Along-strike variations in underthrust sediment dewatering on the Nicoya margin, Costa Rica related to the updip limit of seismicity

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Received 17 October 2003; revised 6 January 2004; accepted 20 January 2004; published 28 February 2004.

[1] Along the Costa Rican subduction zone offshore the Nicoya peninsula, an offset in the updip limit of seismicity coincides with a transition between subduction of warm crust generated at the Cocos-Nazca Spreading Center and cool crust formed at the East Pacific Rise. We evaluate whether the observed difference in thermal state of incoming crust would result in significant differences in sediment dehydration reaction progress along strike, thought to be a control on the updip limit of seismicity. We combine thermal models with models of dehydration reaction kinetics for opal and smectite to estimate the distribution of diagenetic fluid sources. The modeled distribution of diagenetic fluid sources mimics the pattern of the updip limit of seismicity; seismicity begins ∼15–20 km landward of most of the smectite-to-illite transition. This suggests that the location of the updip limit of seismicity may be influenced by the dissipation of fluid overpressures landward of smectite-to-illite dehydration. INDEX TERMS: 3015 Marine Geology and Geophysics: Heat flow (benthic) and hydrothermal processes; 3022 Marine Geology and Geophysics: Marine sediments—processes and transport; 3040 Marine Geology and Geophysics: Plate tectonics (8150, 8155, 8157, 8158); 7209 Seismology: Earthquake dynamics and mechanics; 8105 Tectonophysics: Continental margins and sedimentary basins (1212). Citation: Spinelli, G. A., and D. M. Saffer (2004), Along-strike variations in underthrust sediment dewatering on the Nicoya margin, Costa Rica related to the updip limit of seismicity, Geophys. Res. Lett., 31, L04613, doi:10.1029/2003GL018863.

1. Introduction

[2] Within subduction zones, where large, damaging, and tsunamigenic earthquakes primarily occur [Satake and Tanioka, 1999], processes that control the updip (seaward) limit of seismicity along the plate boundary detachment, or décollement, are not well characterized. Moore and Saffer [2001] posit that diagenetic processes (e.g., opal-to-quartz and smectite-to-illite transformations), low-grade metamorphic reactions, and increasing effective stress (mediated by declining fluid pressure) act together to control the updip limit of the seismogenic zone. Others argue that the updip limit of seismicity results from a change in frictional behavior as smectite is transformed to illite [e.g., Hyndman et al., 1997], although recent laboratory data do not support this hypothesis [Saffer and Marone, 2003].

[3] On the Costa Rica margin, offshore the Nicoya Peninsula, an along-strike offset in seismicity [Newman et al., 2002] coincides with a change in the thermal state of the incoming crust [Fisher et al., 2003]. The temperature history of subducting sediment affects the progress of diagenetic dewatering reactions, and may be important in explaining along-strike changes in seismicity. Better constraining the location where diagenetic reactions occur is important for both understanding factors affecting the updip limit of seismicity, and characterizing subduction zone fluid flow systems.

[4] Compaction and diagenesis control the release of fluids from sediment in subduction zones [e.g., Moore and Vrolijk, 1992]. As the porosity of underthrust sediment decreases during transport farther into the subduction zone, fluid stored within minerals comprises an increasing proportion of the total volume of fluid within the sediment. The hydrologic contribution of underthrust sediment to the plate interface is amplified at non-accretionary margins (e.g., Costa Rica) because the décollement often steps down into these sediments, and they are buried and transported arcward rapidly [e.g., Saffer and Bekins, 1998].

[5] Subduction geometry and the thermal state of crust entering a subduction zone are primary controls on the temperature history of underthrust sediment [e.g., Molnar and England, 1995] and in turn the distribution of diagenetic fluid sources. Here, we model temperatures and the progress of opal and smectite dehydration reactions within sediments underthrust at the Costa Rican margin, to evaluate whether along strike differences in progress of kinetically controlled dehydration reactions are correlated with observed along-strike changes in seismicity. First, we combine a thermal model of the shallow subduction zone with kinetic expressions for reaction progress and observed sediment compositions to estimate the spatial distribution of diagenetic fluid sources. Second, we compare the distribution of estimated temperature and fluid sources to the observed pattern of shallow interplate seismicity.

2. Regional Setting

[6] Along the Pacific margin of Costa Rica (Figure 1a), the Cocos Plate subducts beneath the Caribbean Plate at the Middle America Trench at ∼85 mm/yr [DeMets, 2001]. Most (~99%) of the ~375 m thick incoming sediment column is subducted [Saito and Goldberg, 2001]. The hemipelagic sediment (upper ~150 m) is ~10 wt-% opal and ~60 wt-% smectite on average, with no significant variations along strike [Spinelli and Underwood, 2004].

