We report detailed thermal measurements undertaken during Integrated Ocean Drilling Program (IODP) Expeditions 315 and 316 along the Nankai Trough Seismogenic Zone Experiment (NanTroSEIZE) transect offshore the Kii Peninsula, Japan. Geothermal objectives included determining thermophysical rock properties of the cored material and characterizing the background thermal regime along this transect. New thermal conductivity measurements made with a divided bar are reported and supplement shipboard full- and half-space thermal conductivity measurements for a total of 938 thermal conductivity measurements. Thermal conductivity varies between about 1.0 W m$^{-1}$ K$^{-1}$ near the seafloor to 1.6 W m$^{-1}$ K$^{-1}$ at a depth of 1 km. Thermal conductivity generally increases with depth and correlates with variations in porosity and lithology. Thermal gradients along the transect are characterized by 48 sediment temperature measurements from 6 sites. Thermal corrections for the effects of bathymetric relief and sedimentation improve the confidence with which the heat flow values can be interpreted. Heat flow generally decreases with landward distance from the deformation front and varies from 70 mW m$^{-2}$ just landward of the deformation front to 54 mW m$^{-2}$ at sites characterizing the outer fore-arc high and to 57 mW m$^{-2}$ at the Kumano Basin Site. IODP heat flow measurements are significantly lower than nearby seafloor heat flow measurements. This difference is most likely due to variations in bottom water temperature that have a large effect on values of seafloor heat flow. Thus the heat flow of the Nankai accretionary prism is lower than previously thought. We present thermal models of subduction along this transect and explore the impact of the initial geotherm. Conductive plate cooling based on the age of subducting seafloor (20 Myr) under predicts the observed heat flow. We find a good fit to the data using a geotherm appropriate for 10 Myr seafloor. The extra heat is interpreted in terms of back-arc thermal environments.
1. Introduction

Knowledge of the temperature distribution of accretionary prisms is important for understanding physical and chemical processes such as fluid flow, diagenesis, and the mechanics of faulting [e.g., Langseth et al., 1990; Yamano et al., 1992; Ruppel and Kinoshita, 2000; Booth-Rea et al., 2008]. Heat flow measurements provide the most direct measure of the thermal state of accretionary prisms. These measurements are most commonly based on either seafloor probe determinations that measure the in situ thermal gradient and thermal conductivity to depths of typically 4–5 mbsf, or borehole temperature gradients that extend much deeper into the sediment column, and thermal conductivities measured on returned core material. Because of their limited measurement depth seafloor probe measurements are sensitive to shallow processes. This effect is particularly manifest in dynamic environments such as the overriding margins of subduction zones where processes may significantly change the thermal regime over short temporal and spatial scales. These surficial variations are often diagnostic of near seafloor processes such as variations in bottom water temperature, shallow deformation, and fluid flow but often obfuscate the deeper thermal field. In contrast ocean drilling provides access to deeper environments where downhole tools, acoustic measurements, and logging technologies can provide important scientific insights into the deeper thermal regime. The best chance of obtaining a regional image of thermal processes throughout a subduction margin is through a combination of numerous shallow seafloor probe measurements that capture spatial variability and deeper but necessarily fewer borehole measurements that are more representative of deeper processes.

In this paper we report detailed thermal measurements undertaken during the Integrated Ocean Drilling Program (IODP) Expeditions 315 and 316 to understand the thermal regime of the Nankai accretionary prism offshore the Kii Peninsula. These boreholes provided access for temperature measurements to depths >100 mbsf and targeted the frontal thrust area, the outer fore-arc high where a large splay fault intersects the seafloor, and the Kumano Basin (Figure 1). Geothermal objectives included determining thermophysical rock properties of the cored material and characterizing the heat flow in this area. New thermal conductivity measurements made with a divided bar are reported and supplement shipboard full- and half-space measurements. Forty-nine sediment temperatures from 6 sites and close to 1000 thermal conductivity determinations were used to calculate heat flow along the NanTroSEIZE transect. These heat flow values are then adjusted for the thermal effects of bathymetric relief and sedimentation. We conclude with a simple thermal model of subduction along the transect.

2. Geologic Setting and Stratigraphy

The Nankai Trough and accretionary prism are the result of the Philippine Sea plate subducting beneath Japan. The accretionary prism forms the landward slope of the Nankai Trough and has evolved since the Miocene. The prism is as much as 100 km wide and consists of accreted and underplated trench fill and Shikoku Basin sediments. The accretionary prism has been divided into six structural zones: the trench, frontal thrust fault zone, imbricate thrust zone, megasplay fault zone, Kumano Basin edge fault zone, and the fore-arc basin [Moore et al., 2009].

