Evolution of shallow, crustal thermal structure from subduction to collision: 
An example from Taiwan

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ABSTRACT

We study crustal thermal evolution by examining heat flow patterns along a convergent boundary from a young subduction zone to a more structurally mature collision zone. More than 8000 km of seismic profiles covering an offshore region of 45,000 km² in southern Taiwan show widespread bottom-simulating reflectors (BSRs). We derived 1107 BSR-based heat flows before combining 42 additional, published, offshore thermal probe data and 86 on-land heat flow data to document the shallow forearc thermal structures from the subduction zone to the collision zone. In the subduction zone, the geothermal gradient ranges mostly within 30–80 °C/km, and decreases toward the arc due to slab cooling, intensive dewatering at the toe, sediment blanketing, topographic effects, and other processes. The geothermal gradient ranges mostly from 30 to 90 °C/km in the collision zone, and increases, instead of decreases, toward the arc, possibly caused by exhumation, erosion, topographically induced groundwater circulation, and some upper mantle processes related to collision. Heat flow in the collision zone ranges from 80 to 250 mW/m². The high heat flow in the collision zone correlates with a shallower seismicity zone and high seismic attenuation, while the lower heat flow in the subduction zone might allow the earthquakes to rupture to greater depth. The heat flow increases along the topographic high from subduction to collision zone due to increasing geothermal gradients and higher thermal conductivities of the exhumed basement rocks. This heat flow variation may generate an artificial exhumation rate pattern, if a conventional 30 °C/km geothermal gradient was used in fission-track studies.

Keywords: bottom-simulating reflector, hydrates, geothermal gradient, heat flow, Taiwan, subduction, collision, seismogenic zone, fission-track dating.

INTRODUCTION

Many mountain belts around the world are the result of collisions preceded by subduction. Arc-continent collisions have been interpreted as important processes for the growth of continents. At the same time, many geological and geophysical processes are affected by thermal structures of the crust, e.g., metamorphism, crustal rheology changes, the seismogenic process, seismic attenuation, and maturation of hydrocarbons. Currently, there are few observational studies on crustal thermal evolution from subduction to collision.

Taiwan is located along a convergent boundary with a subduction to the south and an arccontinent collision to the north. Thus, by studying transects from south to north, we can better understand how thermal structures evolve from subduction to collision. In the collision zone on-land, there is an extensive heat flow data set from exploration and thermal wells (Lee and Cheng, 1986). However, regional-scale geothermal gradient and heat flow in the offshore region in the initial collision zone and subduction zone have not yet been studied systematically.

Here we used the gas-hydrate, bottom-simulating reflector (BSR) offshore southern Taiwan to derive 1107 geothermal gradients in the accretionary prism in the subduction zone and initial collision zone. We compared the BSR-based thermal structures in the subduction zone with those in the collision zone. The BSRs were found comparable with recently published in situ heat flow measurements in the Taiwan region. Different mechanisms affecting the regional thermal patterns and the maximum perturbations from these processes were discussed. We also speculated on the implications of the different thermal structures on fission-track data interpretation, pressure-temperature-time (P-T-t) paths of metamorphic rocks, frictional heat on the fault, seismogenic zone width, seismic attenuation, and other geodynamic and seismological processes. Overall, Taiwan is a modern example that shows dramatic changes in thermal structures from subduction to collision both in space and in time.

REGIONAL GEOLOGIC SETTING

The offshore Taiwan accretionary prism (Fig. 1) is located along the boundary between the Eurasian plate and Philippine Sea plate where the oceanic lithosphere of the South China Sea is subducting eastward beneath the Philippine Sea plate (Bowin et al., 1978; Ho, 1986) with a high rate of 7–9 cm/yr (e.g., Yu et al., 1997). Oceanic crust in the subducting South China Sea can be as old as 37 Ma, as inferred from marine magnetic anomalies (Hsu and Sibuet, 2004). The submarine accretionary prism (Fig. 2) contains three distinct structural domains (Reed et al., 1992): (1) a lower slope domain composed of mostly west-vergent ramp anticlines; (2) an upper slope domain with highly discontinuous reflections, suggesting intense deformation; and (3) a backthrust domain located along the rear of the prism. The prism is bounded on the west by the Manila Trench and on the east by the forearc basin of the North Luzon Trough.

Near Taiwan, subduction changes into an arccontinent collision where the Chinese continental margin enters the subduction zone (Suppe, 1981). The initial collision zone is probably near latitude 21°N where the continent-ocean boundary (COB) enters the trench (Fig. 1). As the Luzon arc approached the Taiwan and Chinese passive margin, the volcanism started to cease. The forearc basin is getting consumed until the Luzon volcanic arc is juxtaposed next to the fold and thrust belt of on-land Taiwan in the mature collision zone. Although the timing of initial
collision is still debated, Teng (1990) proposed that it probably began in the late-middle Miocene and is currently propagating southward (Suppe, 1984) toward the modern subduction zone offshore Taiwan.

Further to the north in the mature collision zone on Taiwan, the geologic provinces from west to east include the Coastal Plains, Western Foothills, Hsuehshan Range, Central Range, and Coastal Range (Fig. 2). The Coastal Plains are composed mostly of recent alluvium deposits overlying continental shelf materials of the Chinese passive margin. The Western Foothills are composed of Oligocene to Pleistocene shallow marine materials. The Hsuehshan Range consists of quartz and carbonate sandstone, argillites, and shales. The western side of Central Range consists of slatey materials, while the eastern part is mostly underlain by the pre-Tertiary metamorphic rocks, including schist, gneiss, carbonates, tuffs, and oceanic pelitic and mafic schists (e.g., Ho, 1986). Overall the metamorphic grade increases eastward in the Central Range. Some of the metamorphic rocks have been interpreted as the basement rocks of the Eurasian continental crust (Lan et al., 1990). The Central Range and the offshore upper slope domain (Fig. 1) form a continuous bathymetric and topographic high from collision to subduction. Further east is the Coastal Range, containing both volcanic and sedimentary rocks of Luzon arc affinity. Slab earthquakes illuminate a clear, east-dipping Benioff zone east of the Coastal Range. However, the seismicity disappears abruptly north of 23.3°N. It is still a matter of debate as to whether there is a slab between the latitudes of 23.3°N and 24°N (e.g., Teng et al., 2000; Lin, 2002).

