Heat sources in subduction zones: implications for slab seismicity and arc volcanism

Jason R. McKenna and David D. Blackwell
Department of Geological Sciences, Southern Methodist University, PO Box 750395, Dallas, Texas 75275
jmckenna@post.smu.edu, blackwel@passion.isem.smu.edu

Introduction

The shallow (upper 30–50 kilometers) thermal structure of subduction zones is well-constrained, and with few exceptions, can be modeled utilizing a variety of numerical [e.g., *Hyndman and Wang*, 1993] or even analytical solutions [e.g., *Molnar and England*, 1995]. It is stressed that the intermediate thermal structure of subduction zones remains largely unknown because the mechanism responsible for arc-magma genesis in the depth range of 80 to 120 kilometers remains subject to debate. Attempts to investigate the source region of intermediate-deep slab seismicity are likewise hampered by the poorly-constrained thermal structure at these depths. Constraints on the conditions on the subducting slab face are therefore necessary to any critical assessment of seismogenic conditions within the interior of the slab.

Most comprehensive thermal models have concluded that additional heat-mass transfer from outside the mantle wedge is necessary to explain the occurrence of arc volcanism and back-arc spreading [*Minear and Toksoz*, 1970; *Tatsumi and Eggins*, 1995]. Early thermal models by *McKenzie and Schlater* [1968], *Oxburgh and Turcotte* [1968; 1970] and *Minear and Toksoz* [1970] tested various heat sources to explain the occurrence of arc volcanism including conductive heating from a stagnant mantle wedge, exothermic phase changes, brittle shear-heating and viscous dissipation. All were ultimately rejected because they failed to satisfy the surface heat-flow over the subduction zone or generate the high-eruption temperature basalts observed in subduction zones (see *Davies and Stevenson* [1992] for a discussion).

However, *Thatcher and England* [1998] have demonstrated that significant ductile shearing localized in the lower crust/upper mantle along the San Andreas fault can explain the observed California Coast Ranges heat-flow anomaly. Because the viscosity of a ductile medium is temperature dependent, ductile shearing is rapidly concentrated in a narrow zone about the slip zone [*Yuen et al.*, 1978] and is virtually independent of slip velocity [*Thatcher and England*, 1998].

Thermal modeling results

The thermal structure of the shallowest 30 to 50 kilometers of the subduction zone can be modeled satisfactorily using the forearc surface heat-flow profile as the primary constraint [e.g., *Honda*, 1985; *Hyndman and Wang*, 1993]. However, recent work by *McKenna and Blackwell* [2002] has demonstrated that the shallow thermal regime of even young subduction zones is entirely conductive and can be successfully modeled without consideration of the deeper processes that influence the thermal structure of the subduction zone (i.e., arc magmatism and mantle wedge convection).

A thermal model specific to the subduction of the Juan de Fuca plate near 40°N (northern California) has been constructed assuming a subducting plate velocity of 40 mm/yr (the details of which will be presented elsewhere). Qualitatively, the model is comprised of moderately sedimented ~6 Ma Gorda lithosphere subducting beneath the North American backstop (composed of Mesozoic Franciscan complex and Klamath granite) and is overlain by a localized basin of Eel River sediments [e.g., *Smith et al.*, 1993]. This structure was extended eastward to the volcanic arc using crustal structure from *Mooney and Weaver* [1989]. This thermal model is typical of our estimates for the thermal structure in the upper ~40 kilometers of the southern Cascadia subduction zone system. Figure 1 illustrates the resultant temperature–depth curve along the subducting plate top for the initial model. Also presented is a temperature–depth curve representing the Sierra Nevada paleo-subduction zone to emphasize the difference between the cold subduction that occurred in Mesozoic North America (Sierra Nevada) and the much warmer thermal conditions in the present-day Cascadia subduction zone. The most interesting feature of the Cascadia curves is that temperature along the subducting slab face increases rapidly, attaining temperatures in excess of 450°C by the relatively shallow depth of five kilometers. As depth increases, the temperature along the subducting slab face increases only moderately.

