Thermal models of the Mexico subduction zone: Implications for the megathrust seismogenic zone

C. A. Currie
School of Earth and Ocean Sciences, University of Victoria, Victoria, British Columbia, Canada

R. D. Hyndman and K. Wang
Pacific Geoscience Centre, Geological Survey of Canada, Sidney, British Columbia, Canada

V. Kostoglodov
Instituto de Geofisica, Universidad Nacional Autonoma de Mexico, Mexico City, Mexico

Received 2 August 2001; revised 8 August 2002; accepted 23 September 2002; published 26 December 2002.

[1] It has been proposed that the seismogenic zone of subduction thrust faults is primarily controlled by temperature or rock composition changes. We have developed numerical models of the thermal structure of the Mexico subduction zone to examine the factors that affect the temperature of the subduction thrust fault. Although the oceanic plates subducting beneath Mexico are young, the top of the oceanic plate at the trench is cool, because of the lack of a thick cover of insulating sediments. Marine heat flow observations suggest that hydrothermal circulation may further cool the oceanic plate. This results in a cool subduction thrust fault, where the brittle part of the fault extends to depths of over 40 km. At these depths, even slight frictional heating may have significant effects on temperature along the thrust fault, particularly for regions with a high convergence rate and shallow plate dip. With the addition of a small amount of frictional heating, the temperatures of the deep (30–40 km) thrust fault are increased by over 200°C. As the observed downdip limit of rupture in recent well-constrained megathrust earthquakes is confined to depths above the intersection of the thrust fault and the continental Moho, a temperature of 350°C may control the downdip extent of the seismogenic zone. Thus, in order to be consistent with the observed shallow rupture areas, it is necessary to include a small amount of frictional heating, corresponding to an average shear stress of 15 MPa.

INDEX TERMS: 3015 Marine Geology and Geophysics: Heat flow (benthic) and hydrothermal processes; 3040 Marine Geology and Geophysics: Plate tectonics (8150, 8155, 8157, 8158); 7230 Seismology: Seismicity and seismotectonics; 8130 Tectonophysics: Evolution of the Earth: Heat generation and transport; KEYWORDS: thermal models, subduction zone, megathrust earthquakes, seismogenic zone, heat flow, Mexico subduction zone


1. Introduction

[2] The Mexico subduction zone has experienced numerous megathrust earthquakes over the last century. Some of the largest earthquakes in recent history are the 1985 MW 8.1 Michoacan earthquake and the 1995 MW 8.0 earthquake in the Jalisco region. Such earthquakes pose a significant seismic hazard to the coastal regions of Mexico, as well as areas considerably inland, including Mexico City.

[3] Of particular importance to seismic hazard studies are the seaward (updip) and landward (downdip) limits of the seismogenic zone of the subduction thrust fault. These limits define the maximum earthquake rupture width, which is related to the maximum magnitude that may be expected for the fault. The updip limit is the closest approach of the seismic source zone to cities, and thus is important for ground shaking hazard. The location of the updip limit is important for tsunami generation.

[4] It is hypothesized that temperature and rock composition provide the primary controls on the width and location of the seismogenic zone [e.g., Zhang and Schwartz, 1992; Tichelaar and Ruff, 1993; Hyndman and Wang, 1993; Hyndman et al., 1997; Peacock and Hyndman, 1999]. In this study, we have developed thermal models of the Mexico subduction zone. We examine the factors that influence the temperature along the subduction thrust fault...
and then compare the thermal models to the observed rupture areas of recent megathrust earthquakes.

### 2. Constraints on Megathrust Earthquake Rupture Widths

[5] Globally, it is observed that only a shallow portion of the subduction fault is seismogenic [e.g., Zhang and Schwartz, 1992; Tichelaar and Ruff, 1993]. The seismogenic zone is bounded at its updip and downdip limits by regions that exhibit stable (aseismic) sliding (Figure 1). The transition from stable sliding to stick slip marks a change in fault behavior from velocity strengthening to velocity weakening [e.g., Scholz, 1990]. Although a number of factors may control fault behavior, we focus on temperature and composition.

#### 2.1. Temperature Limits on Seismogenic Zone

[6] The presence of an updip aseismic zone has been attributed to the presence of stable sliding sediments [Byrne et al., 1988; Vrolijk, 1990]. There are many factors that affect sediment properties, including pore fluid pressure and physical and chemical changes in the sediments [e.g., Moore and Saffer, 2001]. One change that may be important is the dehydration of smectite clays to illite and chlorite at temperatures between 100 and 150°C [Wang, 1980]. Whether this change marks the updip limit of the seismogenic zone remains a question [Marone et al., 2001], and the thermal control on the frictional behavior of sediments along the subduction fault is a subject of active research.

[7] The downdip limit of the seismogenic zone is proposed to be limited by a temperature of 350°C for some subduction zones. Laboratory experiments show that quartz-feldspathic continental rocks exhibit a transition from velocity weakening to velocity strengthening at temperatures of 325–350°C [Tse and Rice, 1986; Blanpied et al., 1995]. This temperature agrees well with the maximum depth of earthquakes within continental crust [Brace and Byerlee, 1970; Chen and Molnar, 1983; Tse and Rice, 1986]. There may be a second temperature which limits the maximum rupture depth of megathrust earthquakes, initiated at temperatures less than 350°C. This second limit is proposed to be at about 450°C [Hyndman and Wang, 1993], corresponding to a rapid increase of instantaneous frictional stress in laboratory data [Tse and Rice, 1986].

