Thermo-mechanical Evolution of Oceanic Lithosphere: Implications for the Subduction Process and Deep Earthquakes

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Because subduction involves the return of cold oceanic lithosphere to the warmer mantle, much of our thinking about subduction reflects models of temperatures in subducting slabs. These models, in turn, rely on thermal models of the oceanic lithosphere before it subducts, developed using variations in ocean depth and heat flow with age. Such models predict that subducting slabs are colder, denser, and stronger than the surrounding mantle, in accord with evidence from seismic velocities and earthquake depths. Although the simple models describe the basic observable phenomena which reflect thermal structure, the variation in depth and heat flow between and among plates before they subduct, and in velocity structure and distribution of seismicity when they subduct, illustrates the need for improved models.

INTRODUCTION

Subduction zones are downgoing limbs of the mantle convection system, where slabs of cold oceanic lithosphere formed at midocean ridges return to the deep mantle. As discussed in this volume and reviewed elsewhere [e.g., *Kincaid*, 1995; *Kirby*, 1995; *Kirby et al.*, 1996a], many features of subduction reflect subducting slabs being much colder than surrounding mantle. Because slabs subduct rapidly compared to the time needed for heat conducted from the surrounding mantle to warm them up, they remain colder, denser, and mechanically stronger than the surrounding mantle. Consequently, slabs transmit seismic waves faster and with less attenuation than the surrounding mantle, making it possible to map slabs and to show that deep earthquakes occur within them. The negative thermal buoyancy of slabs should provide a major source of stress within downgoing slabs and appears to be the primary force driving plate motions. The cold slab should be mechanically stronger than its surroundings, and thus sustain higher stresses. This additional strength, and the fact that mineral reactions occur more slowly at lower temperatures, have been suggested as factors permitting earthquakes within slabs to occur to almost 700 km depth, far deeper than in the surrounding hotter mantle. Interaction between the slab and overlying mantle wedge gives rise to arc volcanism, apparently via metamorphism and dehydration of slab tops and partial melting and flow of the wedge.

As a result, considerable attention has been directed toward estimating temperature in subducting slabs. The temperature depends primarily on the temperature in plates when they enter the trench, and how the plates warm up as they descend. The temperatures, in turn, control the properties and behavior of slabs. Here, we review some basic observations and concepts that influence our thinking about these factors, and consider some of their implications for the subduction process.

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THE OCEANIC LITHOSPHERE

Recognition of seafloor spreading led rapidly to the view of plates of strong lithosphere moving over softer asthenosphere [e.g., *Elsasser*, 1971]. This rheological stratification reflects the thermal evolution of oceanic lithosphere, which cools as it spreads away from midocean ridges and reheats upon subduction into the deep mantle. In this view, which relies heavily on a set of observations reviewed next, oceanic lithosphere is the relatively thin and cold upper boundary layer of the mantle convective system (Plate 1), the primary mode of heat transfer from the earth's interior [e.g., *Parsons and Richter*, 1981; *Jarvis and Peltier*, 1989; *Pollack et al.*, 1993]. The lithosphere cools such that when it reaches most subduction zones, it is thought to be about 100 km thick with a basal temperature exceeding 1000°C. Below the lithosphere, temperatures are thought to increase more slowly, rising to only about 1500°C in the mantle transition zone (400-700 km depth) [e.g., *Ito and Katsura*, 1989]. Hence in the lithosphere, where heat transfer occurs primarily by conduction, temperature gradients are much higher (about 10°C/km) than below it, where lower (about 0.3°C/km) adiabatic temperature gradients are expected. Because of heat loss at the sea floor, subducting slabs are much colder than the surrounding mantle.

This temperature structure has major consequences. Because rock strength decreases with temperature, the oceanic lithosphere is also a mechanical boundary layer, which is stronger, i.e. can sustain greater stress, than material below (Figure 1). It arises because strength increases with pressure at shallow depths, where rocks fail by fracture, and decreases with temperature at greater depths, where rocks deform by temperature-dependent creep. Hence once temperature reaches about 800°C, lithosphere should be too weak to support significant stress [*Kirby*, 1977, 1983; *Goetze and Evans*, 1979; *Brace and Kohlstedt*, 1980]. As we will see, this idea is consistent with observations of quantities reflecting strength at depth. The strong lithosphere over a weaker asthenosphere allows the lithosphere to act as a stress guide for horizontal forces [*Elsasser*, 1969] and to sustain vertical loads [e.g., *Turcotte*, 1979], giving rise to familiar aspects of plate tectonics and plate boundary processes.

In addition to forming a thermal and mechanical boundary layer, differentiation at spreading centers causes the crust and uppermost mantle to form a chemical boundary layer [*Oxburgh and Parmentier*, 1977]. As a result, different definitions of "the" lithosphere are used for different purposes. Although the term strictly refers to material strength, it is often applied to the different boundary layers, in part because strength as a function of depth is not directly measurable, and in part because strength is temperature controlled. Thus the "thickness" of "the" lithosphere depends on the property (temperature, strength, chemistry) under consideration, and thus on the criterion or set of observations used to infer its variation with depth. The mechanical thickness inferred from the response to applied loads can also be strain-rate- and time-dependent [e.g. *Turcotte*, 1979; *Kirby*, 1983; *Stein et al.*, 1989].