In the northern part of Taiwan, the polarity of the subduction flipped, and the Philippine Sea plate subducts north underneath the Eurasian plate along the Ryukyu Trench. The Benioff zone is well imaged underneath the eastern part of northern Taiwan, at least north of 24°N (Kao et al., 1998). To reduce complications from processes related to the Ryukyu Trench, in this study we will focus mainly on thermal structures south of 24°N.

**PREVIOUS STUDIES ON CONVERGENT BOUNDARY ZONE THERMAL STRUCTURES**

Most subduction zones around the world show high geothermal gradients and heat flows near the trenches. The heat flows decrease within the accretionary prism before increasing toward the arcs. For example, across northern Honshu in Japan, the heat flow is ~55 mW/m² near the trench, decreasing to ~25 mW/m² near the bathymetric high region, then increasing to more than 100 mW/m² toward the volcanic arc and backarc regions (Jessop, 1990). In Cascadia, the heat flow is ~120 mW/m² near the trench and decreases to ~60 mW/m² in the continental shelf region (Hyndman and Wang, 1995; Hyndman and Lewis, 1999; Lewis et al., 2003). The heat flow then increases toward the Cascade volcanoes and reaches more than 100 mW/m² (Jessop, 1990). The Chilean subduction zone near the Chile triple junction also shows high geothermal gradient near the trench and lower geothermal gradient arcward (Brown and Bangs, 1995). In a local scale near the toe of the accretionary prism, heat flow studies in the Barbados Ridge accretionary prism show high heat flow associated with thrust zones, implying focused fluid flow at the décollement and some thrusts (Foucher et

**Figure 1. Location map of Taiwan. To the south in the offshore region, the South China Sea plate (SCS) is subducting underneath the Philippine Sea plate (PSP), forming the Taiwan accretionary prism. The subduction changes into an arc-continent collision when the Chinese passive margin enters into the convergent boundary. The black arrow shows the direction of the relative motion of the Philippine Sea plate with respect to a stable Chinese continent. COB marks the continent-ocean boundary. The rapid uplift forms the island of Taiwan in the collision zone. On Taiwan, from west to east, are Coastal Plain (CP), Western Foothills (WF), Hsuehshan Range (HR), Central Range (Cen R), and Coastal Range (CR). There is a bathymetric and topographic high that extends from upper slope domain offshore to the Central Range. Eurasian plate is marked as EP. A–A′ and B–B′ show the locations of the two profiles in Figure 2.**
Crustal thermal evolution from subduction to collision

Figure 2. A schematic showing structural domains, heat flow patterns, and possible mechanisms causing different thermal structures in a subduction zone transect and a collision zone transect in Taiwan. In the collision zone at a latitude of 23.8°N (see A–A' in Fig. 1 for location), we have plotted previously published heat flow (Lee and Cheng, 1986) within 5 km of the profile in small circles with error bars, exaggerated topography, and seismicity within 35 km of the profile. Lin (2002) proposed that the exhumation of the crust will reduce the seismicity due to higher temperature at depth, thus the higher heat flow. Other mechanisms, including groundwater circulation, high erosion rate, higher radiogenic heat production rate, topographic effect, and other processes can also increase the heat flow. Lower panel shows a profile in the subduction zone at a latitude of 20.5°N. See B–B' in Figure 1 for location. We have plotted the heat flows derived from this study in large circles with error bars. All these heat flows are within 5 km of the profile. Unlike the heat flow profile in the collision zone, here heat flow decreases toward the arc in a pattern similar to other subduction zones around the world, possibly due to slab cooling, dewatering near the toe of the accretionary prism, increasing blanketing effect toward the ridge, and topographic effects. Triangles are the theoretical heat flows derived from our slab-cooling modeling, which fit the overall pattern of the BSR-based heat flows. See text in Discussion section for more details.
Chi et al. (1990). Similar features have been reported at Nankai accretionary prism (Yamano et al., 1992) and Taiwan (Chi et al., 1998).

Observed data on thermal structures in the collision zone are sparse, particularly for arc-continent collision zones; yet, Taiwan is one of the best-studied regions in terms of crustal thermal structures (cf. Barr and Dahlen, 1989; Hwang and Wang, 1993; Lin, 2002). However, the thermal structures in the initial collision zone and subduction zone offshore southern Taiwan have not been studied systematically until this study (Fig. 2), in which we found a dramatic difference in heat flow patterns between the subduction zone and the collision zone in the Taiwan region.

**BOTTOM-SIMULATING REFLECTOR, GAS HYDRATE, AND GEOTHERMAL GRADIENT**

A bottom-simulating reflector, or BSR (Fig. 3), is a seismic reflector subparallel to the topography of the seafloor, which, in some places, cuts across reflectors generated by sediment layers (Tucholke et al., 1977; Shipley et al., 1979; Bangs et al., 1993). BSRs are commonly observed in continental slope sediments, particularly those associated with accretionary prisms (Hyndman and Davis, 1992; Brooks et al., 1991; Miller et al., 1991). These seismic reflectors are generated by the acoustic impedance contrast at the base of a sediment layer containing methane hydrate (Shipley et al., 1979; Holbrook et al., 1996).