During IODP Expeditions 315 and 316, temperature and thermal conductivity measurements were made in the region of the frontal thrust zone (Sites C0007 and C0006), the megasplay fault zone (Sites C0008, C0004, and C0001) and in the Kumano Fore-arc Basin (Site C0002). The topographic setting and stratigraphy for each region is briefly described to place the thermal data in context. Because clay and quartz content influence thermal conductivity we emphasize this aspect of the stratigraphy.
Sites C0007 and C0006 are within the frontal thrust zone on a topographic high that forms the northeast headwall of a landslide. C0007 is on the slope up to the high and C0006 is on the slope break (Figure 1). Sediments recovered at these sites reflect underthrust and accreted material overlain by slope deposits. The stratigraphy of Sites C0007 and C0006 is described by Kinoshita et al. [2009] and Screaton et al. [2009]. Four units were cored at Site C0007; the upper three of these were also cored at Site C0006 (Figure 2). Unit I comprises the slope deposits and is interpreted as being deposited by hemipelagic settling, turbidity currents, and soft sediment slumping on an oversteepened slope. It is composed of nannofossil-bearing mud, interbedded sand layers, and minor volcanic ash. Unit II is interpreted as accreted trench wedge material; it consists of fine-grained mud to sand- and gravel-rich deposits that coarsen upward. At the base of unit II a significant age transition from Pleistocene and younger sediments above to Miocene (2.5 Ma and older) below marks the boundary with the Shikoku Basin facies (unit III). Unit III consists of mud with ash and tuff layers, interpreted to have been deposited by hemipelagic settling in a deep marine basin. At Site C0007 the base of unit III was cored and is marked by evidence of a fault zone [Kinoshita et al., 2009]. Unit IV consists of sands and recovery was poor. At both sites, clay content increases with depth from ~40% to 60% and quartz content decreases somewhat with depth.
Sites C0008, C0004, and C0001 are in the megasplay fault zone, just southeast of a prominent ridge that forms the seaward boundary of the Kumano Basin (Figure 1). The stratigraphy of these sites is described by Kinoshita et al. [2009] and Kimura et al. [2011]. Sediments recovered from Site C0008 reflect upper slope basin deposits (Figure 2). Two units were identified. Unit I consists of silty clay with a substantial component of calcareous nannofossils, siliceous biogenic debris, and volcanic ash. Lower in the unit, interbedded mud clast gravels and silty clay beds dominate. Unit II consists of sand-rich turbidites with pebbly sandstone and minor silty clay beds dominate. Unit II consists of sand-rich turbidites with pebbly sandstone and minor silty clay beds dominate. Clay mineral content varies between 25 and 55%; quartz content is relatively uniform at ~20%. Site C0004 targeted the conspicuous splay fault at 300 mbsf and recovered accretionary prism sediments in the hanging wall of the splay fault, slope sediments that overlie the prism, and slope basin sediments in the footwall of the splay fault. Four units were identified. Unit I consists of hemipelagic nannofossil-rich mud with rare volcanic ashes and rare thin sands and is interpreted as slope sediment. In the uppermost portion of unit II sedimentary breccias and hemipelagic muds dominate while in the lower portion hemipelagic muds with rare turbidite sands dominate. These sediments represent mass wasting deposits and prism sediment, respectively. Unit III consists of ash-bearing hemipelagic muds with scattered volcanic ashes and unit IV consists of hemipelagic mud with turbidite sands and rare volcanic ashes. In general, clay content increases with depth from 40 to 65% through unit III and then decreases to about 45% in unit IV. Quartz content is relatively uniform at about 20%. Site C0001 documented the hanging wall of the megasplay and intersected two units. The upper section of unit I consists of nannofossil-bearing hemipelagic mud with volcanic ash, and the lower portion contains sand and hemipelagic mud and is interpreted as reflecting slope apron facies. Unit II is dominated by hemipelagic mud representing the upper accretionary prism. Clay minerals increase...
dramatically through unit I from about 30 to 60% and are relatively constant through unit II. Quartz content is relatively uniform at approximately 20%.

[8] Site C0002 is in Kumano Basin and targeted the sedimentary sequence of the upper basin (Figure 2). Four units were cored and consist of hemipelagic muds with fine sands and silt turbidites in the upper sections. Clay content shows small variations with depth and varies between about 30 to 65 with a significant decrease in unit IIIa. Quartz generally decreases with depth and varies between about 15 and 35%.

3. Thermal Conductivity

[9] To understand the thermal structure of the Nankai prism, it is important to have a good knowledge of the thermophysical rock properties. Of these properties thermal conductivity is of primary importance because it exerts a fundamental control on the distribution of temperatures. Thermal conductivity varies as a function of porosity and mineralogy and can vary by a factor of three or more from 0.9 W m\(^{-1}\) K\(^{-1}\) for porous clays to 4.5 W m\(^{-1}\) K\(^{-1}\) for lithified quartz-rich sandstones [e.g., Zoth and Haenel, 1988].

[10] We measured thermal conductivity using three different techniques. In the upper portions of Holes, sediments were generally unlithified and full-space measurements were made [Von Herzen and Maxwell, 1959]. In the deeper portions of Holes, where sediments were lithified and pieces were big enough (>5 cm), we used the half-space technique [Vacquier, 1985]. Both of these methods use a needle probe that contains a thermistor and heating wire that approximates a line source. In the full-space method the needle is inserted through a small hole in the core liner into the unlithified sediment, and in the half-space method the needle is insulated within an epoxy disk but exposed on one face. The face of the disk with the exposed needle is placed against the face of lithified material being measured. The needle is aligned parallel to the core axis. Supplementary measurements made perpendicular to this direction show no evidence of anisotropy in the lithified material. Both the full- and half-space methods are transient techniques that approximate the heating element as an infinite line source. The needle is heated with constant power and the temperature time response yields the thermal conductivity. The experimental uncertainty is estimated to be 5%. Many of the full- and half-space measurements were made in triplicate, and repeated measurements were usually within 1 to 2% of the mean. Both methods were calibrated with a rubber standard (0.517 W m\(^{-1}\) K\(^{-1}\)), and two glass standards (1.237 and 1.623 W m\(^{-1}\) K\(^{-1}\)). These calibrated values showed a small offset relative to the standard. Secondary calibration constants were determined based on a least squares inversion between these measured values and standards [Kinoshita et al., 2009]. These adjustments increased the measured values by approximately 4%.

