the Nicoya Peninsula to
north of the Osa peninsula (Figure 4.1) These data were used to constrain new thermal models of
the Costa Rica margin that provide unprecedented coverage along-strike of the temperature
distribution along and immediately beneath the plate boundary (Figure 4.1).

The new thermal models suggest that the margin is significantly colder than previously
thought (Harris et al., 2010 a,b). For example, the location of the modeled 100°C isotherm is
shifted by as much as 20-km landward in the southern portion of the Nicoya peninsula relative to
previous thermal models (Figure 4.3), which predicts a commensurate shift in the location of
smectite transformation (Figure 4.4). Here, we 1.) Investigate the distribution of fluid release
from clay transformation expected based on this new generation of thermal models, 2.) Explore
the contribution of the associated fluid release to generation of excess pore fluid pressure, and 3.) Evaluate the relationship of smectite transformation to the slip behavior of the plate boundary megathrust.

4.1.3 Implications of revised thermal structure

The thermal structure of Harris et al., 2010ab suggests the margin is colder than previously considered, with the 100°C isotherm shifted as much as 20-km landward of the trench in the southern portion of Nicoya (Figure 4.3). The shift in thermal structure translates to a proportionate shift in the location of kinetically controlled reactions down-dip, where porosity and permeability are further reduced through compaction and tectonic burial. In this region, reaction derived fluid is competing for diminishing pore space and the potential for pressurization is augmented as a result. Here, we use the distribution of the fluid sources as a proxy for potential fluid pressurization, and evaluate the role of smectite dehydration in influencing the slip behavior of the plate boundary.
Figure 4.3: Contoured isotherms of the plate boundary comparing the results of previous thermal model (Spinelli et al., 2006, orange dashed line), with the 2010 Harris et al. model (colored contours labeled with white boxes). The arrows illustrate the shift landward of the 100 and 250°C isotherm.
Figure 4.4: Temperature profiles for the 16-transects used in the current study, compared with previous thermal models. Dehydration models of this margin predict peak dehydration occurs at ~120°C (Spinelli et al., 2006; Lauer and Saffer, 2013, in prep).

4.2 Methods

4.2.1 Methods: Simulation of Smectite Transformation

The largest fluid source from mineral dehydration results from the conversion of smectite-group clay minerals to illite (Moore and Vrolijk, 1992), a reaction that occurs in a series of steps that are well-described by a single “lumped” kinetic expression (eg. Pytte and Reynolds, 1989; Huang et al., 1993). The reaction progress is determined primarily by the temperature field
and exposure time of the sediments, which are dictated by the thermal structure of the plate boundary and plate convergence rate (Saffer et al., 2008). In our models, we consider the incoming sediment package as a 1-D column that is progressively conveyed down-dip through the model thermal structure (Harris et al, 2010ab). In Costa Rica, the thermal state of the incoming plate varies, but the age of the plate and thin sediment cover do not result in temperatures high enough to initiate the reaction until tens of km landward of the trench (e.g., Spinelli & Saffer, 2004). In our simulations, we use the Pytte and Reynolds (1988) kinetic expression, which provides the best fit to the available data that span the range of temporal and thermal conditions experienced by subducted sediments, and matches well with observed reaction progress at other subduction zones (Bekins et al., 1994; Saffer et al., 2008; Lauer and Saffer, 2013). The reaction rate is given by:

$$\frac{\partial S}{\partial t} = -A e^{-E_{RT}/RT} \left( \frac{K^+}{Na^+} \right)^n S^n$$

where $S$ is the mole fraction of smectite; $A$ is a scaling factor $(5.2 \times 10^{-7} \text{ s}^{-1})$; $E$ is the reaction activation energy $(1.38 \times 10^{-5} \text{ J/mol})$; $R$ is the ideal gas constant; $T$ is temperature (Kelvin); $n$ represents the reaction order with respect to smectite, or $n=5$. In applying equation (1), we assume that the ratio of $K^+$ to $Na^+$ is in equilibrium with K-feldspar, and varies with temperature as (Pytte and Reynolds, 1988):

$$\left( \frac{K^+}{Na^+} \right) = 74.2 e^{-2490/T}$$

We define temperature within the subducting sediment section according to the thermal model results of Harris et al., 2010a/b, along the 16-transects shown in Figure 4.1. For our models, we consider an end-member scenario, in which all of the interlayer clay is unreacted smectite, or $S_{\text{initial}}=1.0$. Analysis of sediment inputs indicate a mole fraction of $\sim90\%$ smectite in the mixed
layer clays (Spinelli and Underwood, 2004), slightly less than our end member scenario. The 16-transects are oriented approximately perpendicular to the trench, and extend 500-km N-S along strike.