It is worth bearing in mind that although many tectonic discussions focus on the lithosphere, it is the upper boundary layer of the convective system. Hence although density and strength variations largely reflect contrasts between the lithosphere and the remainder of the convective system, the total variations of each quantity due to convection drive plate motions and control the style of plate tectonics [e.g., *Verhoogen*, 1980].

THERMO-MECHANICAL STRUCTURE OF OCEANIC LITHOSPHERE

Data

Temperatures in subducting slabs are inferred from thermal models of the oceanic lithosphere before it subducts. Because temperatures at depth are not directly measurable, simple models are used, which attempt to provide a general description of average thermal structure as a function of age. The primary surface observables constraining these models are variations in seafloor depth and heat flow with lithospheric age (Table 1). Subsidence relative to the ridge crest, and hence seafloor depth, depends on temperature integrated with depth in the lithosphere, whereas heat flow depends on the temperature gradient just below the seafloor. A third observable is the variation with lithospheric age of the geoid, an equipotential of the gravity field, which reflects a depth-weighted integral of the density distribution, and hence provides a third constraint on the geotherm.

The observation that depth and heat flow vary approximately with the square root of lithospheric age (Figure 2) led to the view that young lithosphere acts largely as the upper boundary layer of a cooling halfspace [*Turcotte and Oxburgh*, 1967]. However, for ages older than about 70 Myr, average depth and heat flow "flatten", varying more slowly with age than for a halfspace. It is thus often assumed that halfspace cooling stops for older ages because heat added from below balances heat lost at the seafloor, causing the geotherm to approach steady state and thus the depths and heat flow to flatten. The plate