[7] Offshore the Nicoya Peninsula, a triple junction trace divides the Cocos Plate into crust formed at the East Pacific...
Rise (EPR) to the north and at the Cocos-Nazca Spreading Center (CNS) to the south [Barckhausen et al., 2001]. Seaward of the trench, seafloor heat flow on the CNS crust is 105–115 mW/m² (consistent with conductive lithospheric cooling models) [Fisher et al., 2003]. Heat flow on the EPR crust is 20–40 mW/m²; the lateral transition between warm and cool crust is abrupt (occurring over a distance ≤ 5 km). The low heat flow on EPR crust likely results from hydrothermal circulation [Fisher et al., 2003]. The location of shallow earthquakes along the plate interface also changes along-strike, coinciding with the transition from CNS to EPR crust. Earthquakes occur shallower (~10 km depth) and closer to the trench (~60 km from trench) on the warm CNS crust, than on the cool EPR crust (~20 km depth and ~75 km from trench) (Figure 1) [Newman et al., 2002].

3. Thermal Models

3.1. Method

We model temperatures in the subducted sediment and the overlying margin wedge using a transient 1-D model that accounts for heat advection and conduction during burial of subducted sediment, as well as frictional heating along the plate interface [Ferguson, 1990]. The crust and most of the sediment are subducted; the remaining sediment is stretched to simulate the growth of a margin wedge. The Costa Rica margin wedge is not actively accreting [Kimura et al., 1997]. Most of the wedge consists of Mesozoic Nicoya Complex rocks draped with slope sediment; the frontal ~15 km consists of deformed slope sediment [Vannucchi et al., 2001]. We use only the upper 5 m of the >300 m thick incoming sediment section to generate the margin wedge. This results in the rapid subduction of underthrust sediment beneath a slowly forming wedge that increases in age dramatically with distance from the trench and deforms slowly enough that heat transfer is dominated by conduction. Our model differs from the 2-D model of Harris and Wang [2002], in that we include more detail (higher spatial resolution and variable sediment thermal conductivity), but examine only the first 80 km from the trench (we do not include the landward portion of the margin that may be influenced by mantle wedge flow).

[9] We model temperatures along two cross-sections, one for the CNS portion of the margin and one for the EPR portion. In all models, seafloor temperature is constant (2°C), heat flow at the base of the model is 105 mW/m² (based on 20–25 Ma crust). For frictional heating, we consider a range of effective friction coefficients ($\mu'$) from 0.03–0.175, and report results for an intermediate value ($\mu'$ = 0.075). This range corresponds to a range of friction coefficients from 0.2–0.5, appropriate for mixtures of clay and quartz [e.g., Saffer and Marone, 2003], and a range of pore fluid pressure ratio from 0.65 to 0.85 [Saffer et al., 2000]. Thermal conductivities of pore fluid, sediment grains, and crust are 0.67, 3.0, and 2.5 W/m-K, respectively. The effective conductivity of the sediment is calculated using a geometric mean mixing model for the pore fluid and sediment grains. The slope of the seafloor on the wedge is 5.4°. The subducting plate dips 6° for the first 30 km of the subduction zone, after which the dip angle increases to 13° [Christeson et al., 1999].

[10] For the CNS case, the initial seafloor heat flow (outboard of the trench) is 105 mW/m², and the initial temperature profile is assumed to be purely conductive. Temperatures in the sediment section and underlying crust are calculated by downward integration from the surface heat flow using the thermal conductivity of each material. For the EPR section, initial seafloor heat flow is 20 mW/m². We examine four hydrothermal cooling scenarios. We vary both the thickness of crust that is initially cooled by hydrothermal circulation and the distance into the subduction zone for which the crustal cooling persists. We initially cool either the upper 1 or 2 km of crust. Temperature is constant with depth in the cooled crust (at the temperature of the base of the sediment column). For both initial cooling thicknesses, we examine the effect of varying the distance into the trench over which the crust remains cooled, by stopping hydrothermal cooling of the crust either at the trench or 10 km landward of the trench.

3.2. Results

[11] Modeled seafloor heat flow along the CNS cross-section decreases dramatically within the first 5 km landward of the trench, then gradually decreases beyond that point (Figure 2a). Modeled heat flow along the EPR transect also decreases at the toe of the margin wedge, then increases, approaching the surface heat flow values for the CNS section. Although there is considerable scatter in the seafloor heat flow observations on the EPR transect, the data are consistent with subduction of hydrothermally cooled crust for most of the hydrothermal cooling and frictional heating scenarios we consider (Figure 2a). Surface heat flow data are not available for the margin wedge in the CNS portion of the study area. Results from our models (with hydrothermal cooling stopping at the trench) are similar to results from 2-D finite element models [Harris and Wang, 2002]. Modeled temperatures along the décollement reach 150°C by 40 km from the trench along the CNS transect, and by 48–60 km from the
trench along the cooler EPR transect (Figure 2b). The point at which décollement temperature reaches 150°C could shift ~15 km landward or seaward if extreme effective friction coefficients are applied. As there are no significant along strike variations in subduction geometry, convergence rate, or incoming sediment composition in the study area, the effective friction coefficient is likely similar on the CNS and EPR portions of the margin, resulting in a consistent difference in décollement temperature between the two sections of the margin.