[8] For southwest Japan, Hyndman et al. [1995] showed that the proposed thermal limits are consistent with both the coseismic rupture width and the seismogenic zone determined through modeling of coseismic and interseismic crustal deformation. For the Cascadia subduction zone, the proposed downdip thermal limit is consistent with interseismic geodetic observations [Hyndman and Wang, 1993; Dragert et al., 1994; Wang et al., 2003]. Both southwest Japan and Cascadia have young subducting plates.

#### 2.2. Alternative Downdip Limit: Serpentinized Forearc Mantle Wedge

[9] For subduction zones with older subducting plates, such as Chile and south Alaska, Oleskevich et al. [1999] showed that the critical maximum temperatures are reached at depths greater than the intersection of the thrust fault with the continental Moho, whereas the maximum depth of rupture is limited to depths of 40–50 km [Tichelaar and Ruff, 1993]. These depths correspond to the depth of the continental Moho, and thus are consistent with the hypothesis that the intersection of the continental Moho and thrust fault provides the maximum downdip limit to the seismogenic zone [Ruff and Tichelaar, 1996]. One possible mechanism for generating stable sliding behavior of the fault below the continental Moho intersection is serpentinization of the mantle wedge. Dehydration reactions within the subducting plate release water into the overlying forearc mantle wedge, resulting in the formation of serpentine minerals and possibly other hydrous minerals, such as talc and brucite [Peacock and Hyndman, 1999]. Laboratory studies indicate that serpentine generally exhibits stable sliding behavior [e.g., Reinen, 2000]. Although there are no laboratory studies of the sliding behavior of talc and brucite, their layered structure suggests that they are probably weak and aseismic. The existence of serpentine within the mantle wedge is supported by thermal and petrologic models [e.g., Peacock, 1993], as well as a number of seismological observations [e.g., Suyehiro et al., 1996; Kamiya and Kobayashi, 2000; Bostock et al., 2002].

### 3. Thermal Modeling of the Mexico Subduction Zone

#### 3.1. Numerical Approach

[10] In this study, the proposed controls on the megathrust seismogenic zone are examined through thermal modeling of the Mexico subduction zone, from the northern Rivera plate (~22°N) to the Tehuantepec Ridge on the Cocos plate (~15°N) (Figure 2). We have developed models for four profiles oriented perpendicular to the Middle America Trench. For each profile, a two-dimensional, steady state thermal model has been developed using the finite element approach described by Wang et al. [1995b]. We have also included subduction-induced mantle wedge flow, following Peacock and Wang [1999]. The thermal effects of mantle wedge flow on the shallow thrust fault are very small. The thrust temperature is increased by a maximum of 6°C at 40 km depth with the introduction of the flow.

[11] The critical parameters for thermal modeling are the geometry and age of the oceanic plate, the thickness and...
deposition history of incoming sediments, and the convergence rate. The thermal conductivity and radioactive heat generation of each rock unit must also be assigned. Two additional factors of potentially first-order importance are frictional heating along the thrust fault and hydrothermal circulation within the upper incoming oceanic crust.

Also of potential importance is the thickness of the continental crust, as this defines the approximate location of the Moho intersection with the thrust fault. The crustal thickness beneath Mexico has been inferred from body and surface wave studies, seismic refraction surveys, and modeling of gravity and magnetotelluric data. In the northern part of the study region, Gomberg et al. [1989] and Bandy et al. [1999, and references therein] propose a crustal thickness of 36 to 40 km. To the south, crustal thickness studies give values of 33 km [Couch and Woodcock, 1981], 44 km [Arzate et al., 1993], 45 ± 4 km [Valdes et al., 1986], and 50 km [Helsley et al., 1975]. In our modeling, we use a crustal thickness of 40 km for all profiles.

The upper boundary of the model has a fixed temperature of 0°C. The base of the model is assigned a temperature of 1450°C, which approximates the mantle temperature at a depth of ~100 km. Because this boundary is located far from the region of interest (shallow subduction thrust fault) and because the primary control on the thermal regime is the seaward boundary condition, the basal boundary condition does not affect our results.

### 3.2. Oceanic Geotherm

The seaward boundary condition for the two-dimensional model is a one-dimensional geotherm for the oceanic plate. There are several factors that control the temperature-depth profile of an oceanic plate: (1) conductive cooling of the plate as it ages, (2) the deposition of lower conductivity sediments on top of the plate that slow the rate of cooling, (3) the increase in sediment thickness over time, resulting in a colder sediment column, (4) compaction of sediments with increasing sedimentation, resulting in the expulsion of pore water, causing advective heat transfer within the sediments, and (5) hydrothermal circulation in the upper oceanic crust. In the following, we neglect the effects of hydrothermal circulation.

The Rivera and Cocos plates are both young oceanic plates. Magnetic anomaly lineations [Klitgord and Mannerickx, 1982] indicate a slight increase in plate age to the southeast, from 11.5 Ma at the Jalisco profile to 15.5 Ma at the Oaxaca profile (Table 1). Multichannel seismic reflection data show that the northern Cocos plate and the Rivera plate are covered with no more than 200 m of sediment [Michaud et al., 2000]. Seaward of the trench, the Cocos plate between the Guerrero and Oaxaca cross sections is covered with 170–200 m of pelagic and hemipelagic sediments.