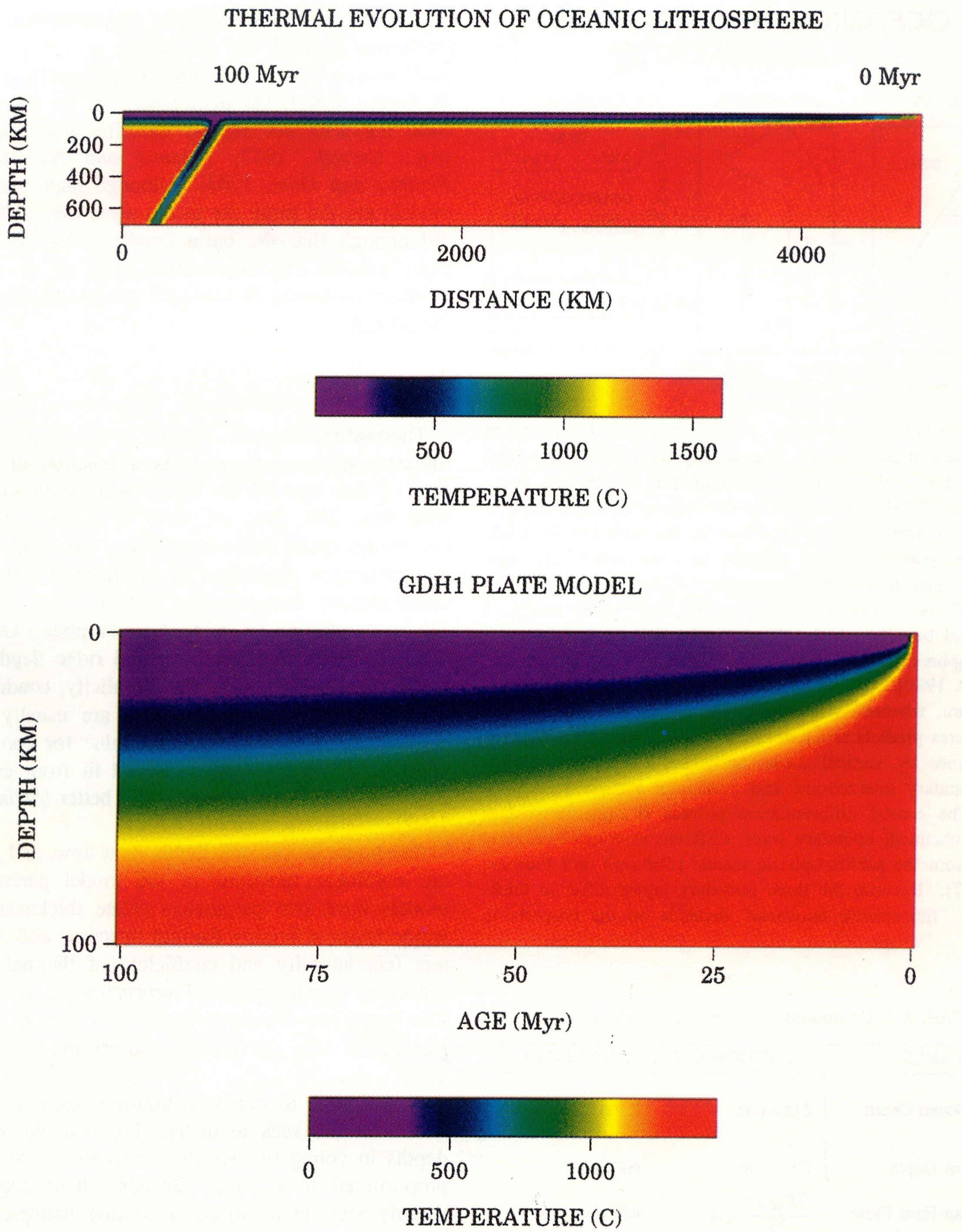

Plate 1. *Top:* Schematic illustration of the oceanic lithosphere as a thermal boundary layer, which cools as it moves away from midocean ridges and reheats as it subducts. Lithospheric temperatures are for the GDH1 thermal model [*Stein and Stein*, 1992] and a half-spreading rate of 4 cm/yr. Slab temperatures are from a finite difference calculation for a convergence rate of 8 cm/yr. Only lithospheric temperatures are calculated, so sublithospheric temperatures are shown as following an adiabatic gradient. *Bottom:* Thermal structure of the oceanic lithosphere for the GDH1 plate model.

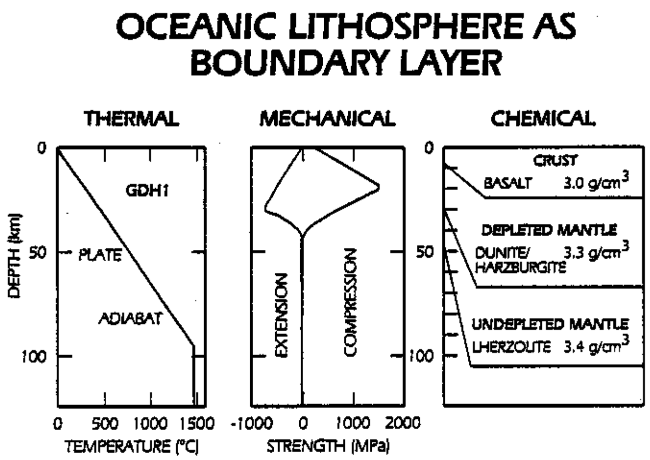

Fig. 1. The oceanic lithosphere forms a boundary layer, due to its thermal evolution. *Left:* The cooling plate forms a thermal boundary layer, illustrated by the asymptotic temperature structure for the GDH1 thermal model for lithosphere older than 70 Myr. The temperature structure controls the variations in depth, heat flow, gravity, seismic velocity and attenuation with age, and gives rise to the density variations causing plate driving forces. *Center:* The thermal boundary layer gives rise to a mechanical boundary layer, illustrated by a strength profile for old lithosphere, computed for a dry olivine flow law [*Brace and Kohlstedt*, 1980]. At shallow depth strength is controlled by brittle fracture, whereas at greater depth the ductile flow at high temperatures predicts rapid weakening. The strength profile controls flexure by vertical loads, horizontal stress transmission, plate boundary interactions, and maximum earthquake depths. *Right:* The crustal differentiation process at midocean ridges yields a chemical boundary layer, different in composition and density from the sublithospheric mantle [*Oxburgh and Parmentier*, 1977]. Because the three boundary layers differ in thickness, the "lithospheric thickness" depends on the property in question.

TABLE 1. Constraints on thermal models $T(z,t)$

OBSERVABLE	PROPORTIONAL TO	REFLECTS
Young Ocean Depth	$\int T(z,t)\,dz$	$k^{1/2}\alpha T_m$
Old Ocean Depth	$\int T(z,t)\,dz$	$\alpha T_m a$
Old Ocean Heat Flow	$\frac{\partial T(z,t)}{\partial z}\Big\vert_{z=0}$	kT_m / a
Geoid Slope	$\frac{\partial}{\partial t}\int zT(z,t)\,dz$	$k\alpha T_m \exp(-kt/a^2)$

T	temperature		
z	depth	t	age
a	plate thickness	T_m	basal temperature
α	thermal expansion coefficient	k	thermal conductivity

model, a simple description of this perturbation, uses an isothermal base of the lithosphere to model its thermal equilibration (Plate 1, Figure 3) [*Langseth et al.*, 1966; *McKenzie*, 1967]. The plate model fits the data reasonably well, but does not directly describe how heat is added [e.g., *Crough*, 1977; *Parsons and McKenzie*, 1978; *Fleitout and Doin*, 1994]. Although plate and halfspace models are the same for young ages, they differ for ages old enough that the basal condition has an effect. In plate models the lithosphere tends to an equilibrium geotherm, whereas in halfspace models cooling continues for all ages.

Thermal Models

Thermal models are solutions to the inverse problem of finding the temperature T as a function of age t and depth z that best fits the variation in depth and heat flow with age. The data are used to estimate the primary parameters (plate thickness a, basal temperature T_m, thermal expansion coefficient α, and thermal conductivity k) characterizing halfspace and plate models. (A halfspace can be considered an infinitely thick plate.) Other parameters (densities, specific heat, and ridge depth) are generally specified a priori. For simplicity, conductivity and coefficient of thermal expansion are usually treated as depth-independent. An a priori value for conductivity is often used, because the improved fit from estimating it from the data is not meaningfully better [*Stein and Stein*, 1992].

As shown in Table 1, depth, heat flow, and geoid slope are nonlinear functions of the model parameters. The models have two parameters (plate thickness and basal temperature) reflecting thermal structure, and two parameters (conductivity and coefficient of thermal expansion) reflecting average physical properties of the lithosphere. The latter two are scale factors which map the thermal parameters into the primary observable features of the data.