Methane hydrates form when water and methane combine to generate a crystalline substance that looks like ice. Most hydrates have been found in deep ocean sediments where there is a sufficient supply of methane and where pressure and temperature ranges are between 0.2 and 5 MPa and 0–25 °C, respectively, as inferred from Kvenvolden and McMenamin (1980). The increase in temperature with depth below the seafloor causes methane hydrate to become unstable and decompose, despite increasing pressure. As a result, the base of the methane hydrate defines a “phase boundary” that separates the stable gas hydrate above from a field of instability below. The acoustic impedance contrast across this phase boundary thus can be imaged in seismic profiles as a BSR.

In particular, P-T conditions at the BSR have been used to estimate heat flow in several accretionary prism settings (Yamano et al., 1982; Yamano et al., 1992; Hyndman et al., 1992; Ashi and Taira, 1993; Zwart and Moore, 1993; Townend, 1997). The temperature at the BSR can be inferred from two-component (methane and water), hydrate-phase equilibria and estimated in situ pressures. By converting the two-way traveltime of the BSR to depth and using a temperature at the seafloor, the regional geothermal gradient and heat flow can be estimated using the relation:

\[ Q = -k \frac{dt}{dz} \]

where \( Q \) is heat flow in mW/m², \( k \) is sediment thermal conductivity in W/mK, \( \frac{dt}{dz} \) is geothermal gradient in °C/km, \( t \) is temperature in °C, and \( z \) is subbottom depth in km.

**DERIVED BOTTOM-SIMULATING REFLECTOR-BASED HEAT FLOW OFFSHORE TAIWAN**

The bottom-simulating reflector (BSR) in the Taiwan region and its relation with gas hydrate was first documented in Reed et al. (1991). Chi et al. (1998) systematically studied the BSR offshore Taiwan and discussed the factors controlling its distribution. Schnurle et al. (2004) used seismic reflection and ocean bottom seismic data to study the BSR in some of this area. The gas component in the hydrates offshore Taiwan is mostly methane. The gas seepage from mud volcanoes near this region on land is mainly composed of methane with minor ethane and other gases; water column samples in this

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Figure 3. (A) An example of a Q1 bottom-simulating reflector (BSR) with reversed polarity reflection located in an anticlinal ridge. The coordinate marks the location of the left edge of the profile. (B) Examples of Q2 and Q3 BSRs in this region. The coordinate marks the location of the left edge of the profile.
offshore region also show high dissolved methane concentrations (Yang et al., 2006).

We use data acquired during two marine geophysical surveys (Fig. 4). The first data set was collected in 1990 aboard the R/V Moana Wave of the University of Hawaii. The second data set is from the 160-channel seismic reflection survey in 1995 aboard R/V Maurice Ewing of Columbia University. The six-channel data were processed using the SIOSEIS processing software. The 160-channel reflection data were processed using ProMAX (e.g., Chi et al., 2003). The processing sequences of these two data sets are listed in Table 1. More than 8000 km of seismic reflection profiles were examined covering a region of 45,000 km².

At least 15,000 km², or 70% of the Taiwan accretionary prism, is covered by a BSR that is identified as a reflector subparallel to the seafloor with reversed polarity. Reversed polarity, increasing subbottom depth with increasing water depth, the distribution of the BSR, and methane emitted on land in the nearby region strongly suggest that this feature marks the base of the methane hydrate stability field.

The BSR typically exhibits a single symmetrical pulse characteristic of a simple interface (Fig. 3). We have made a determined effort throughout the study to separate reflectors associated with water-bottom reverberations from BSR identifications. For example, shallow reflections marked by constant subbottom depths beneath a dipping seafloor reflection will be interpreted as water-bottom reverberations, instead of a BSR. The BSR was identified at more than 1000 locations along the survey lines, and each pick was assigned to one of three categories according to the confidence level of the identification (Q1, Q2, or Q3). The BSR was assigned into the Q1 category if it showed the following reflector characteristics: (1) subparallel to the seafloor in shallow subbottom depths, (2) reverse polarity with a single symmetrical pulse, and (3) a strong coherent reflection that cuts across bedding. The Q2 category, or probable BSR identifications, showed similar characteristics to the Q1 picks except that the reflections were either segmented into pieces or did not cut bedding. The segment lengths of the Q2 reflectors were usually longer than the gaps between segments. The Q3 category, containing possible BSR picks, exhibited very weak and segmented reflections that were subparallel to the seafloor. BSRs in the Q3 group usually could be traced from adjacent Q1 or Q2 categories without abrupt changes in the subbottom depth.

Next, we compiled water depths and the BSR subbottom depths. The two-way traveltime of BSR subbottom depths was converted into BSR subbottom depths in meters by using velocity-depth relations published by Hamilton (1980):

\[ H_{\text{BSR}} = \frac{V_{\text{avg}} \times t}{2} = \frac{1511 + (1041 \times t) - (372 \times t^2)}{2} \times t, \]

where \( H_{\text{BSR}} \) is BSR subbottom depth in m, \( V_{\text{avg}} \) is the average velocity of sediments above BSR in m/s, and \( t \) is the one-way traveltime of BSR subbottom depth in sec. The regional two-way traveltime of the BSR subbottom depth is, on average, ~0.5 s, translating to 460 m of BSR subbottom depth.

We fit the methane hydrate phase boundary data published by Hyndman and Davis (1992) using this equation:

\[ T_{\text{BSR}} = 2.03 \times \log(H_{\text{water}} + H_{\text{BSR}}) - 2 \times 9.75 - 4, \]

where \( T_{\text{BSR}} \) is the temperature at BSRs in °C, and \( \log \) is the logarithms to base 10.