4. Opal and Clay Dehydration

4.1. Method

To estimate reaction progress, we track sediment from its initial position outboard of the trench through the modeled temperature field as it is underthrust [e.g., Bekins et al., 1994]. This provides a thermal history for the sediment, which we combine with kinetic expressions to calculate reaction progress. The smectite-to-illite reaction kinetics we use closely approximate the observed smectite/illite composition of sediment with well-known thermal histories, including those in subduction zone settings [e.g., Pytte and Reynolds, 1988; Bekins et al., 1994]. The opal-to-quartz reaction kinetics employed are derived from laboratory experiments [Ernst and Calvert, 1969].

The process of opal diagenesis is completed at temperatures below ~100°C and releases ~23% of the original volume of opal-A as water [Behl and Garrison, 1994; Moore and Vrolijk, 1992]. Most of the smectite-to-illite reaction progress occurs at higher temperatures (60–160°C) than opal diagenesis [Hower et al., 1976]. The reaction rate decreases as the fraction of smectite remaining in the sediment becomes small. Fluid sources resulting from opal and smectite dewatering, expressed in $V_{\text{fluid}} / V_{\text{sed}} \tau^{-1}$, are calculated based on (1) the progress of the diagenetic

![Figure 2. Modeled seafloor heat flow from the CNS portion (bold black) of the margin is higher than the EPR section (a). The EPR section is modeled with hydrothermal cooling in the upper 1 km (thin black) or 2 km (gray) of crust. Modeled hydrothermal cooling either stops at the trench (solid) or persists for 10 km (dash) past the trench. The gray shaded areas show the range of modeled heat flow and décollement temperature, for the EPR 2km case, resulting from varying the effective friction coefficient ($\mu'$) to extreme values. Lines show results for $\mu' = 0.075$. Modeled décollement temperatures (b) are used to calculate diagenetic reaction progress.](image)

![Figure 3. Diagenetic reactions proceed closer to the trench in the warm CNS portion of the margin than the cool EPR portion. EPR results shown for 2 cases where crustal cooling stops at the trench: 1 km (thin black) or 2 km (gray) of cooled crust, and one case where cooling in the upper 2 km of crust continues 10 km past the trench (dashed gray). Fluid sources from opal dehydration are depleted before peak smectite dehydration (b).](image)
4.2. Results

Vrolijk variations in thermal regime or sediment composition
incoming crust. Over the range where smectite fluid sources
the margin, and 40–65 km into the EPR section. The
op-to-illite reaction occurs 30–50 km into the CNS portion of
EPR section (Figure 3a). The opal-to-quartz transition is a
CNS portion of the subduction zone, and 30–45 km into the
reactions (Figure 3a); (2) the weight percent opal and
smectite in the incoming sediment; (3) sediment porosity;
and (4) the weight percent water in opal and smectite. We
assume that opal and smectite contain 11 wt-% and 20 wt-%
water, respectively [Colten-Bradley, 1987; Moore and Vrolijk, 1992].

4.2. Results

[14] Modeled opal diagenesis occurs 20–25 km into the
CNS portion of the subduction zone, and 30–45 km into the
EPR section (Figure 3a). The opal-to-quartz transition is a
zero-order reaction, therefore, once the reaction begins it
quickly runs to completion (Figure 3). Most of the smectite-
to-illite reaction occurs 30–50 km into the CNS portion of
the margin, and 40–65 km into the EPR section. The
distance into the EPR section at which dewatering reactions
occur varies with the degree of hydrothermal cooling of the
incoming crust. Over the range where smectite fluid sources
peak (~35–55 km), they are larger than the estimated source from sediment compaction (~2 × 10^{-14} – 3 × 10^{-14} V^{-1} s^{-1}).

5. Discussion and Comparison to Seismicity Data

[15] Along the Nicoya margin, large diagenetic fluid sources (>1 × 10^{-14} V^{-1} s^{-1}) are depleted ~15–20 km
seaward of the updip limit of seismicity, along both the CNS
and EPR transects (Figure 4). Thus, an offset in the location
of sediment dewatering based on kinetic models mimics the
observed offset in seismicity along strike [Newman et al.,
2002]. We suggest that fluid produced by dewatering
reactions increases pore fluid pressure in poorly drained
lithology [e.g., Moore and Vrolijk, 1992]. Conversely,
exhaustion of these fluid sources corresponds to an increase
in effective stress as elevated pore fluid pressures associated
with diagenetic dewatering reactions dissipate [e.g., Moore
and Saffer, 2001]. This increase in effective stress may in
part control the updip limit of seismicity [e.g., Scholz,
1998].

[16] Within subducted sediment, significant along-strike variations in thermal regime or sediment composition
should lead to differences in diagenetic reaction progress,
and thus fluid pressure. Along-strike variations in fluid
pressure at the plate interface, in turn, may influence the
location of the updip limit of seismicity.

[17] Acknowledgments. This research was supported by an NSF
Margins post-doctoral fellowship award to Spinelli, NSF OCE-0304946.
We thank two anonymous reviewers for careful and helpful reviews.

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