<table>
<thead>
<tr>
<th>Profile</th>
<th>Plate Age at Trench, Ma</th>
<th>Margin-Normal Convergence Rate, mm yr⁻¹</th>
<th>Plate Dip at 15 km Depth, deg</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jalisco</td>
<td>11.5</td>
<td>38</td>
<td>17.0</td>
</tr>
<tr>
<td>Michoacan</td>
<td>13.3</td>
<td>55</td>
<td>15.3</td>
</tr>
<tr>
<td>Guerrero A</td>
<td>13.1</td>
<td>58</td>
<td>15.3</td>
</tr>
<tr>
<td>Guerrero B</td>
<td>13.1</td>
<td>58</td>
<td>13.1</td>
</tr>
<tr>
<td>Oaxaca</td>
<td>15.5</td>
<td>61</td>
<td>11.3</td>
</tr>
</tbody>
</table>
A uniform sediment thickness of 200 m on the incoming oceanic plate was used for all model profiles. Sediment deposition has occurred at a rate of 135 m Myr$^{-1}$C0 over the last 0.78 Ma, and at a rate of 3–30 m Myr$^{-1}$C0 before that [Shephard and McMillen, 1981]. Because of the small sediment thickness, uncertainties in the thickness and deposition rate have little effect on the oceanic crust thermal structure.

Following the approach of Wang and Davis [1992], the oceanic geotherm can be calculated by allowing the plate to cool from zero years to its age at the trench, using the time-dependent sedimentation history and assuming a constant porosity-depth profile of the sediment column [Hutchison, 1985]. The resulting oceanic geotherms for each profile are shown on Figure 3. The Oaxaca geotherm has a slightly shallower gradient, due to the greater plate age. Despite the young age of the oceanic plates, the Mexico subduction zone is fairly cool relative to the Cascadia and southwest Japan subduction zones [e.g., Wang et al., 1995a, 1995b]. However, the oceanic plates in both of these regions are covered with 1.5 to 3.5 km of sediment, compared to $\sim$200 m for the Mexico margin. The insulating effect of the thicker sediments results in a warmer subducting plate. At the trench, the temperature of the top of the oceanic plate is more than 200°C for the Cascadia subduction zone [e.g., Hyndman and Wang, 1993], whereas the temperature is 30 to 50°C for Mexico. The major uncertainty in the oceanic geotherms is the possible cooling of the upper oceanic plate by hydrothermal circulation (see discussion below).

### 3.3. Oceanic Plate Geometry

For each profile, the geometry of the oceanic plate was defined using (1) single-channel and multichannel seismic reflection data, (2) seismic refraction data, (3) Wadati-Benioff earthquakes (assumed to occur about 5 km below the top of the oceanic plate), (4) intermediate magnitude thrust earthquakes, and (5) large megathrust earthquakes and their aftershocks. Data within 50 km of each profile were projected onto the cross section, and a best fit line was determined using a low-order polynomial (Figure 4). This geometry is similar to that given by Pardo and Suarez [1993, 1995]. The Jalisco profile also agrees well with the crustal structure determined through gravity modeling [Bandy et al., 1999]. Estimated uncertainties in plate depth are 10%.

Between the Orozco and O’Gorman Fracture Zones, it has been proposed that the Cocos plate flattens for a distance of 125 km at $\sim$50 km depth [e.g., Suarez et al., 1990; Kostoglodov et al., 1996]. A line containing a sub-horizontal section was fit to the data (Figure 4). A second geometry was determined using a smooth downward curvature, more consistent with the plate profiles to the north and south. Both profiles have a similar shape in the upper 30 km. The resulting thermal models show that the shallow thermal structure (<30–40 km depth) is not affected by the
presence or absence of the subhorizontal region, as the thermal structure is most sensitive to the shallow curvature (Figure 5).

3.4. Convergence Rate

The convergence rate of the Cocos plate ranges from 5 cm yr\(^{-1}\) in the northwest to more than 7 cm yr\(^{-1}\) in the southeast [DeMets and Wilson, 1997]. The margin-normal component of convergence is given in Table 1. The Rivera-North America convergence rate is much more controversial, with estimates ranging from 2 to 5 cm yr\(^{-1}\) [e.g., Kostoglodov and Bandy, 1995]. The most recent plate motion studies that use data from the last 0.78 Myr give convergence rates between 3.3 and 4.3 cm yr\(^{-1}\) along the Jalisco profile [DeMets and Wilson, 1997; Bandy et al., 1998; DeMets and Traylen, 2000]. We use a steady state convergence velocity of 3.8 cm yr\(^{-1}\). Variations of 0.5 cm yr\(^{-1}\) have little effect on the subduction thrust temperatures. DeMets and Traylen [2000] suggest that the convergence history of the Rivera plate is quite complex and variable over the past 10 Myr. On the basis of magnetic lineations, they propose the cessation or near cessation of convergence between 2.6 and 1.0 Ma. We have generated time-dependent thermal models containing the convergence history of DeMets and Traylen [2000]. The effect on the present thermal structure is minimal, and therefore the model results are not presented here. If a 1.6 Myr hiatus in subduction is introduced, the temperature of the top of the subducting plate increases slightly, and the locations of the 350 and 450°C isotherms are shifted ~7 km seaward of their steady state positions.