[11] In many of the Holes, sections of returned core material were too lithified to insert a needle but so friable that pieces were too small to measure with the half-space disk. In these regions lithic fragments were collected and thermal conductivity was measured using a divided bar apparatus [e.g., Sass et al., 1971, 1984]. The divided bar is a steady state technique in which a heat flux is applied across a sample and the thermal gradient is measured. Samples are placed in a cylindrical cell and vacuum saturated with water to remove air. The bulk thermal conductivity, \(k_b\), of the cell is determined as the ratio of the applied heat flux. The grain thermal conductivity is determined from the geometric mean

\[
k_b = k_g^w k_g^{1-f}
\]

where \(k_w\) is the thermal conductivity of water (0.6 W m\(^{-1}\) K\(^{-1}\)), \(k_g\) is the grain conductivity, and \(f\) is the porosity of the cell. The bulk formation conductivity is then determined using equation (1) a second time by using the just determined grain conductivity and formation porosity. All divided bar measurements were made on a modified tandem apparatus at the University of Utah [Chapman et al., 1981] that was calibrated with both fused quartz and crystalline quartz. Experimental uncertainties for these measurements are estimated to be 3%.

[12] Eleven half-space and divided bar measurements were made on samples within 10 cm of each other. These measurements agree to 5%, although the standard deviation is approximately 13%. Comparisons between five full space and divided bar measurements have a mean agreement and standard deviation of 2 and 22%. The comparisons show that the methods generally give similar results with no systematic offsets.

[13] All thermal conductivity measurements were corrected for in situ pressure and temperature. Thermal conductivity increases with pressure; for each 1800 m of depth the thermal conductivity is
increased by 1% [Ratcliffe, 1960]. In porous rocks temperature influences thermal conductivity in two competing ways. The thermal conductivity of rock matrix is inversely related to temperature [Zoth and Haenel, 1988], whereas the thermal conductivity of water increases with temperature [Keenan et al., 1978]. We apply a temperature correction of +1% for each +20°C change in temperature between the laboratory and estimated in situ temperatures, a value intermediate between +5% suggested by Ratcliffe [1960] for a high-porosity, water-saturated sediment and mean value of −3% derived from data reported by Clark [1966] for several hard rocks. Thermal gradients determined at each site are discussed in section 4. These corrections increased values of thermal conductivity by 1–2%.

Thermal conductivity measurements are shown in Figure 2. In some Holes, scatter is greatest near the deepest full-space measurements (e.g., C0007 and C0004) or the shallowest half-space measurements (e.g., C0002). This scatter likely reflects applying the techniques to questionable material, either too lithified to get good contact with the needle for full-space measurements, or pieces too small to avoid edge effects with the half-space measurements. The generally greater scatter at very shallow depths may be due to low values attributed to gas expansion and/or core disturbance.

In general thermal conductivity varies between about 1.0 W m⁻¹ K⁻¹ near the seafloor to 1.6 W m⁻¹ K⁻¹ at depth. Measurements may be biased toward low values because of core disturbance and greater recovery of muds instead of sands relative to their in situ occurrence. Bulk thermal conductivity is inversely proportional to porosity (Figure 3). This correlation is most pronounced at sites C0007, C0004, C0001 and C0002 where variations in porosity are significant. Additionally, second-order variability is present that can be understood in terms of lithology. High values tend to be associated with sands and higher quartz content while low values tend to be associated with clays and ash layers. For example, the frontal thrust sites (C0007 and C0006) have a higher fraction of quartz content and are associated with generally high thermal conductivities. In other sections where muds or volcaniclastics dominate, thermal conductivity values are low (C0001 and C0002).

4. Temperature Measurements

Equilibrium temperatures were measured using the APCT-3 and SET temperature tools. The APCT-3 tool fits within the cutting shoe of the hydraulic piston core used for coring sediments. The inner core barrel is shot 9.5 m into sediments at the bottom of the hole, sufficiently deep to avoid the thermal influence of drilling. After the core barrel penetrates, it is decoupled from the drill string and left in place from 7 to 10 min. During this period a temperature-time series of the decaying heat pulse from the frictional of penetration is recorded. These temperatures are extrapolated to the equilibrium temperature using a tool response function [Heesemann et al., 2007]. The SET tool measures temperatures in stiffer sediments and requires a wireline trip. This tool extends 1.4 m below the drill bit and is held in place with weight applied by the driller.

Forty-eight high-quality in situ temperature measurements were made at seven different sites on Expeditions 315 and 316 (Figure 4a). Each temperature-depth profile is based on at least five equilibrium temperatures (Holes C0007 and C0004) and up to 10 equilibrium temperatures (Hole...
C0008C). The deepest equilibrium temperature is approximately 250 m below seafloor (mbsf) at Hole C0008A, but most Holes have temperatures extending to only 150 mbsf because of increasing sediment induration. Equilibrium temperatures are plotted as a function of depth and thermal resistance on a relative scale to avoid overlap (Figure 4a).