Fluid production associated with illitization is calculated by:

$$\Gamma_{\text{illitization}} = \frac{\partial S}{\partial t} \cdot H \cdot C \cdot (1 - \phi)$$ (3)

where $\Gamma$ is the source term ($V_{\text{fluid}}/V_{\text{sediment}}/s$), $H$ is the volumetric water content of smectite, $C$ is the volume fraction of smectite in the bulk sediment, and $\phi$ is the sediment porosity. We assume there are two layers of bound water, corresponding to a volumetric water content ($H$) of 0.4 (Colten Bradley, 1987; Bird, 1984), and the initial volume fraction of smectite is 50% of the bulk material ($C=0.5$, Equation 3). Smectite supply to the margin is constrained by XRD analysis of drilling inputs offshore Nicoya (Site 1039; Spinelli and Underwood, 2004) and we assume the same available smectite for each of the 16-transects, with the supposition that small variability along-strike will not affect the spatial distribution of the peak-dehydration sources. Porosity is varied as a function of burial depth (bsf), using the depth of the plate boundary from thermal models, and data from ocean drilling and exhumed subduction complexes to characterize progressive deformation (e.g. Lauer and Saffer, 2013; Bray and Karig, 1995).

For our submarine system, we assume that $K^+$ is readily available, and the reaction is therefore not limited. If we consider a case in which $K^+$ is limited (Perry and Hower, 1970), our simulated fluid production rates would be lower, and the reaction would extend to greater depths.

In our calculation of the source terms, we further assume: 1.) any porosity increase associated with a decrease in the volume of solid phases during illitization of the unlithified sediments is ephemeral, and the pore space collapses to the previous value (e.g., Bangs et al., 1990; Bekins et al., 1995); and 2.) a 5% increase in fluid volume occurs through expulsion of the interlayer water.
(Bethke 1986; Bekins et al., 1995). Considering these assumptions, the dehydration fluid source terms we report should be considered a maximum end-member.

4.3 Results and Discussion

4.3.1 Simulated Reaction Progress

The modeled transformation from smectite to illite initiates 12-30-km from the trench, at corresponding depths of ~2.5 and 7-km below seafloor respectively, with the onset of illitization controlled primarily by the thermal state of the incoming crust. Sediments entering the subduction zone on crust south of the thermal transition are considerably warmer than those conveyed via EPR derived, cooler crust, and the kinetically-controlled reaction initiates closer to the trench, and at shallower depths (Figure 4.6). The reaction midpoint, where the smectite-illite ratio is 50%, is achieved at depths of 12-15-km in warmer crust, approximately 60-km landward of the trench, while on the cold side of the thermal transition, the reaction midpoint is reached ~80-km from the trench, at depths of ~23-km. To illustrate the effect of the thermal structure on reaction progress, Figure 4.7 demonstrates simulated reaction progress as a function of depth for profile bgr44 and cr8, located adjacent to the thermal transition in cold and warm crust respectively (Harris et al., 2010; Figure 4.1).

The transformation of smectite to illite progresses in a series of steps that are not explicitly modeled in the Pytte and Reynolds expression, but may play an important role in determining the mechanical properties of the sediments through chemical processes associated with illitization. In particular, the release of silica through smectite dissolution may promote local precipitation of silica near the reaction site, which has been implicated as a mechanism for an observed marked increase in \( v_p \) and \( v_s \) velocities at associated depths, through an increased
stiffness of the sediment matrix resulting from precipitated silica (Peltonen, et al., 2009; Thyberg et al, 2010). These observations of chemical compaction suggest a temperature threshold of 85°C for the onset of geophysically observable chemical compaction, associated with quartz precipitation that cements the clay matrix. In warm crust, 85°C is achieved at depths of ~7-km, and ~30-km from the trench, while in cold crust 85°C is not reached until sediments achieve a burial depth of ~15-km, or 70-km from the trench.

Figure 4.5: Profiles of modeled transects showing the temperature at the plate boundary, chloride distribution of the fluid sources, fluid sources in terms of $V_{\text{fluid}}/V_{\text{sediment}}*\text{s}^{-1}$, and smectite fraction.
Figure 4.6: Plate boundary geometry and thermal structure as a function of depth, and simulated reaction progress for profiles bgr44 and cr8, located adjacent to the thermal transition in cold (blue dashed line) and warm crust (red solid line) respectively (Harris et al., 2010; Figure 1).