3.5. Thermal Parameters

Each of the two-dimensional models has four units: oceanic plate, sediments, continental crust, and mantle wedge. The thermal conductivity of the entire continental crust is taken as 2.5 W m\(^{-1}\) K\(^{-1}\). This is a reasonable value for continental crust material [e.g., Peacock and Wang, 1999] and is consistent with that measured during continental heat flow studies of Mexico [Smith et al., 1979; Ziagos et al., 1985]. Although there may be localized regions of the crust with varying conductivity, it is only the large-scale crustal conductivity that is of importance in the current study. The upper 15 km of the continental crust is assigned a radioactive heat generation of 1.3 μW m\(^{-3}\), while the lower 25 km is 0.27 μW m\(^{-3}\). These are typical continental values and reflect the roughly exponential decrease in radioactive heat production with depth (e.g., discussion by Peacock and Wang [1999]). These values are also comparable with measurements by Ziagos et al. [1985] which gave values of 1.3 ± 0.6 μW m\(^{-3}\) for the upper 4 km. Variation in continental heat generation by a factor of two has a minimal effect on the temperature of the thrust fault but significantly affects surface heat flow (Figure 5).

The accretionary prism and sediments have conductivities that increase landward and with depth from 1.0 W m\(^{-1}\) K\(^{-1}\) at the seafloor to 2.0 W m\(^{-1}\) K\(^{-1}\) at 10 km depth. The seafloor conductivity is consistent with marine heat flow measurements in this area [e.g., Vacquier et al., 1967; Prol-Ledesma et al., 1989]. A uniform volumetric heat production rate of 1.0 μW m\(^{-3}\) is assigned to the accretionary prism and sediments, which is similar to the average upper continental source rock for the sediments.

![Figure 5. Variations in (top) surface heat flow and (bottom) temperatures on the top of the oceanic plate for various models along the Guerrero profile. The effects of plate dip profile and amount of continental heat generation are investigated. None of the models contain frictional heating or hydrothermal circulation. Heat flow measurements are from Ziagos et al. [1985].](image)

Parameter values for the mantle wedge and oceanic plate are those used in previous modeling studies [e.g., Hyndman and Wang, 1993, 1995; Oleskevich et al., 1999]. A conductivity of 2.9 and 3.1 W m\(^{-1}\) K\(^{-1}\) is assigned to the oceanic plate and mantle, respectively. Both units have a radioactive heat generation of 0.02 μW m\(^{-3}\) and a volumetric heat capacity of 3.3 MJ m\(^{-3}\) K\(^{-1}\). Reasonable variations in these values have only a small effect on the thermal structure [e.g., Wang et al., 1995a].

3.6. Frictional Heating

The temperature of the thrust fault may be affected by frictional heating [e.g., van den Beukel and Wortel, 1987; Molnar and England, 1990; Dimitru, 1991]. Following the procedure of van den Beukel and Wortel [1987] and Wang et al. [1995a, 1995b], frictional heating is incorporated into the thermal models by dividing the fault interface into two zones: a shallow region of brittle frictional sliding and a deeper region of ductile shear. In the brittle region, the shear stress along the fault is given by Byerlee’s law of friction [Byerlee, 1978]. The shear stress increases with depth, due to the increase in normal stress, approximated by the load of the overlying rock column. The pore pressure ratio is assumed to be constant with depth. A decrease in the pore pressure ratio corresponds to an increase in frictional heat-
ing. In the deeper plastic regime, the magnitude of heating is given by the viscous stress along the fault, determined from the strain rate using the power law rheology for diabase [e.g., Caristan, 1982]. In this region, the magnitude of shear stress decreases with depth, due to increasing temperature. Thus the maximum shear stress is found at the transition between brittle and ductile behavior. The transition from brittle sliding to ductile shear is temperature-dependent, and thus, an iterative approach was used in the modeling, following Wang et al. [1995b]. Frictional heating was introduced to a maximum depth of 40 km. Below this, the fault interface is inferred to be in contact with the serpentined mantle wedge, which is believed to be too weak to allow significant frictional heating.

[24] As the incoming oceanic crust at the Mexico subduction zone is cool, the brittle-ductile transition along the subduction thrust fault occurs at depths >30 km, and frictional heating may not be negligible. We have tested the effects of varying amounts of frictional heating on the thrust fault. Decreasing the pore pressure ratio from 1.0 (no frictional heating) to 0.9 corresponds to an increase in average shear stress along the fault from 0 to 31 MPa, or an effective coefficient of friction (defined by the ratio of the shear to normal stresses along the fault) from 0 to 0.068. These are within the ranges determined in previous studies of subduction thrust faults [e.g., van den Beekel and Wortel, 1987; Wang et al., 1995a, 1995b; Wang and He, 1999]. Figure 6 illustrates the effects of frictional heating on the temperatures of the top of the subducting plate. The large peak in the temperature profile near the Moho intersection is a result of the termination of frictional heating at this depth.