The effect of viscous shear-heating on the subduc-

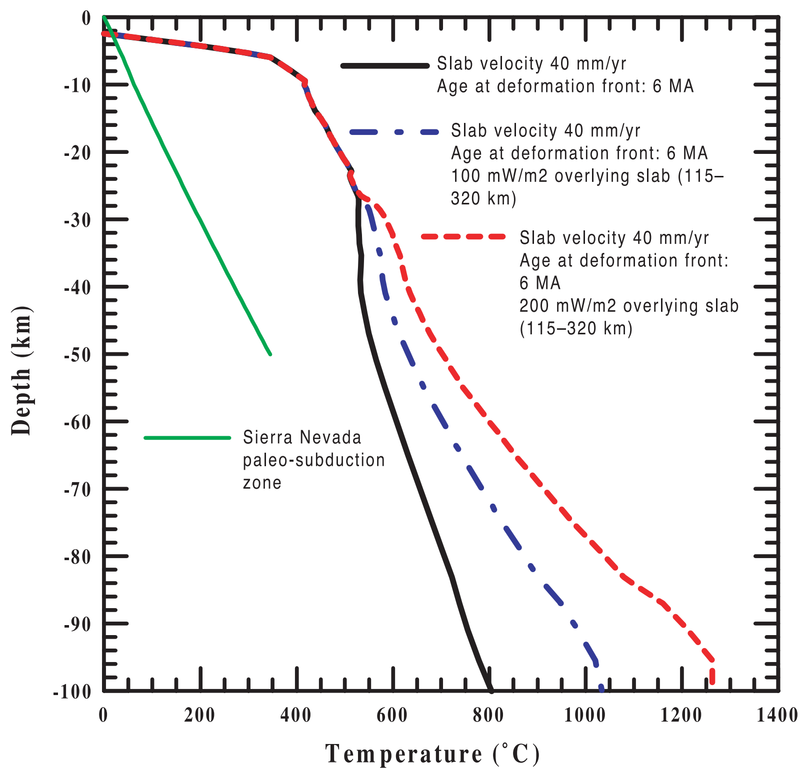

FIGURE 1: Temperature–depth curves along the subducting slab face, northern California. Note the high temperatures encountered along the slab face, even without the additional heat generated by viscous dissipation, when compared to the Sierra Nevada paleo-subduction zone.

tion zone thermal structure is simulated by specifying a heat source within a thin (two kilometers) boundary layer blanketing the slab face for a 200 kilometer horizontal extent (beginning about 115 kilometers landward of the deformation front). The strength of the additional heat source within the boundary layer is between 100 and 200 mW/m^2. We feel that while rather high, these values are not unreasonable estimates for the additional heat generated by viscous dissipation because in the presence of a non-linear (i.e., temperature-dependent) mantle viscosity, even moderate shear velocities can produce large temperature gradients across a thin boundary layer [*Thatcher and England*, 1998]. The resultant temperature–depth curves for the models that include the effects of viscous dissipation are presented in Figure 1. The point at which the heat source is "turned on" is visible as the bifurcation of the initial model curve.

Basalts are volumetrically the dominant lava type in the Cascades [*Blackwell et al.*, 1982; 1990]. Models that satisfy the surface heat-flow profile and still produce the requisite temperature (~1100°C) in the mantle wedge source region are considered successful. Utilizing this criteria, it is clear that only the curve representing the model with an additional heat input of 200 mW/m^2 begins to approach the temperature necessary for arc-basalts.

Figure 2 presents the results of the thermal calculations as pressure–temperature paths along the slab face. It has been suggested that the densification to eclogite via metamorphic dehydration reactions may be responsible for intermediate-depth intraslab seismicity [*Kirby et al.*, 1996; *Peacock and Wang*, 2002]. Again, we show the blueschist generating *P–T* conditions in the Sierra Nevada paleo-subduction zone to illustrate the distinct thermal conditions in present-day Cascadia. Here, even the initial model *P–T* path (no additional heat sources) almost reaches the eclogite stability field. Any additional heat introduced downdip through viscous dissipation promotes conditions in which the slab face readily transforms to eclogite by the relatively shallow depth of ~45 kilometers, about 100–150 kilometers west of the volcanic arc. We conclude from this analysis that seismicity below this depth is probably not related to the densification to eclogite, but most likely mechanically generated, and must therefore, be restricted to the cooler interior of the slab.