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**Table 1. Reflection Data Processing Sequences**

<table>
<thead>
<tr>
<th>Moana Wave Dataset:</th>
<th>Ewing Dataset:</th>
</tr>
</thead>
<tbody>
<tr>
<td>Resample to 4 msec</td>
<td>Resampling at 4 msec</td>
</tr>
<tr>
<td>Geometry</td>
<td>Geometry</td>
</tr>
<tr>
<td>Automatic gain control</td>
<td>True amplitude recovery</td>
</tr>
<tr>
<td>Spiking deconvolution</td>
<td>F-K demultiple in shot domain</td>
</tr>
<tr>
<td>Normal move out</td>
<td>F-K demultiple in receiver domain</td>
</tr>
<tr>
<td>Stack</td>
<td>Dip move out</td>
</tr>
<tr>
<td>Filter</td>
<td>Velocity analysis</td>
</tr>
<tr>
<td>Predictive deconvolution</td>
<td>Radon filter</td>
</tr>
<tr>
<td>Water-bottom mute</td>
<td>40-fold weighted stack at 12.5 m CDP spacing</td>
</tr>
<tr>
<td>Trace weighting</td>
<td>Time variant frequency filter</td>
</tr>
<tr>
<td>Filter</td>
<td>F-K or F-D migration</td>
</tr>
<tr>
<td>Automatic gain control</td>
<td>Automatic gain control</td>
</tr>
<tr>
<td>Fourfold stack section plot</td>
<td>Migrated section plot</td>
</tr>
<tr>
<td>F-K migration at 1500 m/s</td>
<td></td>
</tr>
<tr>
<td>Filter</td>
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<tr>
<td>Automatic gain control</td>
<td></td>
</tr>
<tr>
<td>Migrated section plot</td>
<td></td>
</tr>
</tbody>
</table>

Note: CDP—Common-Depth Point; F-D—Finite-Difference; F-K—frequency-wave number.
Another equation was derived to represent the seafloor temperature by least-squares fitting the water temperature and depth data provided by the National Center for Ocean Research of Taiwan:

\[
T_{sf} = 0.2597 \times (\ln H)^3 - 3.802 \times (\ln H)^2 + 10.67 \times (\ln H) + 26.96,
\]

where \( T_{sf} \) is temperature of the seawater right above seafloor in °C. \( \ln H \) is the natural log of water depth in meter. The water depth in this data set ranges from 600 to 3700 m; therefore, the \( T_{sf} \) ranges from 7.5 to 2.0 °C assuming that the temperature on the seafloor sediment is similar to that of the seawater.

These two quantities (\( T_{sf} \) and \( T_{BSR} \)) were used to determine the temperature difference between the seafloor and the BSR, which then was divided by the BSR subbottom depth to estimate geothermal gradient. The BSR-based regional geothermal gradient ranges from 17 to 150 °C/km. The mean and standard deviation of the geothermal gradient are 44.7 ± 12.9 °C/km.

We have also included previously published geothermal gradients in both subduction and collision zones in Figure 5 (e.g., Shyu et al., 2006; Lee and Cheng, 1986).

These previously published data sets usually also include thermal conductivity measurements. However, the conductivity data from offshore BSR locations at several hundred meters subbottom depth are currently not available. We used a published depth-porosity profile from a well in the offshore region farther to the west (Lin et al., 2003) to derive a 1D, depth-dependent, average thermal conductivity model. The average seafloor conductivity of the 42 thermal probe measurements in this region is 0.97 W/mK (cf. Shyu et al., 2006). With this value as the arithmetic mean of the conductivities of the matrix (37% volume at seafloor) and the pore fluid (63% volume), we used 0.6 W/mK as the pore fluid conductivity to derive the conductivity of the matrix in this region, which is 2.2 W/mK. This matrix conductivity value then was used to convert the depth-porosity relation to a depth-conductivity relation. Using an average regional heat flow of 48 mW/m², we then calculated the depth-temperature profile. For each subbottom depth, we calculated its corresponding average conductivity by dividing the 48 mW/m² heat flow by the temperature increase from the seafloor to that subbottom depth. In addition, we did a sensitivity test by using 60 mW/m², instead of 48 mW/m², as the regional heat flow and found that the average thermal conductivities differ by less than 3%. For the whole BSR data set, the derived thermal conductivities range from 1.11 to 1.29 W/mK with a mean of 1.25 W/mK. This

Figure 5. Geothermal gradient pattern from subduction zone to collision zone in the Taiwan region. Large stars indicate Q1 class, or highest confidence in BSR identification. Medium stars indicate Q2 class, or probable BSR identification. Small stars indicate Q3 class, or possible BSR identification. The large circles are from previously published thermal probe measurements (Shyu et al., 1998; Shyu et al., 2006). Different sizes of the hexagons represent different geothermal gradient data sets on land in the collision zone, which are discussed in Lee and Cheng (1986). Note higher geothermal gradients along the trench in the subduction zone and along the topographic high in the collision zone. The three red boxes mark the locations of the possible hydrothermal vents. BHT—bottom-hole temperature; BSR—bottom-simulating reflectors; CPC—Chinese Petroleum Corporation.
translates to up to ±15% variations of thermal conductivity. We then derived the BSR-based heat flow using the geothermal gradient and thermal conductivity. BSR-based heat flow values (Fig. 6) were estimated at 15–146 mW/m², while both the mean and standard deviation are 43.4 ± 12.5 mW/m². The BSR-based heat flow data set was combined with previous offshore and on-land heat flow in situ measurements (Fig. 7). Note that these data sets were derived from different methods, but the first-order features for the crustal thermal structures from the subduction zone to the arc-continent collision should be robust. All these data sets were derived from shallow depths and range down to several hundred meters. However, some of the on-land measurements from Lee and Cheng (1986) can reach kilometers in depth.