It is useful to consider limiting features of the data which models seek to match. The first, the slope of the depths in young lithosphere versus square root of age, is proportional to $k^{1/2}\alpha T_m$. Because these depths can be equally well fit assuming a cooling halfspace, they are insensitive to plate thickness. In contrast, the predicted behavior at old ages depends on plate thickness. The asymptotic depth for old ocean is proportional to $\alpha T_m a$, the heat lost as the plate cools. Similarly, the asymptotic heat flow for old ocean, $k\ T_m/a$, is proportional to the asymptotic linear geotherm. Hence in a plate model depth and heat flow tend to asymptotic values depending on

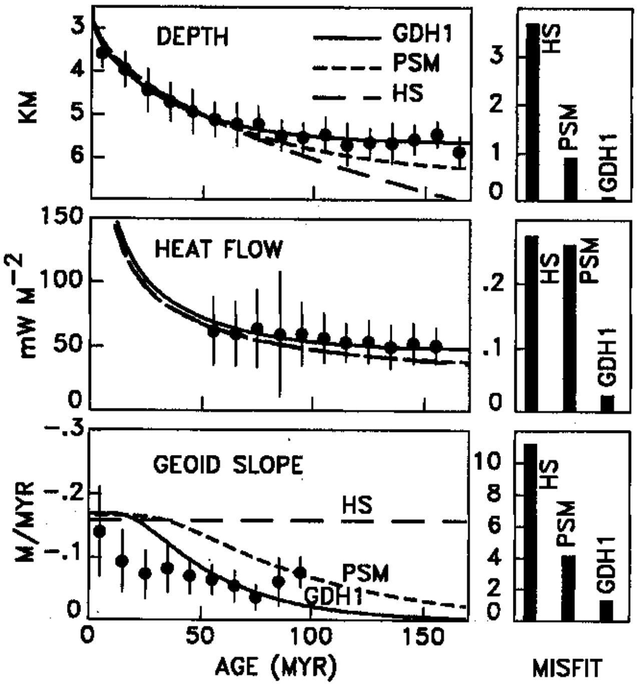

Fig. 2. Data used to constrain thermal models of the oceanic lithosphere. Comparison with the predictions of the three thermal models, which were not derived using these data, shows the general pattern that at older ages the data differ from the predictions of a cooling halfspace, and are better fit by plate models. Global depth data exclude hotspot swells [*Kido and Seno*, 1994]. Global heat flow are from *Stein and Stein* [1992], who used only North Pacific and Northwest Atlantic values to derive model GDH1. Geoid slopes across fracture zones are from *Richardson et al.* [1995]. The misfit values (*right panels*) show that the thin-plate GDH1 model fits better than either a halfspace (HS) or the thick-lithosphere PSM model.

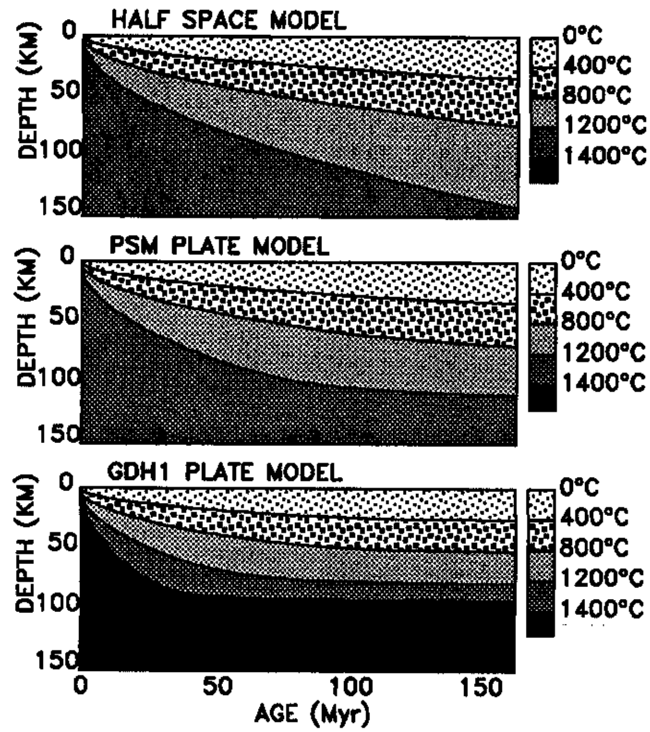

Fig. 3. Isotherms predicted by three thermal models. The lithosphere continues cooling for all ages for a halfspace model, reaches equilibrium for approximately 125 Myr old lithosphere in the thick-lithosphere (125 km) PSM model, and equilibrates for lithosphere about 70 Myr old in the thin (95 km) plate GDH1 model.

plate thickness and basal temperature, whereas in a halfspace model they continue to change with age. The derivative of the geoid with age (geoid slope) is also sensitive to the difference between models. It is constant for a halfspace model. For a plate model, the predicted slope is the same as for a halfspace at young ages, but "rolls off" at older ages at a rate depending inversely on plate thickness [*Cazenave*, 1984]. Like the "flattening" of depth and heat flow, this predicted deviation from halfspace behavior reflects the lithosphere approaching equilibrium thickness at older ages.