4.3.2 Distribution of Fluid Release

Fluid sources released through illitization peak between ~60 and 85-km landward of the trench, at corresponding depths of ~10 and 18-km bsf, south and north of the thermal transition respectively (Figure 4.6, 4.7). The source magnitude is marginally consistent at $1.0 \times 10^{-14} \left[ V_f/V_s^{*} s^{-1} \right]$ throughout the model domain, although colder sediments yield 10% more fluids than sediments delivered via warm crust, owing to the reaction initiating at a greater depth, and a corresponding lower porosity environment.
There is a clear shift in the distribution of fluid sources on either side of the thermal boundary, with reactions in the cold EPR crust initiating and reaching their peak deeper within...
the subduction zone. The transition bisects the Nicoya peninsula between profile CR8 to the northwest and CR9 to the southeast, as depicted in both profile and contour view of the plate boundary (Figures 4.7, 4.8). The peak reaction generally follows the 120-degree isotherm, which wraps around the coastline of Nicoya peninsula, and continues to follow the coast to the southeast. The modeled distribution of fluid sources is consistent with previous clay dehydration studies, that predict peak fluid sources ~120°C, corresponding to a smectite fraction of ~0.5 (Moore and Saffer, 2001; Spinelli & Saffer, 2004; Saffer et al., 2008). The position of the 120°C isotherm shifts farther from the trench to the southeast, where the taper angle is shallower (Figure 4.2), resulting in a smaller geothermal gradient, and a commensurate shift in reaction progress.

We consider the potential contribution of clay transformation to the generation of excess pore pressure in the subducting sediments using a metric that acknowledges the dynamics between fluid production, permeability structure of the region, and the distance to a “free” surface, or in this case, the relatively permeable décollement. Neuzil (1995) showed that for a variety of geologic settings, length scales, and forcing mechanisms, anomalous pore pressures are expected when:

$$\frac{\Gamma L}{K} > 1$$

where \(\Gamma\) is the strength of fluid sources (\(V_{\text{fluid}}/V_{\text{sediment}}\ s^{-1}\)), \(L\) is characteristic drainage path length (discussed below), and \(K\) is the effective (bulk) hydraulic conductivity along the drainage path (m/s), which decreases with progressive burial. For our system, the pressurization potential \((P_{pot})\) is tied to smectite dehydration and associated fluid release (\(\Gamma\)) in the overlying or hemipelagic section of the underthrust sediments, and the distance to the décollement, \(L\), is half of the layer thickness.

We consider the décollement a “free” surface, based on the permeability distribution; initial sediment permeability in the sediments is several orders of magnitude lower than the plate
boundary décollement (e.g. Spinelli and Saffer, 2006; Lauer and Saffer, 2012), and the sediment permeability decreases monotonically with progressive subduction. At depths associated with clay transformation, the sediment permeability is likely to be more than six orders of magnitude lower than the décollement (e.g., Spinelli et al., 2006; Saffer & Tobin, 2011). Drilling of sediment inputs identify a 150-m section of hemipelagic sediment (DSDP Leg 84; ODP Legs 170 and 205), with >60% initial porosity. To determine the evolution of the layer thickness (and L), we use a porosity distribution that honors available data for exhumed subduction complexes and experimental data from recovered samples, and assume a constant solid height for the sediment column (e.g. Lauer and Saffer, 2012). At a distance of ~40-km from the trench, the porosity is estimated to have reduced from 70% to 5%, and the projected height of the sediment column is ~47-m, corresponding to a path length, L of ~23-meters. Permeability values for the sediments are calculated from experimentally derived relationships for porosity-permeability evolution within the hemipelagic sediments (Saffer and Mckiernan, 2005), and we calculate hydraulic conductivity directly from permeability, with viscosity varying as a function of temperature (Chapter 3, Lauer PSU dissertation, 2013).

Calculated values for $P_{p_{\text{sat}}} \left( \frac{\Gamma L}{K} \right)$ exceed unity between 35 and 50-km from the trench, indicative of expected fluid overpressures driven by clay transformation in the underthrust sediments (Figure 4.8). The region of these expected overpressures extends landward ~135-km from the trench on both sides of the thermal transition, where fluid sources associated with dehydration (\( \Gamma \)) are essentially exhausted.
Figure 4.8: Profiles of transect cr8 (red) and bgr 44 (blue), showing the variability in thermal structure, and forearc geometry that give rise to differences in fluid source distribution and magnitude. The calculated Neuzil parameter predicts overpressures at a distance of 35-km from the trench in “warm” CNS crust (cr8, red shaded area), and 50-km from the trench in the cooler, EPR crust (bgr44, blue shaded area).