DISCUSSION
Uncertainty in Estimating Geothermal Gradient

In this study, we discuss both geothermal gradient and heat flow. Overall, the geothermal gradients derived from this study are more robust than the heat flow because of relatively larger uncertainty on the thermal conductivity.

There is a large scattering of the BSR subbottom depths, ranging mostly from 250 to 600 m (0.2–0.7 s in two-way traveltime). This scattering represents the wide range of thermal gradients derived from this study. The uncertainties in measuring the two-way traveltime of the seafloor and the BSR are less than 0.05 s. Subbottom depths of BSRs in two-way traveltime can be influenced by a number of factors, including spatial and temporal changes in P-T conditions and lateral variations in sediment velocity above the BSR.

Variations in BSR subbottom depths, particularly along the surface of modern accretionary prisms, are commonly attributed to changing P-T conditions beneath the seafloor. The time span for the hydrates to respond to P-T condition changes is still not clear. Torres et al. (2002) showed that gas hydrate stability responds rapidly to tidal-sea–level changes, implying that the BSR depths reflect the modern P-T regime. An experiment conducted by a remotely operated vehicle in the deep waters also shows that hydrate formation occurred within minutes (Brewer et al., 1997).

We studied the uncertainty in the estimates of BSR-based geothermal gradients. Assuming the picked BSR correctly represents the hydrate phase boundary, we estimated the uncertainty based on error analysis of seafloor temperature, BSR temperature, and BSR subbottom depth.

Based on the data from the National Center for Ocean Research of Taiwan, the maximum seawater temperature variation is about ±3 °C. For an average BSR subbottom depth of ~500 m, it translates to ~6 °C/km of uncertainty in geothermal gradient estimates.

For the BSR in situ temperature, previous studies found that using hydrostatic pressure yields a good result (e.g., Ashi and Taira, 1993). But we also calculated the in situ lithostatic pressure using 2.3 g/cm³ as the average density of the sediment layer above the BSR. All the geothermal gradients were recalculated using lithostatic pressure and compared with our geothermal gradient model based on hydrostatic pressure. From the histogram of the discrepancies between these two types of geothermal gradients, we found that most of the geothermal gradients will increase between 3–7 °C/km, if lithostatic pressure is used.

It is difficult to estimate the errors on BSR subbottom depth because currently there is no detailed velocity analysis on a regional scale. However, we found that, for an arbitrary 30% error in estimating the BSR subbottom depth, the geothermal gradients will change only 0.5–2.5 °C/km, which is relatively small. This is because the BSR subbottom depth error will also be mapped into error in hydrostatic pressure estimates, which in turn will change the estimated BSR temperature, thus compensating for the effect that came from errors in BSR subbottom depth in estimating the geothermal gradient. The degree of lateral velocity variation
Above the BSR in this region is still unclear. The depth-velocity relation of Hamilton (1980) might be better suited for the sediments in the lower slope domain, while the sediments in the upper slope domain might have higher P velocity at a given subbottom depth due to compaction from more intensive deformation. Saturation of the hydrates above the BSR could also increase the velocity. However, this effect might be relatively minor. In summary, the total error should be less than 15 °C/km, whereas most of the BSR-based geothermal gradients will have error within 10 °C/km. The thermal probe geothermal gradients are consistently slightly higher than those derived from the BSRs (Fig. 5), possibly due to the smaller thermal conductivity on the seafloor compared to that in the sediments above the BSR.

Regional Thermal Conductivity Model at Shallow Depths

Local heterogeneity in thermal conductivity will also affect the local geothermal gradient pattern. Currently we do not have detailed information on the thermal properties of the crust in the offshore region. The 42 seafloor in situ measurements that cover 70% of this region (Shyu et al., 1998; Shyu et al., 2006) give thermal conductivities on the seafloor that range from 0.79 to 1.19 W/m°C. Most of them are between 0.85 and 1.05 W/m°C and show little systematic spatial variation.

The thermal conductivities from this study range from 1.11 to 1.29 W/mK with a mean of 1.25 W/mK. The higher conductivities occur because we have considered the reduced porosity in sediments from seafloor to the BSR. The ±15% thermal conductivity variation derived in this study is similar to that of thousands of in situ thermal conductivity measurements from drilling in Peruvian, Cascadia, and Blake Ridge margins (cf. Figure 5 of Grevemeyer and Villinger, 2001). However, larger variations of ±40% were found in the Nankai and Chile margins (Grevemeyer and Villinger, 2001). Without drilling data from the offshore Taiwan region, it is difficult to estimate the error ranges of our conductivity estimates. However, because the geothermal gradients in the offshore region are mostly between 31 and 56 mW/m², the errors that propagate into our heat flow estimates should be relatively small (mostly <20 mW/m²).

This thermal conductivity model does not take into account the possible influence from different lithology. However, the BSR is most likely found in sedimentary layers, not in the basement rocks. As a result, the thermal conductivity above the BSR should be similar and does not vary a lot such as that in the collision zone, where different rock types were found in the outcrops.

Regional Heat Flow Model

We multiply the BSR-derived geothermal gradient with depth-dependent thermal conductivity to get the heat flow. In total, 1107 BSR-based heat flows were derived from this study (Fig. 6).

In situ heat flow measurements in and near this region (Nissen et al., 1995; Xia et al., 1995; Shyu et al., 1998; Shyu et al., 2006) are consistent with the BSR-derived geothermal gradients and heat flows from this study. Twenty-two out of the 42 thermal probe heat flows were measured near the BSRs and, except at shallow-water depth regions, the deviation ranges mostly from 0 to 16 mW/m², which is within 30% of the regional heat flow. Because most of the geothermal gradients fall between 31 and 56 °C/km, most of the errors will be within 10–17 °C/km, which is similar to the results from our uncertainty analyses from the previous section. The absolute values of the resulting heat flow are less accurate, but the spatial variations are well determined.