Subbottom temperatures can be expressed as [Bullard, 1939]

\[
T(z) = T_0 + q_o \sum_{i=1}^{n} \frac{\Delta z_i}{k(z)_i}
\]

where \(T_0\) is the bottom water temperature, \(q_o\) is the background heat flux, \(\sum \Delta z_i/k(z)_i\) is thermal resis-
\[ \Delta z_i \] is the depth spacing and \( k(z)_i \) is the thermal conductivity corresponding to that depth spacing. In general thermal resistances appear linear and are in general well fit by a constant gradient suggesting constant heat flow. The standard deviation of the fit to the thermal resistance is less than or equal to 0.5 m\(^2\) W\(^{-1}\) K\(^{-1}\) (Table 1). Residuals are plotted in Figure 4b and give a better sense of discrepancies between individual thermal resistances and best fitting lines. Holes C0008A and C0008C are notable because they are only separated by \( \sim 200 \) m. Temperatures in Hole C0008A are generally more noisy than those in Hole C0008C but the heat flow differs by only \( 5 \) mW m\(^{-2}\), or about 10\%, indicating internal consistency. The reason for the increased noise at Hole C0008A is unknown.

[18] Values of heat flow range from 60 to 36 mW m\(^{-2}\) (Table 1) and in general decrease with distance landward of the deformation front, consistent with subduction. The heat flow value at Hole C0006 appears anomalously low.

### 5. Corrections to Heat Flow Determinations

#### 5.1. Bathymetric Correction

[19] Bathymetry has the potential to distort the thermal field by increasing the thermal gradient under bathymetric lows and decreasing it under bathymetric highs. High-resolution multibeam bathymetry with a grid size of approximately 100 m [Ike et al., 2008] shows pronounced relief at the drilling sites (Figure 1). Our strategy for correcting the thermal field is based on techniques formulated by Blackwell et al. [1980]. For each borehole we construct a grid of seafloor temperatures based on three-dimensional bathymetry and CTD casts from the world databank. Seafloor temperatures are projected onto a horizontal plane defined at the top of the borehole using the best fitting thermal gradient. Finally this surface is fit with a Fourier series and upward (depth positive down) continued to generate the thermal perturbations created by the bathymetry. Because the best fitting gradient used in the projection step is also the unknown being solved for, this scheme is necessarily iterative. The bathymetrically induced gradients are subtracted from the observed gradient to generate the background thermal gradient.

[20] Heat flow values have been corrected for bathymetry and are reported in Table 1. Heat flow increases for Sites C0007, C0006, C0001, and C0002 consistent with being sited on local ridges and topographic highs (Figure 1). Heat flow slightly decreases for Sites C0008A and C0008C, and C0004 consistent with being sited in local topographic lows. The largest change is associated with Sites C0006 (+19\%) and C0001 (+16\%) that sit on a slope break and local seafloor high, respectively. These large changes show the importance of this correction in regions of large relief.

#### 5.2. Sedimentation Correction

[21] Sedimentation and subsidence can transiently depress surface heat until the deposited sediments are warmed to the background equilibrium temperature. Similarly, unroofing of sediments through erosion or mass wasting events can transiently increase the surface heat flow until the unroofed section of sediments is cooled to the background equilibrium temperature. Thrust faults, normal faults, biostratigraphic and paleostratigraphic age reversals, and porosity data indicate a combination of these mechanisms have likely perturbed the thermal field. The linearity of the temperature-depth profiles suggests that these processes do not dominate the current thermal field, although the vertical spacing of temperature measurements degrades resolution to these events. Here we focus...
on the most recent large events because these will have the biggest impact on the heat flow. Our sedimentation correction is based on the work by Hutchison [1985] as implemented in a one-dimensional finite difference algorithm [Wang and Davis, 1992] that includes the thermal effects of variable sedimentation rates, compaction, and different thermal properties between the sediment and basement rocks. This correction is most sensitive to the combination of sediment thickness and its rate of deposition [Hutchison, 1985; Wang and Davis, 1992; Martin et al., 2004].

[22] Sedimentation rates and intervals of erosion at these sites, particularly in the megasplay fault zone is episodic due to irregular periods of fault and structural activity, as well as slumping and mass transport deposition [Strasser et al., 2009, 2011]. Sediment accumulation rates were calculated from biostratigraphic and paleomagnetic data documenting sedi-
We convert accumulation rates, \( S \), sites C0001, C0004, and C0008 have been subject to recent erosion, with surface hiatuses at 0.30, 0.44, and 0.06 Ma [Kinoshita et al., 2009]. These erosive events are likely due to slumping and mass wasting [Strasser et al., 2011]. Corrections for slumping to marine heat flow data are seldom applied because heat flow observations mostly try to avoid these areas. Here we assess the impact of these surface hiatuses on heat flow by modeling the thermal impact in terms of an instantaneous step change in seafloor temperature, \( T(z, t) \), expressed as

\[
T(z, \tau) = \Delta T \text{erf}
\left( \frac{z}{\sqrt{4\alpha \tau}} \right)
\]  