Differences in source magnitude and pressurization potential result from a combination of forearc geometry or taper angle, the thermal structure, and the porosity distribution calculated as a function of burial depth. Transect bgr44 exhibits lower heat flow values, and temperatures required for clay dehydration to initiate are not reached until 35-km from the trench, compared with cr8, where reactions kick in ~20-km from the trench. The peak magnitude of the fluid source is notably higher along the cooler transect, bgr44, potentially associated with a longer sediment exposure time in the dehydration temperature window. This difference in the source magnitude is also reflected in the calculated $P_{pots}$, with higher magnitudes calculated for transect cr8, although both transects exhibit pressurization potential at distances >35-km from the trench.
Previous studies have established that the magnitude of excess pore pressure scales systematically with the ratio $GL/K$ [e.g., Saffer & Bekins, 2006], and we propose that this difference in the magnitude of the pressurization potential suggests a correspondingly higher potential for pressurization related seismicity. The pattern observed in the current study is consistent across several subduction margins, which span a wide range of permeability, subduction geometry, burial rate, and sediment thickness, suggesting a simple and useful framework for considering the underlying processes that govern excess pore pressure generation and maintenance. This conceptual model implies that margins characterized by thick and/or low-permeability incoming sediment should be poorly drained, leading to high pore pressure, low fault strength, and small taper angles (e.g., Barker et al. 2009, Saffer & Bekins 2002).

4.3.3 Correlation of reaction progress and fluid release with plate boundary slip behavior

To evaluate the potential impact of smectite transformation on fault slip behavior, we first consider the dehydration-derived source distribution along the surveyed 500-km region of the plate boundary and identify the location of peak sources with respect to the spatial distribution of seismicity, considering the full spectrum from large magnitude earthquakes to slow-slip and tremor. We find that for all major earthquakes in the region since 1950, the rupture patches for events $\geq M_w 7.0$ correlate spatially with simulated peak dehydration reactions(Figure 4.9). In particular, the 1950 (Güendel, 1986; Avants et al., 2001; Norabuena et al., 2004) and 1978 Nicoya earthquakes (Güendel, 1986; Avants et al., 2001; Norabuena et al., 2004), the 1990 Nicoya Gulf earthquake (Protti et al., 1995), and the 1999 Quepos earthquake (DeShon et al., 2003) all coincide with modeled peak clay dehydration sources within the study area. The epicenters of the 2012 events, as well as the 1950 7.8 event lie at the edge of the peak dehydration reactions, but their rupture patch also correlates with this zone of peak clay dehydration. An
additional large magnitude rupture occurred landward of Osa peninsula that lies outside the current area of investigation.

Figure 4.9: Regional map of Costa Rica showing a range of slip behavior, from large magnitude earthquakes to slow slip, plotted together with contours representing peak dehydration sources. Shown are the location of large earthquakes epicenters since 1950 (red stars), and their associated rupture patch (grey polygons); interplate seismicity (blue and white circles), and patches of low magnitude earthquakes.
Plate boundary seismicity is well documented in the Costa Rica margin through a combination of geodetic, and OBS deployments that extend coverage offshore, and promote a comprehensive assessment of fault behavior that encompasses slow-slip and tremor and low frequency earthquakes, as well as regular seismicity. In particular, comprehensive studies of the Nicoya region (Newman et al., 2002; Deshon et al., 2006; Outerbridge et al., 2010), and the area northwest of Osa peninsula (Deshon et al., 2003) provide excellent records of plate boundary seismicity in these regions, including a record of slow-slip and LFE beneath the Nicoya peninsula (Outerbridge et al., 2010; Brown et al., 2009). The location of LFEs and slow-slip correlate spatially with a decline in dehydration fluid source magnitude, both immediately down-dip of a smectite/illite fraction of 0.5, where the mole fraction of Smectite-Illite in the mixed layer clays is 50% Smectite and 50% Illite(Figure 4.10). The frequency of microseismicity in the Nicoya region is highest at 18-20-km depth, with a second smaller peak at 35-km depth, correlated with seismicity north of the thermal transition, on the relatively cool EPR crust where dehydration reactions occur deeper in the forearc. A similar bimodal distribution is present in the Quepos earthquake aftershocks to the southeast (yellow circles, Deshon et al., 2003), where the frequency of seismicity is highest at 20-km depth, with a second peak at ~30-35-km bsf. A composite of regional seismicity ≥ Mw 5.0 suggests a peak frequency of large magnitude EQs at 35-km depth, greater depths than those associated with microseismicity, slow-slip, and ETS, although consistent with the deeper aftershocks of the Quepos earthquake.
Figure 4.10: Locations of plate boundary microseismicity (red circles, Newman et al., 2002; Deshon et al., 2006), well-located aftershocks (yellow circles, Deshon et al., 2003), low frequency earthquakes (light blue circles, Brown et al., 2009), slow-slip (dashed grey circle, Outerbridge et al., 2010), and regional seismicity > Mw=5.0 (green circles). The blue solid line delineates the modeled Smectite-Illite transition where S/I is 0.5, or 50% of the mixed layer clays are Smectite, and 50% Illite. Histograms showing the depth distribution of earthquake frequency correspond with colored circles plotted on regional map.
4.3.5 Choice of kinetic expression: Sensitivity analysis