BSR-derived heat flow is an average heat flow for the top 200–400 m sedimentary column, compared with thermal probe that samples...
very shallow depths. As a result, BSR-derived geothermal gradient is less influenced by periodic surface temperature variation. Actually, the deep-sea stable temperature environment will dramatically reduce the temperature perturbation from periodic seasonal or even paleoclimatic temperature changes that might have affected the modern heat flow measurements as proposed by Kohl (1998) and others. For the shallow-water depth region where the seafloor temperature varies greatly, the average 300 m BSR subbottom depth in the shallow depth region is still deeper than the skin depth for temperature perturbation with periods longer than 10,000 yr (cf. equation 4–90 of Turcotte and Schubert, 1982). In summary, the BSR-based heat flows derived from this study are consistent with the thermal probe measurements, especially in the deep water regions. By combining the BSR-based heat flow data sets with the published heat flow measurements, we will next discuss the thermal structures in both subduction and the collision zones.

Crustal Thermal Structures

At a latitude of ~20.5°N in the subduction zone, heat flow decreases from 75 to 40 mW/m² from the trench to the upper slope domain of the accretionary prism (Fig. 2). To the east in the forearc basin, a BSR in a fold gives a heat flow of ~25 mW/m² (Fig. 6). The heat flow pattern along this transect is consistent with the three in situ heat flow measurements farther to the south at ~19°N (Xia et al., 1995). Heat flows in the satellite basins in the arc region are ~50 mW/m² and are derived from some low-quality BSRs.

Farther to the north in the initial collision zone (Fig. 5), the continent-ocean boundary (COB) enters into the trench near 21.2°N. Nissen et al. (1995) have collected 12 thermal probe measurements along a transect from continental shelf (117°E, 22.8°N) to continental slope (118.1°E, 19.3°E) that is 220 km west of and parallel to the trench. We can treat this data set as the initial condition before the Chinese passive margin enters into this convergent boundary. Nissen et al. (1995) found that heat flows are ~80 mW/m² in the continental slope and decrease to 70 mW/m² in the abyssal plain. From BSR-based heat flow, we found slightly increased heat flows once the incoming sediments were scraped off and incorporated into the toe of the accretionary prism, where intensive dewatering occurs.

High geothermal gradients (40–80 °C/km) and heat flows (50–105 mW/m²) were found in a thick basin near the toe in this region east of the COB (Fig. 6). This might be a result of intensive dewatering in this basin, which covers a circular region with a diameter of 60 km centered at 119.8°E, 21.6°N. Seismic reflection data show conjugate fault plane reflections within the basin, even though the displacements across the faults are small, suggesting possible fluids within the fault zones. Cores recovered in this region show manganese nodules, indicating active hydrologic processes in this region (N. Lundberg, 2003, personal commun.). Similar high heat flows also were found across continent-ocean transform boundaries around the world (e.g., Vogt and Sundvor, 1996). Vogt and Sundvor (1996) proposed that the great sedimentary thickness of the continental slope will enhance dewatering, thus generating high heat flow on the Norwegian-Barents-Svalbard continental slope. Moreover, along the Australia margin, the high heat flow was modeled as a result of thermal uplift, local isostatic rebound, and other processes (Lorenzo and Vera, 1992).

Away from the toe in the initial collision zone, the heat flows decrease toward the arc as the sediments stack thicker, especially from lower slope domain to the upper slope domain. Some clear BSRs were identified in the backthrust domain with heat flows ranging from 30 to 60 mW/m² and increasing toward the arc.

On Taiwan in the mature collision zone, the heat flows are high in the mountains, mostly between 80 and 250 mW/m². Whether the heat flow data from some of the “geothermal wells” are representative of the regional heat flow pattern is still under debate. But several studies are able to fit the high heat flow pattern by thermal modeling using different crustal kinematic models (Barr and Dahlen, 1989; Hwang and Wang, 1993; Lin, 2002). The high heat flow pattern is also consistent with observed shallowing of seismicity (Lin, 2002). Overall, the high heat flow in the mountainous regions is interpreted as a result of rapid uplift and exhumation of the warmer material from depth. Other processes, including groundwater circulation, topographic effects, and higher radiogenic heat production rates in the continental crust will also contribute to the high heat flow. Complex deformations in lower crust and upper mantle following the collision might have also affected the thermal structures in this region.

FACTORS AFFECTING SHALLOW THERMAL STRUCTURES

Cooling Effect of the Subducting Plate

We compared the theoretical heat flow values derived from the age of the subducting plate with the BSR-derived heat flow. The predicted heat flow ranges from 90 to 84 mW/m², based on the ages of the subduction plate (32–37 Ma; Hsu and Sibuet, 2004) and the updated seafloor age–heat flow relation (Stein and Stein, 1992).

The average 44.5 mW/m² derived from BSR is about half the average predicted heat flow of 87 mW/m².

Here we studied the cooling effect due to the rapid subduction of the South China Sea plate based on Molnar and England (1990) that was slightly modified by Von Herzen et al. (2001):

$$Q = \frac{Q_{0\text{flux}}}{S}$$

where $Q_{0\text{flux}}$ is the reference heat flow in the region, $v$ is the rate of orthogonal plate convergence, and $\tau$ is the shear stress associated with slip on the fault. The term $\nu \tau$ is for frictional heating on the fault. $S = 1 + b(z, \nu \sin \alpha)$, where $z$ is depth of the fault below the heat flow measurement, $\nu$ is the dip of the fault separating the upper and lower plates, and $\alpha$ is the average thermal diffusivity of the plate.