where \( z \) is depth, \( \tau \) is time before present that the event occurred, \( \Delta T \) is the difference between the original seafloor temperature and the new seafloor temperature due to unroofing, \( \text{erf} \) is the error function, and \( \alpha \) is thermal diffusivity. Equation (5) shows that the impact of slumping events depends on the magnitude of unroofing, through \( \Delta T \) and \( z \), and the time of the event in the past, \( \tau \). We compute percent changes in heat flow as a function of unroofing thickness and time since the event (Figure 6). For these perturbations we used a background thermal gradient of 0.06°C/m, a thermal conductivity of 1 W m\(^{-1}\) K\(^{-1}\). Near subbottom perturbations are much larger than indicated by the figure; we fit the temperature profile in a least squares sense over a depth range of 0 to 100 m to assess its impact on our heat flow determinations. For this set of parameters, largest perturbations correspond to unroofing events between 10 and 100 years before present. After approximately 10,000 years the thermal effects have attenuated. Extending the depth range of the calculation extends the time period over which the gradient is sensitive to erosive events.

Table 2. Sedimentation Correction

<table>
<thead>
<tr>
<th>Site</th>
<th>Initial Porosity</th>
<th>Compaction Constant (km)</th>
<th>Duration ( a ) (Myr)</th>
<th>Accumulation Rate (m/Myr)</th>
<th>Duration ( b ) (Myr)</th>
<th>Accumulation Rate (m/Myr)</th>
<th>Duration ( c ) (Myr)</th>
<th>Accumulation Rate (m/Myr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>C0007</td>
<td>0.50</td>
<td>1.06</td>
<td>0.5</td>
<td>24</td>
<td>1.0</td>
<td>480</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>C0006</td>
<td>0.50</td>
<td>1.85</td>
<td>0.6</td>
<td>40</td>
<td>1.2</td>
<td>150</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>C0008A</td>
<td>0.61</td>
<td>1.09</td>
<td>1.8</td>
<td>27</td>
<td>0.5</td>
<td>520</td>
<td>1.7</td>
<td>35</td>
</tr>
<tr>
<td>C0008C</td>
<td>0.61</td>
<td>1.09</td>
<td>1.8</td>
<td>27</td>
<td>0.5</td>
<td>520</td>
<td>1.7</td>
<td>35</td>
</tr>
<tr>
<td>C0004</td>
<td>0.61</td>
<td>1.09</td>
<td>2.5</td>
<td>30</td>
<td>2.0</td>
<td>140</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>C0001</td>
<td>0.63</td>
<td>1.99</td>
<td>2.0</td>
<td>100</td>
<td>2.0</td>
<td>13</td>
<td>2.0</td>
<td>320</td>
</tr>
<tr>
<td>C0002</td>
<td>0.62</td>
<td>2.07</td>
<td>1.0</td>
<td>140</td>
<td>1.0</td>
<td>1300</td>
<td>5.0</td>
<td>140</td>
</tr>
</tbody>
</table>

\( a \)Duration of sedimentation beginning with the most recent phase.

Sedimentation accumulation [Kinoshita et al., 2009]. Representative accumulation curves for each of the drilling areas are shown in Figure 5. We have fit the accumulation rate data with two or three linear trends and the durations and rates are given in Table 2.

We convert accumulation rates, \( V_o \), to sedimentation rates, \( V_s \), through [Hutchison, 1985]

\[
V_s = \frac{V_o (1 - \phi_o)}{(1 - \phi(z))}
\]

where \( f_o \) is the initial porosity and \( f \) is the porosity as a function of depth, \( z \). Porosity curves were fit to porosity data at each borehole using Athy’s law

\[
\phi(z) = \phi_o \exp\left(-\frac{z}{L}\right)
\]

where \( L \) is an empirically derived constant related to compaction. Porosity data for each region (Figure 5) are best fit in a least squares sense and best fitting parameters are reported in Table 2. In general the best fitting porosity trends fall in a small range with the exception of sites in the frontal thrust region. The initial porosity associated with C0007 and C0006 is 50% and suggest that this material undergone compaction by material that has been removed [Screaton et al., 2009]. At the other sites porosity does not decrease to <50% until 150 or 200 mbsf in the slope region or Kumano Basin, respectively. The erosion or removal of material would increase the surface heat flow, but because heat flow values at these sites appear relatively low, the transient effects of erosion may have diminished.

Slope sites C0001, C0004, and C0008 have been subject to recent erosion, with surface hiatuses at 0.30, 0.44, and 0.06 Ma [Kinoshita et al., 2009]. These erosive events are likely due to slumping and mass wasting [Strasser et al., 2011]. Corrections for slumping to marine heat flow data are seldom applied because heat flow observations mostly try to avoid these areas. Here we assess the impact of
prism slope vary between approximately 5 and 10%. Site C0002 within the Kumano Basin has a sedimentation correction of 22% consistent with rapid Pleistocene sedimentation.