Mineral transformation from smectite to illite, and accompanying fluid release occurs in a series of discrete steps that can be lumped into a single, higher-ordered kinetic expression (Pytte and Reynolds, 1988; Huang et al., 1993). Constraints for the kinetic expressions are derived empirically; using reconstructed thermal histories to interpret observed smectite content (Pytte and Reynolds, 1988), and through high temperature laboratory experiments (Huang et al., 1993). We evaluate two kinetic expressions to simulate illitization within the subducted sediments: the empirically derived expression of Pytte and Reynolds (1988), and the experimentally derived expression of Huang et al. (1993).

In the Huang model, Smectite transformation is described by:

$$\frac{\partial S}{\partial t} = -Ae^{-E_a/RT} [K^+] \cdot S^2$$

(5)

where \([K^+]\) is assumed to be constant, and assigned a value based on pore water composition obtained during Leg 205 drilling (Site(s) 1253, 1254, and 1255; 10 mM). For both kinetic expressions, the reaction progress should be considered as a maximum end member, as we do not consider the potential for potassium depletion, which effectively shuts down the reaction in both models (equation (1) and (5). (Hower et al. 1976; Boles and Franks 1979; Huang et al. 1993).

The Huang model predicts a higher peak magnitude fluid source than the Pytte and Reynolds model (Figure 11), located deeper in the margin, where sediment porosity and permeability are further reduced, and pressurization potential is consequently enhanced. By comparison, the Huang model predicts a greater potential for pressurization, at greater depths than predicted by the Pytte and Reynolds (1988) expression employed in this study. Therefore, the results presented here represent a conservative estimate of the pressurization potential, as compared to the results of the Huang model.
4.4 Summary/Conclusions

The Costa Rica margin is known to experience large magnitude earthquakes that originate at the plate boundary thrust (Güendel, 1986; Protti et al., 1995; Avants et al., 2001; Deshon et al., 2003; Norabuena et al., 2004), where progressively compacted sediments juxtapose crystalline rock, and the décollement is thought to localize along weak surfaces within these underthrust sediments (Wallace et al., 2003; Tobin and Saffer, 2009). Our understanding of the processes that control the sliding behavior of the plate boundary décollement, and the up-dip limit of the seismogenic zone relies on estimates of frictional properties and fluid pressures, which each contribute to weakening through distinct mechanisms. Previous studies interpreted the updip-limit of seismicity to occur where fluid sources decline, and the effective stress increases through a decline in pore pressure which allows the plate boundary to strengthen, and ultimately host a rupture (Moore and Saffer, 2001; Spinelli and Saffer, 2004; Ranero et al., 2008)).

The results of the current study suggest a spatial correlation between plate boundary seismicity, and simulated peak fluid sources from smectite transformation, with both large
magnitude earthquakes and microseismicity associated with regions of peak fluid production (Figure 4.10, Figure 4.12). Additionally, both slow-slip and ETS occur in a region where the predominant interlayer clay transitions from smectite to illite (S/I=0.5), down-dip of peak dehydration, which correlates spatially with large magnitude earthquakes in this region. The locus of seismicity in regions of peak dehydration suggests that these reactions may weaken portions of the plate boundary through pressurization, although the connection between fluid pressure and seismicity is not well understood.