We used the following parameters to calculate the heat flow profile at the latitude of 20.5°N in the subduction zone: $Q_{0\text{flux}} = 87$ mW/m², $v = 85$ mm/yr *cos(50°), which is the full convergent rate times cosine of the angle between relative plate motion and the strike of the trench axis. Based on a crustal model derived from a seismic profile and gravity modeling (Chi et al., 2003), we used an average dip of 8° to calculate the depth of the fault at different distances from the trench axis. We used a diffusivity of 1.1*10^-6 m²/s as used by Von Herzen et al. (2001). We used one (1) for the dimensionless factor $b$, because Molnar and England (1990) found this a reasonable number for most of the cases. The results are plotted in Figure 2, which shows a first-order match to the BSR-based derived heat flows in this region. Similar observations and full numerical simulations have been applied to subduction zones in other regions (cf. Hyndman and Wang, 1995; Wang et al., 1995; Currie et al., 2004).

Because the BSR-based and in situ heat flows are low in this region, we excluded the frictional heating term, i.e., we set the shear stress $\tau$ as zero. If we include the frictional heating term, a uniform stress of 25 MPa will increase the heat flow by 18 mW/m². We have also tested the cases of increasing stress with depth, but the higher frictional heating from deeper fault segments will increase the heat flow toward the arc, opposite to the decreasing trend derived from the BSR-based heat flows.

Near the collision zone to the north, we observed very high heat flows between the latitudes of 23.3°N and 24°N (Fig. 7), where there is no Benioff seismicity east of Taiwan island. Whether the slab is detached in this region is still under debate (e.g., Teng et al., 2000), and its influence on the heat flow pattern is still unknown.

Other tectonic events might have influenced the thermal structures in this region, including

Geological Society of America Bulletin, May/June 2008 687
sedimentation, because seismic profiles do not show strong evidence of simple shear-style thickening of the sediments arcward of the trench. However, is not very important for the Taiwan region.

We used the relation derived by Von Herzen and Uyeda (1963) to study the range of heat flow reduction due to sedimentation in this region:

$$\frac{Q^*}{Q} = 1 - \text{erf}(X) - \frac{2X}{\sqrt{\pi}} e^{-X^2} + 2X^2 \text{erfc}(X),$$

where $Q^*$ is the modified heat flow due to the sedimentation, $Q$ is the initial surface heat flow, $\alpha$ is the diffusivity,

$$X = 0.5 \times U \times \sqrt{\frac{t}{\alpha}},$$

and erf() and erfc() are the error function and its complement. The time ($t$) is in seconds. We use a thermal diffusivity of $2 \times 10^{-2} \text{ m}^2/\text{s}$. $U$ is the sedimentation rate.

Although no sedimentation rate data are available in this study region, a wide range of sedimentation rates (0.17–7.3 cm/ka) were derived from the nearby regions (cf. Table 8 of Chang, 2002). This translates to a 0.5%–18% reduction in heat flow based on our calculations. It is likely that the sedimentation effect is less important than the cooling effects from the subducting slab. The frictional heating and sedimentation calculations are assumed to increase and decrease the heat flow of the whole study region, respectively. Thus they can partially cancel out each other. In general, they will generate a “direct current” (DC)-like shift in heat flow profiles that could not explain the observed decreasing heat flow toward the arc in the subduction zone. However, if the sedimentation effect is small, the frictional heating must be small. Large frictional heating will not be consistent with the low heat flows observed in the offshore Taiwan region.

Opposite to the sedimentation processes reported in the offshore region, one of the highest erosion rates ever documented in the literature was found in the mountain region, on land, in the Taiwan collision zone (e.g., Dadson et al., 2003). These high erosion and exhumation rates will likely contribute to the high heat flow observed in the collision zone.

### Topographic Effects

Heat flow that escapes from the crust will take the shortest path to the low-temperature regions, in this case, the seafloor. The irregular bathymetry will cause nonvertical heat transfer, and the geothermal gradient increases under valleys while it decreases under the bathymetric highs in the upper slope domain. The dense heat flows derived from this study provide one of the first opportunities for us to directly study the topographic effects over a wide range of scales.

To quantitatively study the effects of the topography of the accretionary prism as a whole on the geothermal gradient, we have set up a finite-element numerical experiment using a transect in the subduction zone at the latitude of 20.5°N (i.e., the lower panel of Fig. 2). Preliminary results give more than 20 °C/km of perturbation from topography alone, which is consistent with the results from the analytical solution using a sinusoidal topography (cf. equation 4–66 of Turcotte and Schubert, 1982).

On a smaller scale, some of the local variations in the general trend of BSR subbottom depths is due to its occurrence in regions of folded strata. The curvature of the BSR is usually slightly less than that of the seafloor topography. A similar pattern is shown in seismic profiles from the western North Atlantic (Tucholke et al., 1977). Ganguly et al. (2000) have also quantitatively studied the topographic effects on 1–2 km scale based on BSR data from the Cascadia margin, and they found ~50% variation in heat flow due to topographic effect in that region. This local observation suggests a lower geothermal gradient on the crest than on the limbs of a fold, similar to the patterns from topographic effects. But this could also be influenced by the more intense upward fluid migration on the limbs, due to higher strain. In addition, note that many processes generating the topography, including thrusting and exhumation, will also affect the heat flow pattern.

We found strong correlations between high geothermal gradient and valleys over scales ranging from hundreds of kilometers to kilometers. Topographic effects, along with some processes generating the topography, can be important factors contributing to the observed heat flow pattern.