[25] The effect of frictional heating at shallow depths (<20 km) is quite small. With a pore pressure ratio of 0.9, the 100°C and 150°C isotherms are shifted a maximum of 10 km seaward from models without frictional heating. The inclusion of the maximum amount of frictional heating results in a seaward shift of the 350°C isotherm of 23 km (Jalisco) to 89 km (Oaxaca) from its position in models with no frictional heating. The position of the 450°C isotherm is shifted seaward by 30 km (Jalisco) to 170 km (Oaxaca). Frictional heating has the largest effect for profiles with a high convergence rate and shallow plate dip (e.g., the Oaxaca profile). At 40 km depth, the thrust fault along the Oaxaca profile is 200°C warmer with the addition of frictional heating, compared to an increase of ~120°C for Jalisco.

3.7. Hydrothermal Circulation

[26] An additional uncertainty in the shallow thrust temperatures comes from possible hydrothermal circulation within the upper oceanic plate. Marine heat flow studies have shown that the heat flow near the trench offshore Mexico is lower than that predicted for a cooling oceanic plate overlain by sediments (Figure 6). This suggests that hydrothermal circulation in the upper oceanic crust may occur in this region. This mechanism has been used to explain the anomalously low heat flow values observed offshore Costa Rica [Langseth and Silver, 1996; Harris and Wang, 2002].

[27] We developed a “cool crust” thermal model to approximate the effects of hydrothermal circulation for all profiles, using the method of Langseth and Silver [1996] and Harris and Wang [2002]. In the model, the oceanic geotherm was designed to contain a 3 km thick convectively cooled layer that produces a surface heat flow of 30 mW m⁻², consistent with the observations (Figure 3). This model examines the maximum thermal effects on the shallow thrust fault and may not accurately model the deep temperatures. The effect on the surface heat flow is confined to the regions near the deformation front (Figure 6). At distances >100 km, the resulting surface heat flow is indistinguishable from the original models. With hydrothermal cooling to a depth of 3 km, the location of the 100°C and 150°C isotherms are shifted ~10 km landward. For hydrothermal cooling to depths <3 km, the landward shift is less.

3.8. Uncertainties in the Thermal Models

[28] Figure 6 illustrates the range of temperatures along the top of the subducting plate from our modeling, for variations in frictional heating (pore pressure ratio of 1.0–0.9) and the inclusion of hydrothermal circulation. The amount of frictional heating provides the major uncertainty in the models, particularly the deep (>20 km) temperatures.

[29] For each of the models shown in Figure 6, an additional uncertainty comes from the other modeling parameters. In western Mexico, the age of the oceanic plate, the incoming sediment thickness and convergence rate are relatively well constrained. The subducting plate profile remains fairly uncertain, particularly at depths greater than 30–40 km. However, a variation in plate dip by more than 20% is required to produce significant changes in the temperature of the top of the oceanic plate. This is much greater than the estimated uncertainty of 10% in the shallow plate profile. While the values of the thermal parameters (thermal conductivity and radioactive heat production) are relatively uncertain for the continental crust, these parameters have little effect on the temperature of the plate interface. A reasonable variation of 0.5 W m⁻¹ K⁻¹ of the thermal conductivity of the entire continental crust and accretionary prism results in a change in thrust temperature of <15°C. Radioactive heat production in the continental crust significantly affects the surface heat flux, but only has a small effect on the plate surface temperature (Figure 5). We estimate that these factors result in an additional uncertainty in the position of the 100°C isotherm of 15 km. The additional uncertainty in the location of the 350°C and 450°C isotherms is 20 km.

4. Surface Heat Flux Observations

[30] An independent constraint on the thermal models is surface heat flux observations (Figures 5 and 6). Low values of heat flow observed near the trench on the Michoacan and Oaxaca profiles are better fit by models containing hydrothermal circulation. For all models, the forearc has a modeled heat flux of 40–50 mW m⁻² (Figure 6), which is slightly larger than most of the borehole heat flow measurements by Ziagos et al. [1985]. The surface heat flow is highly dependent on the amount of radioactive heat generation in the continental crust. By reducing the crustal heat generation, the modeled values will better fit the observations (Figure 5).