In addition to facilitating potentially seismogenic *P–T* conditions at shallow depths, the addition of viscous dissipation through a boundary layer overlying the subducting slab can induce local convective upwelling via boundary layer separation. In fluid dynamics, the Reynolds

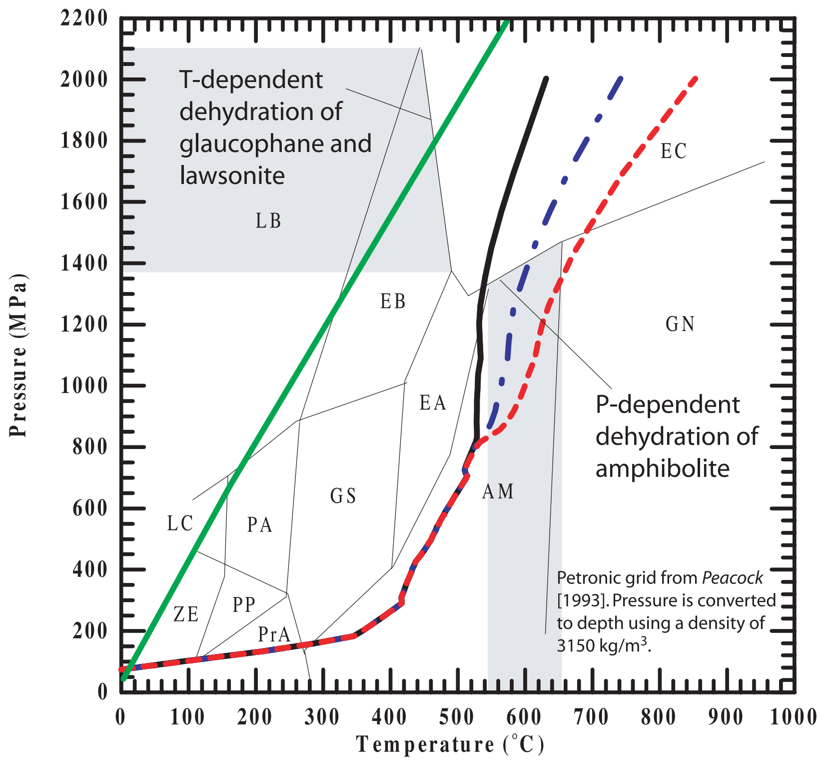

FIGURE 2: Pressure–temperature paths, northern California, curves are the same as in Figure 1. The potentially seismogenic phase transition to eclogite occurs around 45 kilometers depth, and about 100–150 kilometers west of the volcanic arc. If the densification to eclogite is largely completed by this depth, any observed seismicity is likely mechanically generated, and must therefore, be restricted to the cooler interior of the slab.

Number describes the stability of some particular flow field and is defined as $Re=UL/\nu$, where U is the free-stream velocity, L is the characteristic length of the flow, and ν is the kinematic viscosity. When $Re<<1$, viscous forces dominate the flow and the velocity is inversely proportional to (dynamic) viscosity. Furthermore, any inertial contributions to the flow may be neglected, greatly reducing the complexity of the analysis. Most, if not all, subduction zones can be characterized by an extremely small Re due to the initial large mantle viscosity, typically, $1x10^{19}$ or $1x10^{20}$ Pa·s in the uppermost mantle. For the northern California subduction zone, $U=40$ mm/yr, $\nu\sim1x10^{19}$ Pa·s, and $L=200$ kilometers, yielding a $Re\sim2x10^{-23}$.