Figure 4 illustrates estimation of model parameters from data. The joint fit to depth and heat flow is shown as a function of assumed plate thickness and basal temperature. The best-fitting model, termed GDH1 (Plate 1, Table 2) [*Stein and Stein*, 1992], fits significantly better than either a halfspace (HS) model or a plate model with the parameters used by *Parsons and Sclater* [1977] (PSM). (The halfspace shown has parameters from *Carlson and Johnson* [1994] but the results would not differ significantly for other proposed parameters.) In particular, GDH1 reduces the systematic misfit to the depth and heat flow in older (>70 Myr) lithosphere (Figure 2), where PSM or a halfspace predict depths deeper and heat flow lower than generally observed.

The improved fit occurs because relative to PSM, GDH1 has thinner lithosphere with a higher basal temperature, and hence a steeper geotherm, higher heat flow, and shallower depths. An F-ratio test indicates the improved fit is significant at 99.9%. The improved fit going from PSM to GDH1 is comparable to that of PSM relative to a halfspace. Of the models, GDH1 fits geoid slope data best, though no model fits well.

The process of reestimating model parameters is conceptually the same as for global seismic velocity structure or relative plate motions. The goals are the same: to provide a better average description of the data and the pro-

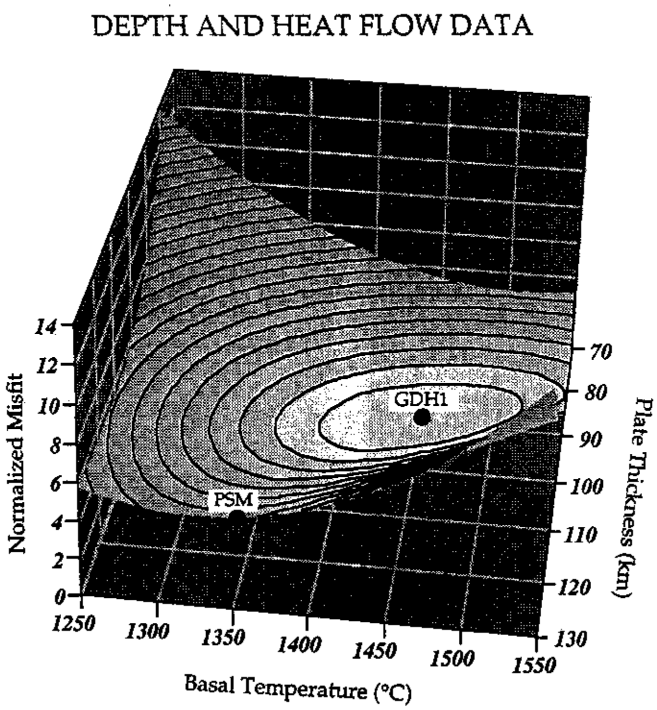

Fig. 4. Fitting process used for thermal model parameters. The misfit surface for the depth and heat flow data as a function of plate thermal thickness and basal temperature is shown, with the positions of the GDH1 and PSM models. Values are normalized to the GDH1 misfit, and the contour interval is 0.5. The misfit for PSM is five times that for GDH1. The surface is plotted for the GDH1 coefficient of thermal expansion, so PSM which has a different value plots slightly above the surface. The data and fitting function are discussed in *Stein and Stein* [1992].

cess causing it, and facilitate study of regions still poorly fit by the new, better-fitting, model. The improvements can be significant; GDH1 fits about five times better than PSM, compared to the plate motion case where the recent NUVEL-1 model [*DeMets et al.*, 1990] gives a factor of three improvement over the earlier RM2 model [*Minster and Jordan*, 1978].

The resulting model should better describe the average thermal state of oceanic lithosphere. In addition, because the model better fits the data, it makes it easier to assess which regions are "anomalous", in that their depths and heat flow differ from most lithosphere of that age. A difficulty with using a halfspace or a thick plate as reference models is that because they systematically mispredict depth and heat flow for old lithosphere, almost any old lithosphere appears "anomalous" relative to these models, although it is normal for old lithosphere. It is thus harder to assess which areas differ from average

TABLE 2. GDH1 model parameters

a	plate thickness	95 km
T_m	basal temperature	1450 °C
α	thermal expansion coefficient	3.1×10^{-5} °C^{-1}
k	thermal conductivity	3.138 W m^{-1} °C^{-1}
C_p	specific heat	1.171 kJ kg^{-1} °C^{-1}
ρ_m	mantle density	3330 kg m^{-3}
ρ_w	water density	1000 kg m^{-3}
d_r	ridge depth	2600 m

old lithosphere [*Stein and Stein*, 1993]. This situation is reminiscent of the mythical town of Lake Wobegon in the radio show *Prairie Home Companion*, "where all children are above average".*

Assessment of Models

Several points about such models are worth noting. Although the models fit data reasonably well, clear misfits remain, which presumably reflect both processes acting in addition to those incorporated in the simple thermal models, and variability between and within plates. For example, heat flow for ages younger than about 65 Myr is lower than predicted, presumably due to hydrothermal circulation [e.g., *Wolery and Sleep*, 1976; *Stein and Stein*, 1994a] (The extent to which this circulation affects temperatures is unknown, but may not be significant except very close to the ridge axis, [e.g., *Stein et al.*, 1995], so it is assumed to not affect depths significantly). The misfit to geoid slope data for young ages may reflect the geoid offset across fracture zones incorporating effects in addition to that purely of the thermal age contrast, such as flexure, thermal stresses, or local asthenospheric flow [*Sandwell*, 1984; *Parmentier and Haxby*, 1986; *Robinson et al.*, 1988; *Wessel and Haxby*, 1990]. Moreover, the thermal models describe only average temperature structure as a function of age, using a few depth- and age-independent parameters. Hence these models are simple representations of a complex thermal structure which do not address the variations in depth, heat flow, and geoid slope as functions of age between and within plates [e.g., *Calcagno and Cazenave,* 1994] which may reflect both variations in temperature and other perturbations such as those due to intraplate volcanism, crustal thickness, or asthenospheric flow.