Another process that may play an important role in determining the mechanical properties of the sediments is the release of silica through smectite dissolution, which may promote local precipitation of silica near the reaction site, which stiffens the matrix through chemical compaction. This process has been considered a mechanism for the marked increase in $v_p$ and $v_s$ velocities observed at associated depths, through an increased rigidity of the sediment matrix resulting from precipitated silica (Peltonen, et al., 2009; Thyberg et al, 2010). These observations of chemical compaction suggest a temperature threshold of 85°C for the onset of chemical compaction, which is achieved at depths of ~7-km, and ~30-km from the trench in warm crust, and a depth of ~15-km, or 70-km from the trench in cold crust (Figure 4.12). We suggest that the change in rigidity associated with this reaction, although not implicitly modeled here, may provide a mechanism for the onset of seismicity, given that the temperature associated with the onset of this process occurs at the same depth as the updip limit of the seismogenic zone. While there are certainly many processes that contribute to the transformation of a soft sediment into a lithified material capable of sufficient strain to rupture, chemical compaction appears well correlated with a sharp change in plate boundary behavior, either through changes in stiffness, or by further enhancing pressurization through precipitation that reduces the matrix permeability.
Figure 4.12: Plate boundary temperature for cold (blue) and warm (red) crust for each of the modeled transects, with vertical lines indicating the onset of smectite transformation (60°C, solid), and quartz cementation associated with observed velocity increases and chemical compaction. The horizontal dashed lines show the depth of peak fluid release in warm (red) and cold (blue) crust.
4.5 References


Appendix A

Permeability and consolidation behavior of sediments subducting at the Japan Trench: Implications for physical properties in the high slip region of the 2011 Tohoku-Oki earthquake

1.0 Introduction

In subduction zones, high porosity sediments on the oceanic plate undergo rapid consolidation and commensurate permeability reduction due to tectonically-driven loading. The evolution of sediment permeability during consolidation mediates drainage and fluid pressure distribution, and the position of these sediments puts them in direct contact with the overlying plate boundary megathrust. Therefore, their in situ conditions and physical properties hold important implications for fault behavior, in terms of weakening the plate boundary and contributing to excessive slip. Excess fluid pressures, likely facilitated by low sediment permeability, have been implicated as a potential mechanism for slow-slip and low-frequency earthquakes at many convergent margins (Kodaira et al., 2004; Shelley et al., 2006; Ito et al., 2007; Kitajima and Saffer, 2012), and in the 2011 Tohoku earthquake, elevated fluid pressures provide a potential mechanism for the extraordinarily high slip that propagated to the trench (Kimura et al., 2012). The pressure distribution within the subducted sediments is controlled by the evolution of porosity and permeability as sediments are progressively loaded, and these parameters can be empirically derived through consolidation experiments designed to emulate conditions during tectonic loading.

Deep Sea Drilling Project (DSDP) Leg 56, Hole 436, represents the most proximal sediment inputs to the rupture site (Figure A-1), and therefore an ideal opportunity to experimentally determine the trajectory of sediment parameter evolution from the trench, to depths associated with the onset of excessive slip. Deformation and permeability tests were
conducted under stress paths of isotropic loading and uniaxial strain, at mean effective stresses ranging from ~350 kPa to 65 MPa. Under ideal conditions, the uniaxial consolidation tests enable continuous measurements of porosity and permeability, and flow through tests during isotropic loading in the triaxial vessel are used to confirm these measurements. Together, these results represent 1.) the first efforts to characterize the mechanical and hydraulic behavior of sediments subducting at the Japan trench, as they are exposed to an increasing axial load simulating their burial under progressive subduction (up to 100 MPa axial load), and 2.) the first constraints for extrapolating these parameters farther down-dip, in the high-slip region of the plate boundary.

2.0 Site 436: Setting, Samples, Experimental Method

Site 436 is located near the crest of a swell in the Japan basin, ~100-km seaward of the Japan trench and 200-km north of the high slip region in the March 11, 2011 (Mw) 9.0 Tohoku-Oki earthquake(Figure A-1). The rupture initiated on the plate boundary delineated by subduction of the Pacific Plate beneath Japan at a rate of 8 to 8.5 cm/yr., and extended to the trench where co-seismic slip estimates are as high as 60-meters (Lay et al., 2011). The sediments sampled at Site 436 represent the best example of sediments subducted at the rupture site, and in the area of excessive slip.