6. Discussion of Heat Flow Observations

[27] Figure 7 shows the new corrected heat flow results in the context of the uncorrected measurements and existing data along the NanTroSEIZE transect. Including both the bathymetric and sedimentation correction leaves the heat flow at Site C0006 anomalously low. The source of this low value is not known. It may be that bottom water infiltrated the lithology during the temperature measurement, although temperature-time series of each measurement appear reasonable [Expedition 316 Scientists, 2009]. Neglecting the anomalously low value at Site C0006, the borehole values of heat flow generally decrease with distance from the deformation front consistent with the downward advection of heat by the subducting plate.
It is interesting to compare nearby seafloor measurements of heat flow with the borehole values of heat flow. We project existing heat flow from shallow probe (4.5–6 m) measurements (H. Hamamoto et al., Heat flow distribution and thermal structure of the Nankai Subduction zone off the Ki-I Peninsula, submitted to Geochemistry Geophysics Geosystems, 2010) onto the profile from distances as great as 10 km. The majority of these stations are within 2 km of the transect and lie near sites C0004 and C0008 (Figure 1). We focus on these data and for comparison we have also corrected these data for the effects of bathymetry using an assumed maximum measurement depth of 5 m. Reasonable variations in this parameter have a negligible influence on the corrected heat flow value. With the exception of two values at a distance of about 8 km from the deformation front, this correction decreases observed heat flow values on average by less than 5%. These other two values sit on a narrow trough and have an exceptionally large correction that decreases their magnitude by about 40%. It is interesting to note that the bathymetric correction for borehole heat flow systematically increase values, whereas for shallow probe data, values are systematically decreased. This may be due to placing probe measurements in basins or trough where the probe is more likely to penetrate sediments, and placing boreholes in basement highs or slopes. We have not corrected probe heat flow values for the effects of sedimentation but based on the correction at sites C0004 and C0008, the effect would increase these values by about 10%. Heat flow values for the shallow probe data are based on an average of in situ thermal conductivity determinations from the shallow probe measurements of 1.04 W m\(^{-1}\) K\(^{-1}\). This value is nearly identical to thermal conductivity measurements at C0004 and C0008 that are less than 10 mbsf, indicating that the difference in heat flow is due to the difference in thermal gradients. The shallow probe measurements have a mean and standard deviation of 73 and 11 mW m\(^{-2}\), respectively, 15 mW m\(^{-2}\) greater than the borehole heat flow measurements that have a mean and standard deviation of 58 and 8 mW m\(^{-2}\), respectively, excluding Site C0006.

It is difficult to unequivocally attribute the discrepancy between these two data sets to a particular cause or process because downhole measurements and shallow probe measurements have uncertainties that are both unique and common. Accretionary prisms are dynamic environments and discrepancies may be related to real issues of spatial heterogeneity. However, because these three borehole measurements and nine shallow probe measurements are systematically offset from each other it seems reasonable to assume that this reflects curvature in the temperature field of the subbottom. Two processes not already considered that could produce this curvature include the upward advection of heat by fluids and bottom water temperature variations.

The upward flux of fluids could cause the appropriate curvature. Low chloride cold seeps and chemosynthetic communities are observed southeast of the transect where splay faults intersect the seafloor [Toki et al., 2004]. Goto et al. [2008] measured heat flow in a biological community and found heat flow in excess of 180 mW m\(^{-2}\), much higher than the seafloor heat flow measurements near the transect. We solve the one-dimensional steady state advection-diffusion equation subject to a constant seafloor temperature and constant gradient at depth and express the mean vertical fluid velocity, \(V_z\), as a function of the surface heat flow, \(q_o\) [e.g., Harris et al., 2010]

\[
V_z = \frac{k_f}{\phi\rho_w c_w H} \ln \left( \frac{k_f \Gamma}{q_o} \right)
\]  

where \(f\) is the porosity, \(r_{w} c_{w}\) is the heat capacity of water, \(H\) is a length scale, and \(G\) is the thermal gradient at depth. An upward fluid flux of approximately 50 mm/yr would be required to increase the subbottom heat flow by 15 mW m\(^{-2}\). However, pore water analysis made on returned core material is consistent with in situ diagenetic reactions and there is no clear evidence of fluid flow along the decollement and associated splay faults at the sites drilled during Expeditions 315 and 316 [Solomon et al., 2008; Wheat et al., 2008]. These results suggest that advective fluid flow is unlikely responsible for the difference between seafloor and deeper values of heat flow in this area.

Bottom water temperature variations can perturb shallow thermal gradients and has been suggested as an explanation of the scatter in heat flow on the accretionary prism slope (H. Hamamoto et al., submitted manuscript, 2010). We summarize their argument and explore its implications for reconciling the seafloor probe measurements with borehole measurements of heat flow. The Nankai Trough is affected by the Kuroshio current, a strong western boundary current that imparts bottom water temperature variations to the shallow subbottom [Yamano et al., 1984; Hamamoto et al., 2008]. The seafloor probe measurements were made at bathy-
metric depths that range between approximately 2700 to 3050 m. Although there are no long-term records of bottom water temperature variations at these depths, bottom water temperatures at a depth of 2525 m were recorded over a two and half year period and shows annual peak-to-peak temperature variations of about 0.3°C \cite{Hamamoto et al., 2008}. This variation can produce a maximum temperature change of 0.115°C at 5 mbsf; corresponding to an apparent change in heat flow of about 23 mW m\(^{-2}\). The majority of these measurements, six out of nine, have a penetration of only 2 m \cite{Hamamoto et al., submitted manuscript, 2010} that can lead to greater perturbation than that calculated above. There is an additional clue that the discrepancy between these sets of data may be due to bottom water temperature variations. The shallow probe data come from three cruises (Figure 7). We note that cruise KT-05 and KT-06 form two distinct clusters suggesting that the heat flow values may in part a function of cruise timing. The means of these two data sets are 63 and 82 mW m\(^{-2}\), a difference of 19 mW m\(^{-2}\). The standard deviations of the two subsets are 4 and 11 mW m\(^{-2}\), respectively, values less than the difference between their means. Two additional values are from a third cruise KR-04 overlap both of the clusters. Together this evidence suggests that part of the variation in heat flow is a function of cruise date and may therefore be due to variations in bottom water temperature. Cruise KT-05 was 25 February through 3 March 2005, and cruise KT-06 was later in the spring 11 April through 22 April 2006 \cite{M. Yamano, written communication, 2011}. It is interesting to note that heat flow at Site C0007 lies on a trend between the heat flow values collected on cruises KR-02 and KR-04. These data at greater water depths (>3 km) suggest that significant bottom water temperature variations may not extend to these depths.