Along the subducting slab, parameters change much faster normal to the slab than along it, so we may expect that a boundary layer of thickness d to be present along the slab. If we define our coordinate system so that the horizontal flow parallels the slab dip, the vertical component of velocity is normal to the slab and should be laminar, so long as: $d<<L$ and $Re<<1$. This is generally true if viscosity is constant. However, if the material properties are temperature-dependent, rapid departures from laminar flow may occur. We have determined the velocity in a boundary layer two kilometers thick parallel to the subducting Gorda slab in northern California. We allowed the viscosity to depend on temperature as follows: $\mu=\mu_o\cdot\exp[Q/RT]$, where $\mu_o=1x10^{19}\cdot\exp[-Q/R(T+1073.15)]$. Q, the activation energy is 522 kJ/m^3, R, the universal gas constant is 8.317 J/K·mol, and T is absolute temperature. Once the velocity field is known within the boundary layer, the point at which the boundary layer separates from the slab is simply where the shear-stress is zero on the slab: $(du/dz)_{slab}=0$.

Figure 3 illustrates the distance at which boundary layer separation occurs for the thermal models incorporating viscous dissipation. The initial thermal model does not attain the vertical velocity gradient reversal necessary for separation within the model geometry. In the model geometry, the position of the volcanic arc corresponds to a distance of 100 kilometers. It is clear that only the models with an additional heat source produce the upwelling beneath the position of the volcanic arc.

Conclusions

1. Reasonable heat source estimates (100–200 mW/m^2) in a thin boundary layer blanketing the subducting slab-face raises the ambient mantle temperature by at least

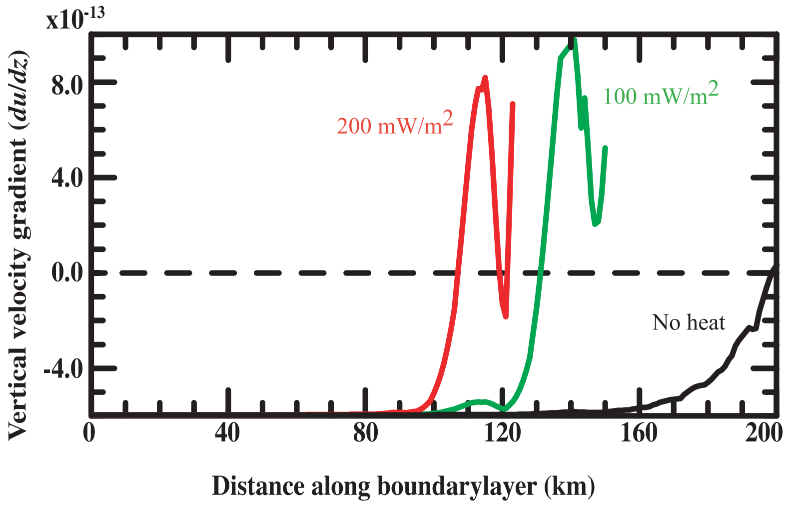

FIGURE 3: Boundary layer separation in the northern California subduction zone resulting from viscous dissipation. When the vertical velocity gradient reverses sign, the thin boundary layer that blankets the subducting slab begins to peel off from the slab, generating convective upwelling. Here, the separation point for 100 mW/m^2 is about 105 kilometers, while for 200 mW/m^2, it is closer to 125 kilometers. Both distances are almost directly below the volcanic arc, suggesting that boundary layer separation may contribute mass-flux to the volcanic arc system.

200–400°C beneath the volcanic arc. Viscous dissipation of heat produced in this layer can produce the minimum source temperature (>1100°C) necessary for the generation of the volumetrically significant basalts observed in the Cascades.

2. High temperatures along the subducting slab predict the seismogenic phase transition to eclogite at relatively shallow depths, even without viscous dissipation, resulting in little intraslab seismicity below ~45 kilometers in the subducting Gorda slab.
3. Because the high slab face temperature restricts seismicity in the upper portion of the slab, the subduction zone thermal structure is a boundary condition on the occurrence of intraslab seismicity that can aid in the interpretation of the subducting slab morphology.
4. The separation of the boundary layer from the slab is a direct result of the strong temperature-dependence of the viscosity and occurs approximately beneath the volcanic arc. This separation should produce local convective upwelling. However, the interaction of the separating boundary layer and any deeper, "corner-flow" remains uncertain. Future work will attempt to model this interaction and quantify the mass-flux off the slab in this region.

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