[31] However, the observed heat flux values have large uncertainties. The boreholes used were very shallow (many were <200 m depth). Few thermal conductivity measurements were taken for each hole, and steep topography at many sites required a terrain correction to the data, further increasing the uncertainty. In addition, the complex geo-
Figure 6. Variations in (top) surface heat flow and (bottom) temperatures on the top of the oceanic plate for varying amounts of frictional heating. The thin dashed line indicates a model with no frictional heating; the solid line is a model which uses a pore pressure ratio of 0.94 (medium frictional heating); the thick dashed line is a high frictional heating model, using a pore pressure ratio of 0.90. The introduction of frictional heating causes the peak in the temperature profile near the Moho intersection, as heating was introduced only to this depth. Dotted lines indicate the results for models with hydrothermal circulation to a depth of 3 km within the incoming oceanic plate (thin dotted line is a model with no frictional heating; thicker dotted line is a model with medium frictional heating). For models with hydrothermal circulation to depths <3 km, the effects on thrust temperature are smaller. The rupture widths of past earthquakes are indicated. The Guerrero profile uses the Guerrero B geometry. Heat flow measurements are from Ziagos et al. [1985] (circles) and Prol-Ledesma et al. [1989] (squares).
logical setting of Mexico results in small-scale lateral variations in surface heat flow due to local geological or hydrogeological effects [Prol-Ledesma, 1991]. For example, in the forearc near the Jalisco profile, heat flow values of 13 and 180 mW m$^{-2}$ were recorded within 200 km of each other [Ziagos et al., 1985]. On the basis of these factors, we have assigned an uncertainty of 20% to each heat flow observation. The introduction of frictional heating results in a surface heat flow increase of <10 mW m$^{-2}$ per each earthquake were located between 20 and 100 km from the region (Figure 7a). The majority of aftershocks for this earthquake extend from 20 to 100 km from the trench [Singh et al., 2000a]. The width of the aftershock area is 35 km, while the width of the 1957 earthquake may not extend all the way to the seafloor [Stewart et al., 2000b]. For example, in the forearc near the Jalisco profile, heat flow values of 13 and 180 mW m$^{-2}$ were recorded within 200 km of each other [Ziagos et al., 1985]. On the basis of these factors, we have assigned an uncertainty of 20% to each heat flow observation. The introduction of frictional heating results in a surface heat flow increase of <10 mW m$^{-2}$ per each earthquake were located between 20 and 100 km from the region (Figure 7a). The majority of aftershocks for this earthquake extend from 20 to 100 km from the trench [Singh et al., 2000a]. The width of the aftershock area is 35 km, while the width of the 1957 earthquake may not extend all the way to the seafloor [Stewart et al., 2000b].

5. Mexico Seismogenic Zone
5.1. Past Megathrust Earthquakes
[33] The thermal models can be compared to the rupture widths of past megathrust earthquakes to investigate possible thermal controls on the seismogenic zone. Below, we summarize well-studied earthquakes for each profile.

5.1.1. Jalisco
[34] On 3 and 18 June 1932, two earthquakes occurred north of the proposed Rivera-Cocos plate boundary ($M_c$ 8.2 and 7.8, respectively). Aftershocks from these earthquakes have a seaward limit 10–15 km from the trench and extend to 90–95 km from the trench [Singh et al., 1985]. In 1995, an $M_w$ 8.0 earthquake occurred in the southern part of this region (Figure 7a). The majority of aftershocks for this earthquake were located between 20 and 100 km from the trench [Pacheco et al., 1997]. As the aftershocks were located using a local station array and a local velocity model, we estimate the location uncertainty to be as small as ±5 km. The aftershock area is in good agreement with the results of waveform modeling that determined a nearly rectangular rupture area, oriented parallel to the Middle America trench, with a margin-normal width of 90 km [Mendoza and Hartzell, 1999].

5.1.2. Michoacan
[35] Geodetic observations provide additional constraints on the updip and downdip limits of rupture. Modeling of tsunami records from two local tide gauges indicates that the maximum seafloor uplift was located 24 ± 5 km from the trench [Ortiz et al., 2000b]. The downdip limit was constrained through GPS observations from a dense GPS network located directly onshore. Both Melbourne et al. [1997] and Hutton et al. [2001] were able to fit GPS observations with heterogeneous coseismic slip along the thrust fault to a depth of ~23 km using an elastic model. This corresponds to a downdip limit located 90 km from the trench.

[36] Primarily using the relocated aftershocks for the 1995 earthquake [Pacheco et al., 1997], we have taken the seismic rupture width in the Jalisco region to be 75 km, located between 15 and 90 km from the trench (Figure 7a). On the basis of the good agreement among the various types of observations, uncertainties in this width are small.

5.1.2. Michoacan
[37] On 19 September 1985, an $M_w$ 8.1 earthquake occurred on the Michoacan segment of the Cocos-North America interface (Figure 7b). This was followed 2 days later by an $M_w$ 7.7 earthquake to the southeast. The majority of aftershocks for both events were located between 40 and 120 km from the trench [UNAM Seismology Group, 1986; Stolte et al., 1986]. The width of the aftershock area is slightly narrower in the northwest part of the 19 September rupture area, with the landward limit of the aftershocks ~80 km from the trench. Waveform inversion of the 19 September earthquake indicates that slip occurred between depths of 6 and 39 km [Mendoza and Hartzell, 1989], in good agreement with the aftershock rupture area. The locations of the updip and downdip limits of rupture from both types of studies are relatively well constrained due to a close station spacing and the use of a locally derived velocity model. There are no published tsunami models constraining the updip limit of rupture of the 19 September earthquake. However, distant stations recorded a tsunami with a very low amplitude, which suggests that rupture did not extend all the way to the seafloor [UNAM Seismology Group, 1986]. On the basis of the above studies, seismic rupture in the Michoacan region appears to have occurred between 40 and 120 km from the trench (Figure 7b).

5.1.3. Guerrero
[38] The region northwest of the Guerrero profile is referred to as the Guerrero gap as there have been no large megathrust earthquakes for at least 90 years [e.g., Ortiz et al., 2000a]. However, southeast of the profile, there were two small thrust earthquakes in 1962 within eight days of each other, with magnitudes of 7.1 and 7.0. Southeast of these earthquakes, there was a magnitude 7.8 earthquake in 1957. Although the rupture area of these three earthquakes has been defined using the aftershock distribution (Figure 7c), this area has large uncertainties due to a poor regional station distribution [Ortiz et al., 2000a]. Modeling of the tsunamis generated by these earthquakes has provided stronger constraints on the rupture area. The modeled updip limit of all three earthquakes is located at 60 ± 10 km from the trench [Ortiz et al., 2000a]. The width of the 1962 rupture area is 35 km, while the width of the 1957 earthquake may be as much as 70 km. Uncertainties in the updip rupture limit are large, as this limit is not well constrained by tsunami observations.