### Fluid Migration

Figure 5 shows higher geothermal gradient at the toe of the accretionary prism. Several studies have shown that sediment dewatering and presumably fluid flow are most extensive along the toe of accretionary prisms (Bray and Karig, 1985; Suess et al., 1998). Fluids are also released by dehydration reactions within sediments (Moore and Vrolijk, 1992). These fluids are forced to migrate upward toward the seafloor, advecting heat. Other evidence of overpressuring and fluid migration within the lower slope domain, especially in the collision zone, includes the presence of mud diapirs and volcanoes, on-land gas seepages, bright spots beneath BSRs and fault zones, and the small taper angle of the prism (Chi et al., 1998). This implies that the fluid is trapped in some locations and can only migrate in several specific locations like fault zones or diapirs, causing spatial variations in heat flow. Three possible hydrothermally active sites with BSR-based geothermal gradients exceeding 100 °C/km are identified in this region (Fig. 5).

In summary, the arcward-decreasing heat flow pattern in the Taiwan subduction zone can be best explained by the slab cooling effect. We also found evidence of topographic effects and focused fluid migration. It is less clear how much sedimentation and shear heating from the fault affect the heat flow pattern. On the other hand, the arcward-increasing heat flow in the collision zone has been previously modeled as a result of rapid exhumation. In addition, topography, groundwater migration, higher radiogenic heat, and some lower crust and upper mantle processes might also affect the heat flow pattern there.

### Implication for Other Geological and Geodynamical Processes

Zircon fission-track studies (Liu et al., 2001; Willet et al., 2003) in the collision zone show very fast cooling rates, some as large as 220 °C/m.y. for the past 2 m.y. Actually, Liu et al. (2001) found the width of the zircon reset zone, marked as a 260 °C isotherm, increases as the collision becomes more mature to the north. Although the increased cooling rate could be caused by increased exhumation rate, it could also be partially related to the northward increased geothermal gradient. In northern Taiwan where the fission-track studies were mainly conducted, the average geothermal gradient at shallow depth is slightly higher than the assumed 30 °C/km. The presumably higher thermal conductivity at greater depths might reduce the geothermal gradients at depth where the fission-track data sampled. However, the much higher geothermal gradients in southern Taiwan might cause overestimates of exhumation rates, while the lower geothermal gradients in the initial collision zone and subduction
zone might cause underestimates. The large variation of the geothermal gradient, from the upper slope domain in the subduction zone, to the Central Range in the collision zone, implies that the 30 °C/km assumption in calculating the uplifting rate might not be correct at all times even though it has been shown to be able to capture the overall evolution of thermal structure in this region. The geothermal gradients derived from this study might provide a closer approximation. However, even these spatially varying, geothermal gradients might not be representative at greater depths. New techniques and data sets are needed to better constrain the temperature field at depths.

Heated materials will have higher seismic attenuation. We found that the high geothermal gradient region derived from this study correlates with the high seismic attenuation regions proposed by Liu and Tsai (2005). The correlation is strongest in the southern part of the Central Range where geothermal gradient measurements in the mountainous regions are available. This suggests that this high geothermal gradient region can extend to at least mid-crust depth where the seismic waves have sampled.

The high heat flow in the collision zone will make the overall seismogenic zone on land shallower (Lin, 2002). The lower limits of the seismogenic zone in the convergent boundary is proposed to be defined by the 350 °C isotherm, which marks the change from velocity weakening to the velocity strengthening as the temperature increases (Tse and Rice, 1986). The slip distribution of the 1999 Chi-Chi, Taiwan earthquake on an east-dipping fault has a width of ~40 km (e.g., Chi et al., 2001). The deeper part of the mainshock slip terminated to the east at a region where the heat flow increases dramatically. No Chi-Chi aftershocks were found that ruptured into this region (Chi and Dreger, 2002; Chi and Dreger, 2004). Actually, very few earthquakes were found at greater depths in that region (Fig. 2). These results suggest that the high heat flow in some of the collision zone constrains the rupture extent of large earthquakes. On the other hand, the lower heat flow in the subduction zone implies that the seismicity in the offshore region probably can go deeper, thus increasing the width of the seismogenic zone.

CONCLUSIONS

Most of the mountainous regions around the world usually only have sparse heat flow data due to their rough terrain. This study is one of the first to document dense heat flow data from a region that covers both subduction and collision. In total, more than 1107 BSR-based heat flows were derived offshore southern Taiwan. We then combined this data set with previously published in situ heat flow measurements to study the possible thermal evolution from subduction to arc-continent collision.

In the subduction to the south, most of the BSR-derived heat flows range from 35 to 85 mW/m². The heat flow is high along the trench and decreases toward the arc. On the other hand, the in situ heat flow measurements in the collision zone mostly range from 70 to 250 mW/m². The heat flows are low near the frontal thrust, and increase, instead of decrease, toward the mountains.

The increased heat flow from subduction zone to collision zone results from increasing rock exhumation, reduced sedimentation and increased erosion, and topographically induced groundwater circulation after the upper slope domain emerged from the sea surface. Other geodynamic processes at depth might have also affected the thermal structures. The observed arcward-decreasing heat flow pattern in the subduction zone can be modeled with slab cooling, which shows ~50% reduction in heat flow. We found highest heat flows located above the regions without Benioff zones or near the edges of the slabs, which might relate to the lack of slab cooling. Other lower crust and upper mantle processes might also affect the thermal structures.

From this study, we found the thermal structures vary both in space, and very likely in time, when the crust evolves from subduction to collision. This dense data set depicts low heat flows in the subduction zone, suggesting low frictional heat on the décollement in this region. The increased geothermal gradient from subduction to collision along the bathymetric and topographic high will give faster exhumation rates for fission-track studies, if a constant geothermal gradient is used. The decreasing heat flow toward the subduction zone also implies that the seismogenic zone in the subduction zone in Taiwan can be wider and reach to greater depth. This dense and large heat flow data set might provide a unique opportunity to study some upper mantle, crust, and ocean thermal processes, in addition to P-T-t paths of metamorphic rocks in convergent boundary.

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