Because these models are solutions to an inverse problem, we can assess how they fit data, but have no direct

*Similarly, S. Peacock pointed out to us in his review that 90% of motorists are said to consider themselves above-average drivers.

way of telling how well they describe temperature in the earth. Some insight can be derived by using models to predict data not used in deriving them. For example, GDH1 fits depth and heat flow data that were not inverted in deriving it (Figure 2) better than PSM or a halfspace model [*Johnson and Carlson*, 1992; *Stein and Stein*, 1993; *Shoberg et al.*, 1993; *Kido and Seno*, 1994]. Similarly GDH1, derived by inverting depth and heat flow data, predicts geoid data better than the other models. Nonetheless, GDH1 or any simple model does not fully describe the thermal structure, as illustrated by the misfit to the geoid data at young ages or the variations in depth, heat flow, and geoid slope as functions of age between and within plates.

Estimation of thermal structure of the lithosphere thus faces difficulties common to inverse problems. The models are oversimplifications of the real situation, and even for a given model, the parameters estimated depend on the choice of data and fitting function, and are nonunique [*Stein and Stein*, 1992, 1993]. The usual question arises when more complicated models better fit data, whether the improved fit exceeds that expected purely by chance from the model's having more free parameters [*Stein and Stein*, 1992, 1993, 1994b]. Similarly, there is the issue of how best to incorporate other information. For example, should parameters like the average coefficient of thermal expansion be determined from inversion, specified a priori from extrapolation of laboratory results, or estimated by combining these approaches? This question is illustrated by the observation that the GDH1 basal temperature is slightly (7%) higher than the approximately 1350°C often inferred for the temperature of midocean ridges from the thickness of oceanic crust [e.g., *Sleep and Windley*, 1982; *McKenzie and Bickle*, 1988]. However, the improved fit of GDH1 over a model with basal temperature fixed at 1350°C is significant as measured by F-ratio test [*Stein and Stein*, 1993]. If the ridge temperature is known to sufficient precision (a question beyond our scope here), the discrepancy could have several causes. For example, although the thermal model uses a single basal temperature for all ages, the estimate of this parameter (rather than its product with the coefficient of thermal expansion) depends largely on data for old ages. Hence the discrepancy may reflect heat addition to old lithosphere by a process analogous to mantle plumes, which are thought to be several hundred degrees hotter than ridges [*Sleep*, 1992].

Investigation of the thermal evolution of oceanic lithosphere remains an active research area. Even the basic question of whether old lithosphere approaches thermal equilibrium is still under discussion [e.g., *Carlson and Johnson*, 1994]. Although the most straightforward interpretation of the depth, heat flow, and geoid data is in terms of a plate model [*Richardson et al.*, 1995], other interpretations are possible. In particular, explanations other than thermal equilibration have been offered for flattening of the depth curve. In one, flattening is analogous to that associated with seafloor traces of mantle plumes [e.g., *Heestand and Crough*, 1981; *Davies and Pribac*, 1993], and so is due largely to dynamic pressure of plumes, with some heating of the lithosphere. Another possibility is that apparent flattening results from excess volcanism masking continued subsidence due to halfspace cooling. A third possibility is that depths are perturbed by pressure differences driving asthenospheric flow [*Schubert and Turcotte*, 1972; *Schubert et al.*, 1978; *Phipps Morgan and Smith*, 1992, 1994; *Stein and Stein*, 1994b]. Moreover, even if flattening is a thermal effect, the physical process of heat addition has yet to be satisfactorily explained. These questions remain unresolved because all proposed perturbations are at least qualitatively consistent with the observed flattening, because it is difficult to isolate the effects of possible different mechanisms, and because flattening differs enough in different locations [*Calcagno and Cazenave*, 1994] that multiple mechanisms may operate.

Implications for Subduction Zone Thermal Structure

Fortunately, for many subduction zone applications, the choice of temperature model is not crucial. For example, the model predictions in Figure 3 are similar, especially at shallow depths (consider the 400°C isotherm). The differences between models are, however, of possible significance for the common case of the subduction of old lithosphere. As discussed shortly, various subduction zone phenomena seem vary with the age of the subducting plate, and hence presumably its temperature structure. Thus where old lithosphere subducts, it matters somewhat whether we consider a plate model (in which all lithosphere older than about 70 Myr is similar) or a halfspace model (in which temperatures still vary with age for old lithosphere). The choice of model matters largely for the deepest portions of the lithosphere.

Mechanical Structure

The mechanical structure of the oceanic lithosphere before it subducts also has implications for the subduction process. The primary determinant of mechanical structure (Figure 1) is variation in strength with depth and age resulting from the temperature and pressure. A secondary factor is the presence of structural heterogeneities.