5.1.4. Oaxaca
[39] The Oaxaca profile passes through the rupture area of the 1978 $M_w$ 7.6 earthquake (Figure 7d). The 1-month aftershocks for this earthquake extend from 20 to 100 km from the trench [Singh et al., 1980; Stewart et al., 1981]. As
the aftershocks were located relative to the main shock, the
relative positions of each earthquake are very well con-
strained. In addition, the aftershocks extend well under the
cost and were recorded using a number of local portable
seismic stations. Thus the uncertainties in the absolute
lateral location of the group of aftershocks are likely less
than ±5 km (J. Cassidy, personal communication, 2002).

5.2. Thrust Intersection With Mantle

The observed rupture widths for megathrust earth-
quakes near each profile are shown on Figure 7. The
intersection of the continental Moho with the subduction
fault may mark the maximum downdip limit to the seismo-
genic zone. As discussed above, the continental crustal
thickness is not well resolved for the Mexico margin. The
Moho intersection with the subduction fault was taken to
occur at a depth of 40 km. This corresponds to a lateral
distance of 117 km between the Moho intersection and the
trench on the Jalisco profile, increasing to 160 km on the
Oaxaca profile, due to the shallower plate dip in the south-
east. If the crust is as thin as 33 km, the Moho intersection
will be shifted seaward by 15 km (Jalisco) to 25 km
(Oaxaca). For a Moho at 50 km depth, the intersection with
the fault will be shifted landward by 15–35 km. Thus, for
all Moho depths the observed earthquake rupture areas in
Jalisco, Guerrero and Oaxaca are updip of the intersection
of the continental Moho with the thrust fault. In the middle
of the 1985 Michoacan earthquake rupture area, observa-
tions suggest that rupture may have extended to the Moho.
However, to the northwest and southeast, rupture was much
narrower and significantly updip of the Moho intersec-
tion.

5.3. Comparison of Rupture Areas to Thermal Models

A downdip change in properties along the thrust fault
appears to limit the downdip extent of rupture to depths
shallower than the Moho intersection. We focus on the
hypothesis that thrust fault temperature provides the primary
control on the sliding behavior of the fault. As discussed
above, we take the 100 and 350°C isotherms as the thermal
limits on the seismogenic zone of the thrust fault. Assuming
that the proposed thermal limits are correct, it is necessary
to introduce a small amount of frictional heating to the
thermal models of Mexico to be consistent with the

Figure 7. (opposite) Comparison of the aftershock areas
of recent megathrust earthquakes with the thermal model
results. (a) The 11-day aftershocks for the 1995 Jalisco
earthquake [Pacheco et al., 1997]; (b) 11-day aftershocks
for the 1985 Michoacan earthquakes [UNAM Seismology
Group, 1986], also shown are rupture areas and magnitudes
of past earthquakes in this region [after Kostoglodov and
Pacheco, 1999]; (c) 1962 earthquake rupture limits from
tsunami modeling [Ortiz et al., 2000a] and from the poorly
constrained aftershock distribution [Kostoglodov and Pa-
checo, 1999]; and (d) 33-day aftershocks for the 1978
Oaxaca earthquake [Stewart et al., 1981]. The shaded bars
represent the range in the position of the 100 and 350°C
isotherms, based on the models shown in Figure 6. The
solid line indicates the position of these isotherms for the
preferred model. The Moho intersection at 40 km depth is
also shown (dashed line).
observed shallow downdip rupture limit. For models with no frictional heating, the location of the 350°C isotherm is landward of the fault intersection with the Moho and much deeper than the observed rupture limit. If the alternative hypothesis that the Moho intersection provides the maximum downdip limit of the seismogenic zone is correct, the seismogenic zone width would be 100–145 km along the Mexican margin, significantly wider than the observed rupture widths.

Thermal models that include a small amount of frictional heating, using a pore pressure ratio of 0.94, provide the best agreement with the observed earthquake rupture areas along the entire margin. This corresponds to an average shear stress of \( \sim 15 \) MPa, or an effective coefficient of friction of 0.041, consistent with estimates for other subduction faults [e.g., van den Beukel and Wortel, 1987; Wang et al., 1995b; Wang and He, 1999]. It is also consistent with a previous thermal model for Mexico that concludes that frictional heating must be fairly small at shallow depths [Ziagos et al., 1985]. Most subduction thrust faults are believed to be weak. Wang et al. [1995a, 1995b] assumed negligible frictional heating in Cascadia and southwest Japan thermal models. For these subduction zones, a pore pressure ratio of 0.94 leads to an average shear stress of \(<7\) MPa because the brittle part of the fault is confined to depths \(<20\) km. The thermal effect of this small shear stress is well within model uncertainties. Frictional heating was not included in the thermal models of the south Alaska and Chile subduction zones [Oleskevich et al., 1999]. For south Alaska and Chile, even with frictional heating, the position of the 350°C isotherm is much deeper than the Moho intersection with the thrust fault, and therefore does not change their conclusion that the downdip limit of megathrust earthquakes in these regions may be limited by the Moho intersection.