Together these results suggest that heat flow on the accretionary prism along the NanTroSEIZE transect may be lower than previously estimated for two reasons. First bathymetric corrections generally lower seafloor probe values of heat flow. This may reflect a bias toward placing these measurements in troughs where there is a greater likelihood of probe penetration. Second, because the borehole temperature measurements come from deeper in the prism and are therefore much less susceptible to bottom water temperature variations, we suggest that heat flow determinations for the borehole sites are more reflective of the thermal state of the accretionary prism.

### 7. Thermal Models

The thermal state of the shallow subduction is primarily a function its convergence rate and dip, the thermal properties of the margin, and importantly the temperature structure of the incoming plate \cite[e.g.,][]{Cloos, 1985; Dumitru, 1991}. We use a transient 2-D finite element model \cite{Wang et al., 1995} to simulate temperatures in a cross section along the NanTroSEIZE transect. Heat is transferred advectively with the subducting plate and conductively through the fore arc. The Philippine Sea plate is subducting at a rate of 4 cm yr\(^{-1}\) \cite{Seno et al., 1993}. The geometry of the model is derived from seismic reflection data for the shallow portion of the subduction zone \cite{Moore et al., 2009} and tomographic imaging for the deeper slab geometry \cite{Hirose et al., 2008}. We parameterize the geometry of the model in terms of 5 units (Table 3). Model parameters are based on a combination of NanTroSEIZE drilling results \cite{Kinoshita et al., 2009} and parameters used along the Muroto transect \cite{Hyndman et al., 1995; Wang et al., 1995; Spinelli and Wang, 2008}.

The upper boundary for the model has a fixed temperature of 0°C, and the base of the subducting plate is assigned a temperature of 1400°C. The thermal regime of the subduction thrust is insensitive to this basal boundary condition. The landward boundary is also placed sufficiently far from the seismogenic zone to avoid boundary effects. At the landward boundary, we prescribe a surface heat flow of 70 mW m\(^{-2}\) and an adiabatic gradient through the mantle wedge. The thermal effects of mantle wedge flow are coupled to the subducting slab using an isoviscous mantle rheology \cite{Peacock and Wang, 1999}. The details of the mantle rheology have only a small effect on the seismogenic
portion of the subduction thrust [Currie et al., 2002].

[35] The temperature structure of the incoming Philippine Sea plate plays an important role in the model outcome but also introduces significant uncertainty. There are at least two factors that increase this uncertainty here: (1) the tectonic history is transient and may impact the temperature field [Wang et al., 1995], and (2) the Philippine Sea plate was formed as a back arc to the Izu-Bonin subduction zone and the thermal structure of back arcs are not well known in detail [Peacock, 2003; Currie and Hyndman, 2006]. Both of these factors are discussed below.

7.1. Tectonic History of Subduction

[36] The Philippine Sea plate formed as a back-arc basin to Pacific plate subduction at the Izu-Bonin Trench (Figure 1, inset). Back-arc spreading originated along a now fossil spreading center located offshore Cape Muroto and marked by the Kinan seamount chain [Kobayashi and Nakada, 1978; Chamot-Rooke et al., 1987; Okino et al., 1994]. Sdrolias et al. [2004] analyzed the tectonic history of the Shikoku Basin based on magnetic, gravity, and bathymetric and the salient features are briefly summarized here. The Shikoku Basin originated at approximately 28 Ma due to a propagating rift in response to slab rollback of the Izu-Bonin Trench. At this time the Izu-Bonin Trench was south of 22°N and old (>75 Ma) Pacific plate was subducting at the position of the Kii Peninsula. Contemporaneously with slab roll back of the Pacific plate, the Izu-Bonin Trench migrated to the northeast such that the trench was approximately at the position of the Kii Peninsula transect at 14 Ma. Rifting of the Shikoku Basin occurred between 28 and 15 Ma. Since about 14 Ma, when the Izu-Bonin Trench migrated past the position of the Kii Peninsula, the Nankai Trough has subducted back-arc material. Currently back-arc seafloor with a magnetic age of 20 Ma is being subducted.

7.2. Regional Heat Flow and Thermal Structure of Shikoku Basin

[37] Currie and Hyndman [2006] surveyed observations of back-arc basins that bear on their thermal state and found that nearly all back-arc basins are anomalously warm even in the absence of recent extension. They also found that the thermal anomaly extended over wide areas of the back arc (250 to >900 km). Thus we expect the thermal state of the incoming Shikoku Basin may be anomalously warm.