Three lithologic units were identified through coring at Site 436 before encountering chert at the base of the section, with total penetration of 397.5-meters below seafloor (mbsf). Unit I and II together comprise 360-meters of silty clay and claystone overlying a 30-m thick section of pelagic clay (Unit III)(The Shipboard Science Party, 1980; Figure A-2). Samples were selected based on their suitability for trimming down to one-inch samples for loading in the experimental apparatus. We obtained two samples from Unit I (100-mbsf; 436-12R), and four
samples from Unit III (360-370-mbsf; 436-39-04, 436-40R-03) for deformation experiments to investigate the evolution of porosity and permeability with progressive subduction.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Sample</th>
<th>depth</th>
<th>Lithologic unit</th>
<th>Initial porosity</th>
<th>Maximum stress (s)</th>
<th>Experiment type</th>
</tr>
</thead>
<tbody>
<tr>
<td>271</td>
<td>436_12R_03</td>
<td>103 mbsf</td>
<td>1A</td>
<td>0.75</td>
<td>97 MPa</td>
<td>Triaxial</td>
</tr>
<tr>
<td>272</td>
<td>436_12R_03</td>
<td>103 mbsf</td>
<td>1A</td>
<td>0.7</td>
<td>86 MPa</td>
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</tr>
<tr>
<td>279</td>
<td>436_12R_03</td>
<td>103 mbsf</td>
<td>1A</td>
<td>0.74</td>
<td>91 MPa</td>
<td>Uniaxial</td>
</tr>
<tr>
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<td>436_39R_04</td>
<td>364 mbsf</td>
<td>2A</td>
<td>0.55</td>
<td>91 MPa</td>
<td>Uniaxial</td>
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<tr>
<td>295</td>
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<td>2A</td>
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<td>93 MPa</td>
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</tr>
<tr>
<td>296</td>
<td>436_40R_03</td>
<td>374 mbsf</td>
<td>2A</td>
<td>0.56</td>
<td>20 Mpa</td>
<td>Triaxial</td>
</tr>
</tbody>
</table>

Table A1. Experimental matrix for consolidation experiments conducted on Site 436 samples.

Consolidation tests were conducted under both isotropic and uniaxial consolidation conditions, if a sufficient volume of sample material was present, in order to verify the consistency and reliability of the permeability measurements, and consider the effects of a different stress path. In the triaxial vessel, permeability was measured during flow through tests, at a range of effective mean stresses from 500 kPa to 90 MPa. A Constant Rate of Strain (CRS) test was conducted in the uniaxial configuration, which enables continuous measurements of porosity and permeability under ideal loading conditions up to axial stresses of 100-MPa (Table A1).

3.0 Methods: Deformation Experiments

3.1 Uniaxial CRS experiments

Each sample was carefully trimmed to a 25.4-mm diameter cylinder and heights ranging from ~15-20-mm. The cylindrical sample was placed within a steel ring inside the consolidation cell and saturated for 24 hours at a back pressure of 300kPa. Upon
saturation, an axial force was applied to the sample up to 45kN at a constant displacement rate of 0.5-1.5 µm/min using a computer controlled load frame, with displacement measured externally via a linear variable differential transformer (LVDT) mounted on 50kN load cell. Throughout the consolidation test, we maintained a constant back pressure of 300 kPa at the top of the sample, with undrained conditions at the base of the sample, where (base) pressure is logged by a pressure transducer.

The pressure gradient across the sample drives flow from the base, to the top of the sample, enabling calculation of hydraulic conductivity using the excess pressure at the base of the sample and the strain rate (ASTM International, 2006; Valdez et al., Site C0019 Data Report, 2013):

\[
K = \frac{\dot{e} H_n H_o \gamma_w}{2 \Delta u} \tag{1}
\]

Where,

\( \dot{e} \) = strain rate,
\( \gamma_w \) = unit weight of water,
\( H_n \) = current sample height,
\( H_o \) = initial sample height, and
\( \Delta u \) = base excess pressure.

Permeability is calculated from hydraulic conductivity using:

\[
k = \frac{K \mu}{\rho_w g} \tag{2}
\]

where \( \rho_w \) is the density of seawater (1024 kg/m3), \( \mu \) is the viscosity at room temperature, \( (10^{-3} \text{ Pa•S}) \), and \( g \) is the gravitational acceleration (9.81 m/s\(^2\)).
Porosity is “back calculated” from the final porosity according to:

\[ \phi_{\text{test}} = \phi_{\text{final}} + \frac{z_{\text{axial}}}{H_o} \]  

(3)

where \( H_o \) is the original sample height, \( \phi_{\text{final}} \) is the final porosity at the end of the test, and \( z_{\text{axial}} \) is the displacement measured by the LVDT.