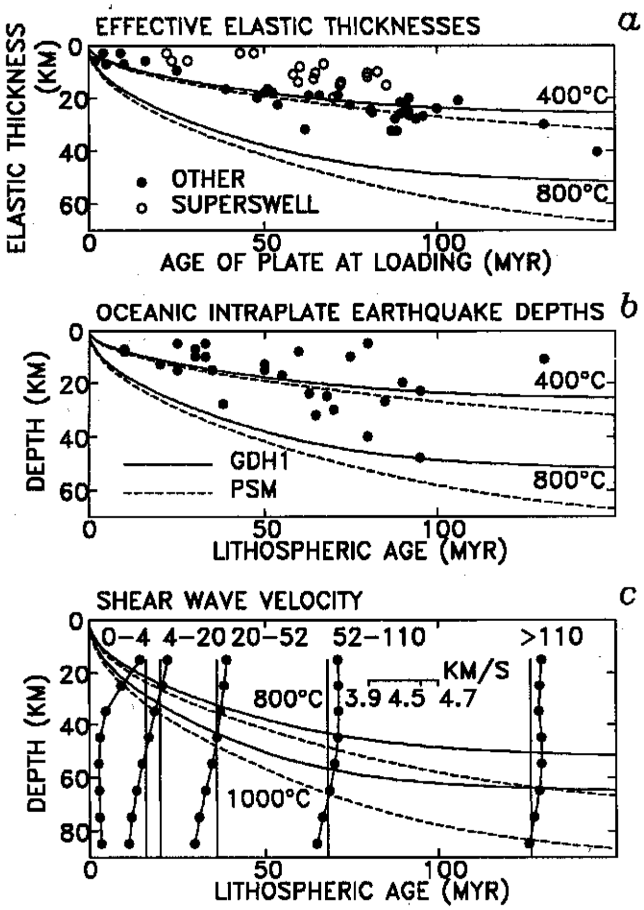

Fig. 5. Other data types whose variation with age is consistent with cooling of the lithosphere, as illustrated by isotherms for two thermal models. Except for the oldest lithosphere, the isotherms corresponding to the effective elastic thickness data [*Calmant et al.*, 1990] *(a)* and deepest intraplate seismicity [*Wiens and Stein*, 1983] *(b)* are similar for the two models. The difference between the temperatures for the low velocity zone *(c)* is greater. The vertical line for the velocity structure [*Nishimura and Forsyth*, 1989] in each age range (e.g., 0-4 Ma) corresponds to 4.5 km s^{-1}.

The predicted weakening of the lithosphere below a temperature-controlled depth is consistent with several observations (Figure 5). Effective elastic thickness inferred from loads on the lithosphere, a measure of the depth to which the lithosphere is strong enough to support significant stress, increases with age approximately as the 400°C isotherm [e.g., *Bodine et al.*, 1981; *Calmant et al.*, 1990]. The maximum depth of intraplate seismicity, which presumably reflects the depth to which the lithosphere is strong enough to support seismogenic stresses, increases with age approximately as the 700°C isotherm [*Wiens and Stein*, 1983; *Chen and Molnar*, 1983]. Similarly, the depth to the low velocity zone inferred from seismic surface wave dispersion increases with age [e.g., *Nishimura and Forsyth*, 1989]. Although these observations appear to reflect lithospheric cooling, using them to discriminate between thermal models is difficult, as it requires rheological models, assumptions about the strength required to support specific loads, and assumptions about the variation in seismic velocity with temperature. For subduction zone considerations, however, they jointly indicate that the upper portion of the slab should be strongest, and that the geometry of this strong region should be temperature- and pressure- controlled as the slab subducts.

In addition to the oceanic lithosphere being weak at depth, tectonic features may cause weak zones. Oceanic intraplate seismicity occurs preferentially on no-longer-active tectonic features, such as fossil spreading ridges and hotspot tracks, suggesting that these zones are weaker than normal lithosphere and move easily in response to intraplate stress [e.g., *Stein and Okal*, 1978; *Stein*, 1979; *Bergman and Solomon*, 1980; *Geller et al.*, 1983; *Wysession et al.*, 1991, 1995]. Similarly, unusual seismicity occurs where seafloor tectonic features enter trenches [e.g., *Vogt et al.*, 1976; *Chung and Kanamori*, 1978; *Stein et al.*, 1982]. In addition, earthquakes, some of which are large, occur in the subducting plate as it approaches the trench [e.g., *Kanamori*, 1971; *Chapple and Forsyth*, 1979], perhaps due to plate bending. Thus preexisting faults, also including those remaining from near-ridge processes, may survive as weak zones once the plate subducts and be the loci of intermediate [*Kirby et al.*, 1996b] and deep earthquakes [*Silver et al.*, 1995]. It thus seems that some of the variation in seismicity along subduction zones depends on weakness in the lithosphere before it subducts.

THERMAL STRUCTURE OF SUBDUCTING SLABS

Models

Predicting temperatures in subducting slabs is more challenging than in lithosphere before it subducts, because temperatures are not only unconstrained by direct observations, but less easily inferred indirectly. Hence caveats raised earlier about thermal models of the lithosphere apply even more strongly. Fortunately, the basic ideas about slab temperatures from simple models are relatively insensitive to the details of the model.