Figure 8 shows the thermal models for each profile, using a pore pressure ratio of 0.94. The temperature along the top of the oceanic plate at depths \(<20\) km is similar for all profiles. The intersection of the 100°C isotherm occurs \( \sim 15–20\) km landward of the trench. The intersection of the 350°C isotherm occurs 92 km from the trench on the Jalisco profile. It is further landward on the other profiles (\( \sim 110\) km). This is due to the younger age of subducting crust, the slower convergence rate, and the steeper plate dip of the Jalisco profile. The downdip transition zone is inferred to extend to the Moho intersection for all profiles. The Moho intersection coincides with a temperature of \( \sim 450\) °C along the Jalisco, Guerrero, and Oaxaca profiles. The temperature at the Moho intersection is \( \sim 400\) °C along the Michoacan profile. Using the proposed thermal limits on the seismogenic zone, the above thermal models give a seismogenic zone downdip width 80–100 km along the margin (Figure 9). There is a very good agreement between the well-constrained observed downdip limit of rupture and the proposed downdip temperature limit of 350 °C in the Jalisco and Oaxaca regions. The agreement is poorer in the Michoacan and Guerrero regions. The majority of seismic rupture of the 1985 Michoacan earthquakes occurred further updip. The downdip limit of rupture in the Guerrero region is poorly constrained, but observations suggest that it was also significantly updip of the proposed thermal limit. The 1962 and 1957 earthquakes in the Guerrero region were quite small. As there are no large scale differences in the factors controlling the thermal regime between the four regions, it is possible that these earthquakes did not rupture the entire downdip width of the proposed thermally limited seismogenic zone.

The shallow fault temperatures are relatively unaffected by frictional heating, and thus, in all models, the proposed updip limit of 100°C is in good agreement with the observed seaward limit of rupture for the 1995 Jalisco and 1978 Oaxaca earthquakes. The updip limit of rupture of the 1985 Michoacan earthquakes corresponds to a temperature of \( \sim 150\) °C. In the Guerrero region, the updip rupture limit of the 1962 earthquakes, as determined through tsunami modeling, is at a temperature of over 200°C. With the inclusion of hydrothermal cooling, the fault temperatures are decreased, and the observed updip rupture limits are more compatible with the proposed thermal limit (Figure 9). The good agreement of the observed updip limit of rupture for most of the megathrust earthquakes with the
proposed thermal limit of 100–150°C suggests that hydrothermal cooling is probably a local effect.

6. Conclusions

[45] We have developed two-dimensional, steady state thermal models for four cross sections through the Mexico subduction zone. Despite the young age of the subducting Rivera and Cocos plates, they are much cooler than oceanic plates in other warm slab subduction zones, such as Cascadia and southwest Japan. This is primarily due to only a thin cover of insulating sediments on the oceanic plates. The thin sediment cover may also allow hydrothermal circulation in the upper oceanic crust, resulting in an even cooler plate surface.

[46] Because of the cool subducting plate, the overall thermal structure of the Mexico subduction zone is intermediate between subduction zones with young, hot oceanic plates (e.g., Cascadia and southwest Japan) and those with older oceanic plates (e.g., south Alaska and Chile). The cool subducting plate surface means that the brittle part of the subducting plate surface extends to depths of 30–40 km. When brittle friction occurs at depths >20 km, even a very weak fault may produce significant frictional heating. Our results show that a small amount of frictional heating can have a large effect on the thrust temperatures. The deep (30–40 km) part of the thrust fault can be increased by over 200°C with the inclusion of only a small amount of frictional heating. The effect of frictional heating is largest for areas that have a shallowly dipping plate and a high convergence rate.

[47] For the Mexico subduction zone, megathrust earthquake rupture is confined to depths <30 km. If the downdip limit of rupture is controlled by a temperature of 350°C, a small amount of frictional heating must be included in the thermal models. The models that best fit the observed rupture along the entire margin use an average shear stress of 15 MPa, consistent with the conclusion of past studies of a weak subduction fault and low frictional heating. The observed updip limit of coseismic rupture, although poorly constrained, is in agreement with a thrust temperature of 100°C for much of the margin. For all models, the proposed downdip thermal limit of the seismogenic zone occurs landward of the coast, with a transition zone that extends up to 80 km inland. This is important for seismic hazard studies for Mexico, as the seismic source zone could underlie populated areas.

Acknowledgments.

We thank S. Mazzotti, S. Peacock, S. Schwartz, and an anonymous reviewer for their thoughtful reviews which improved the manuscript. We acknowledge helpful discussions with J. Cassidy, H. Dragert, W. Hutton, T. Lewis, G. Rogers, and S. Singh. Maps were generated using GMT software [Wessel and Smith, 1995]. Support for C. Currie provided by an NSERC graduate scholarship and the Canadian GEOIDE project. Research support provided by USGS NEHRP grants 00HQGR0061 and 01HQGR0058. Geological Survey of Canada contribution 2001086.

References


V. Kostoglodov, Instituto de Geofisica, Universidad Nacional Autonoma de Mexico, Coyocan, DF 04510 Mexico City, Mexico. (vladimir@ollin.igeofcu.unam.mx)