### 3.2 Triaxial Deformation Experiments

Samples were isotropically loaded by continuously increasing confining pressure from \( \sim 1 \text{-MPa} \) to \( \sim 90 \text{-MPa} \). To measure permeability, we discontinue loading and hold the confining pressure constant, while establishing a pressure gradient across the sample using syring pumps connected to the upstream and downstream ends of the sample. We calculate permeability using a steady-state constant head approach, by monitoring flow into and out of the pumps, using:

\[ k = \frac{Q \cdot \mu \cdot H}{A(P_{\text{upstream}} - P_{\text{downstream}})} \]  

(4)

### 4.0 Results

The porosity-effective mean stress relationships defined by CRS testing are shown in Figure A-4. The shallow sample (U-272-436-12R,U-279-436-12R; 100-mbsf) exhibits a decline in porosity from 0.7 to 0.35 over effective mean stresses up to 80-MPa. The pelagic clay samples follow a similar path in the porosity effective stress plot, with porosity reduction from \( \sim 0.5 \), to a final
value of 0.13 and 0.20 at an effective mean stress of 80-MPa for samples U-279-436-39R(365-mbsf) and U-295-436-40R(371-mbsf), respectively.

The permeability-porosity and permeability-effective mean stress relationships reveal a different trend for the shallow silty clay sample and the pelagic clays (Figure A-4, A-5). In log-log parameter space, the permeability decline with increasing effective stress is approximately linear for sample 12R, with an order of magnitude (OOM) decline in permeability, over an OOM increase in effective mean stress (Figure A-5). The pelagic clay samples exhibit a similar decline in permeability with increasing effective stress, but calculated permeability values are nearly an OOM lower than in the shallower silty-clay section (12R).

To compare the experimental results with other low permeability clays and shales, we overlay our data on Neuzil’s compendium publication for argillaceous media (Neuzil, 1994; shaded area), and delineate the findings of Skarbek and Saffer, 2009, using inputs to the Nankai trough, southeast Japan, Site 1173. (Skarbek and Saffer, 2009; Figure A-6; blue lines). Calculated permeability-porosity values for 12R (100-mbsf) lie within the range reported by Neuzil, and for the Nankai sediments, contrasted by the pelagic clays, which both (39R and 40R) fall below the lower limits for the Nankai inputs, and the basal unit (40R) plots at the edge of Neuzil’s compilation for low-permeability media (Neuzil, 1994). In particular, the experimental data for sample 40R trace the lower limits of Gulf of Mexico deposits (labeled ‘6’; Bryant et al., 1975), Pierre Shale claystones (labeled ‘8’; Neuzil., unpublished data, 1987), and clayey siltstones from western Canada (labeled ‘9’; Young et al, 1964), and permeability values approach $10^{-21}$ m$^2$ at a porosity of 20%.
Figure A-1: Map showing the location of the March 11, 2011 ($M_w$) 9.0 Tohoku-Oki earthquake, and the only reference site for inputs to the Japan trench (Site 436).
Figure A-2: Stratigraphic column of sediments identified through drilling at Site 436.
Figure A-3: Example of data obtained during consolidation testing of sample 436_40R_03.
Figure A-4: Porosity reduction plotted as a function of effective mean stress for the silty shallow lithology (12R, 50-mbsf), and the pelagic clays at the base of the inputs (39R, 40R). Axial load was increased up to ~90-MPa MPa for each of the tests.
Figure A-5: Permeability as a function of effective mean stress for uniaxial and triaxial stress paths. Permeability measurements were conducted under flow through conditions, and calculated from Darcy’s law.
Figure A-6: The results of the current study overlain on the compendium publication for argillaceous sediments (grey shaded area, Neuzil, 1994). Blue lines delineate the findings of Skarbek and Saffer, 2009, using inputs to the Nankai trough, southeast Japan, Site 1173. (Skarbek and Saffer, 2009).
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Professional Experience
2008-today: Graduate Student, Research Assistant, Pennsylvania State University
Using numerical modeling to obtain fluid budgets, and pressure distribution within subduction zones. Initial work focused on Costa Rica, with additional soil mechanics work conducted using samples obtained through IODP drilling.

Participated in all aspects of large-scale electrical resistivity surveys including data collection, processing, interpretation and report development. Follow up data collection included the use of the SASW method (Spectral Analysis of Surface Waves) to confirm the location of rock/soil transitions.

2002-2004: Graduate Teaching Fellow, Lab Coordinator, Radford University

1999-2002: Geophysical Team Leader, NAEVA Geophysics, Inc. Charlottesville, VA
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