[38] The Nankai Trough and Shikoku Basin have been the focus of heat flow studies for decades [e.g., Watanabe et al., 1970]. Because oceanic heat flow is most often understood in terms of conductive cooling models as a function of age, Yamano et al. [1992, 2003] viewed heat flow perpendicular to the fossil spreading center. Heat flow is a maximum near the fossil spreading ridge offshore Cape Muroto (Figure 8). Although heat flow generally decreases with distance from this fossil spreading center both westward toward Cape Ashizuri, and eastward toward the Kii Peninsula, it is significantly higher than conductive cooling curves based on seafloor age [Kinoshita and Yamano, 1995; Kinoshita et al., 2008]. Given the scatter in the data, the heat flow can also be approximated by a constant value. Offshore the Kii Peninsula, and seaward of the Nankai Trough, seafloor values of heat flow are about 110 mW m⁻². The cooling effect of sedimentation is thought to transiently depress this value by about 20%, putting the equilibrium heat flow at about 130 mW m⁻². Using half-space cooling [Davis and Lister, 1974], \( q = C/\sqrt{\text{Age}} \), with a value of \( C \sim 450 \text{ mW m}^{-2}/\text{Ma}^{1/2} \) yields a thermal age of 10 Ma.
Our model focuses on the shallow portion of the subduction zone where we can best estimate its thermal state, and we explore two models that bracket the thermal state of the Philippine Sea plate offshore the Kii Peninsula. In both models we assume that mantle dynamics have reset the geotherm of the back-arc crust to a maximum thermal age. We depress the geotherm through the sedimentary section by adding sediments. Once the crust subducts it is out of the back arc and ages relative to the reset thermal age. With this assumption, steady state thermal models of subduction can be employed, at least for the shallow portion of the subduction zone. Typical conductive time constants are calculated from the expression $\exp(-z^{2}/4\alpha t)$, where $z$ is depth, $\alpha$ is thermal diffusivity ($32 \text{ km}^2/\text{Myr}$), and $t$ is time. If the modeled configuration has been approximately steady for the past 5 Ma then temperatures in the prism to a depth of 25 km are $\sim$83% of equilibrium. In the first model (model A) we use the age of the presently subducting oceanic crust so that the thermal regime of the crust at the Nankai Trough is equivalent to 20 Ma oceanic lithosphere. In model B the

![Figure 9. Thermal model for NanTroSEIZE transect. (a) Heat flow values and model profiles for the two models considered. (b and c) Finite element model showing units. Yellow corresponds to subducting sediment, green corresponds to accreted sediment, and brown corresponds to oceanic crust. Black lines show plate boundary (lower) and splay fault (upper). White lines show finite element model structure. Isotherms corresponding to models B and A are shown in Figures 9b and 9c, respectively.](image-url)
age of the oceanic crust at the Nankai Trough is equivalent to 10 Ma oceanic lithosphere. Results are shown in Figure 9. Model A is relatively cool and fits the lowest values of heat flow. In contrast, model B is warmer and fits the heat flow at C0007 and the trend in probe values between 10 and 20 km landward of the trench. The model is only marginally warmer than the borehole heat flow on the upper accretionary prism. In model A the subduction thrust intersects the 100°C and 150°C isotherms at approximately 22 and 48 km, respectively. In model B the subduction thrust intersects the 100°C and 150°C isotherm at approximately 15 and 25 km landward of the deformation front, respectively. Thus both models predict dehydration reactions landward of the megasplay fault zone.

These two models have implications for predicted temperatures along the megasplay. Mechanical models have been used to argue that the megasplay fault is the primary coseismic plate boundary near the updip extent of slip [Wang and Hu, 2006]. The splay fault branches upward from the plate boundary where the temperature is approximately 150°C and 200°C in models A and B, respectively. If the updip extent of seismongenesis is limited by the 150°C isotherm [Hyndman et al., 1995], then model A predicts the megasplay fault is largely aseismic whereas model B is consistent with seismonogenesis on the megasplay fault. Future ocean drilling in the Kumano Basin will test these predictions.

8. Conclusions

On the basis of this study we conclude the following:

1. There is good agreement between full- and half-space measurements of thermal conductivity with those made by the divided bar method. Together these measurements characterize the thermal conductivity of sediments encountered during drilling. Thermal conductivity increases with depth from about 1.0 W m⁻¹ K⁻¹ to 1.6 W m⁻¹ K⁻¹ and is inversely proportional to porosity. Other variations are related to the clay and quartz content of the material.

2. Along the NanTroSEIZE transect thermal corrections for the bathymetry increase heat flow values by an average of 3% but are as large as 16% (C0001). Corrections for sedimentation increases heat flow on average by about 10% but in the Kumano Basin is as large as 22%. In general these corrections increase heat flow by about 12% and heat flow varies from 70 mW m⁻² near the toe of the deformation front to 57 mW m⁻² at the Kumano Basin.

3. Heat flow measurements from IODP sites are less than seafloor values of heat flow in the area of the megasplay. This discrepancy likely reflects the influence of bottom water temperature variations.

4. Steady state thermal models of subduction are consistent with the heat flow data when an effective age of 10 Ma for the incoming Philippine Sea plate is used.

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