A simple analytical model, based on the time required for a slab to heat up by conduction as it subducts into a hotter isothermal mantle [*McKenzie*, 1969], illustrates several ideas. Isotherms in the slab extend downward, such that the maximum depth reached by a isotherm is proportional to the product of the vertical descent rate (trench-normal convergence rate times the sine of the dip) and the square of plate thickness. Hence for a halfspace thermal model, in which the thickness of the subducting lithosphere is proportional to the square root of its age at the trench, the depth reached by an isotherm is proportional to the product of the vertical descent rate and age, a quantity known as the thermal parameter.

Similar results emerge from numerical thermal models. Plate 2 shows thermal models from a program derived from one by N. H. Sleep, based on a finite difference algorithm [*Toksöz et al.*, 1971] widely used in slab thermal modeling [e.g., *Sleep*, 1973; *Hsui and Toksöz*, 1979] The temperature structure of the lithosphere before it subducts and the thermal diffusivity are from GDH1. In such cooling plate models, temperatures approach steady state at about 70 Myr, such that all older lithosphere has about the same geotherm. Hence the geotherm in subducting lithosphere does not vary directly with thermal parameter, as would be true for a halfspace thermal model. For simplicity, lithosphere entering the trench is assumed to be of constant age. Models are computed by allowing subduction to go on long enough that a stable temperature structure results. The models shown are for a relatively younger and slower-subducting slab (thermal parameter about 2500 km), approximating the Aleutian arc, and an older and faster-subducting slab (thermal parameter approximately 17,000 km), approximating the Tonga arc. As expected, the slab with higher thermal parameter warms up more slowly, and is thus colder. These predicted temperature distributions are typical of slab thermal models, though the depth to individual isotherms vary, as shown by comparison of various models, including those listed by *Helffrich et al.* [1989] and a more recent study [*Davies and Stevenson*, 1992].

The thermal models also give insight into the rate at which slabs should equilibrate with the mantle. Figure 6 shows the predicted minimum temperature within a slab as a function of time since subduction. The coldest portion reaches half the mantle temperature in about 10 Myr, and 80% in about 40 Myr. Hence because most subduction zones have been active far longer than the time (about 10 Myr) required for the slab to first reach 670 km [e.g., *Engebretson et al.*, 1992], the maximum depth of earthquakes in each subduction zone does not simply indicate the maximum depth that the slab has reached, a possibility suggested by *Isacks et al.* [1968] before an adequate magnetic anomaly record became available.

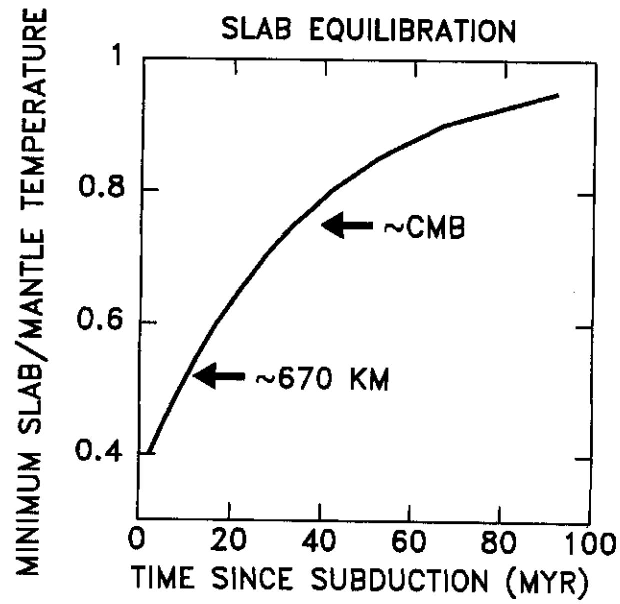

Fig. 6. Minimum temperature within a slab as a fraction of the mantle temperature, as a function of the time since subduction, computed using the analytic model of *McKenzie* [1969] for a slab with GDH1 model parameters. The coldest portion reaches half the mantle temperature in about 10 Myr, by which a typical slab is approximately at 670 km depth, and 80% in 40 Myr, by which a slab which continued descending at the same rate would reach the core-mantle boundary. Slabs can thus remain thermally distinct for long periods of time.

Moreover, because the time required for subducted material to reach 670 km depth is far less than that required for thermal equilibration with the surroundings, the restriction of seismicity to depths shallower than 670 km does not indicate that the slab is no longer a discrete thermal, and thus mechanical, entity. Hence from a thermo-rheological standpoint, there is no reason for slabs not to penetrate into the lower mantle, in accord with seismological observations and convection modeling, discussed shortly. In fact, if a slab descended through the lower mantle at the same rate, it would retain a significant thermal anomaly at the core-mantle boundary. As a result, cold slab remnants in the lower mantle are thought to give rise to thermal and thus density heterogeneity [e.g., *Richards and Engebretson*, 1992].

Such thermal models are used to predict approximate temperatures within slabs. Clearly the geometry assumed is simplified and the uniform thermal diffusivity adopted is an approximation. As the model shown only allows slab heating by conduction from the surrounding mantle and does not include shear or radiogenic heating or latent heat release, the predicted temperatures are lower bounds.

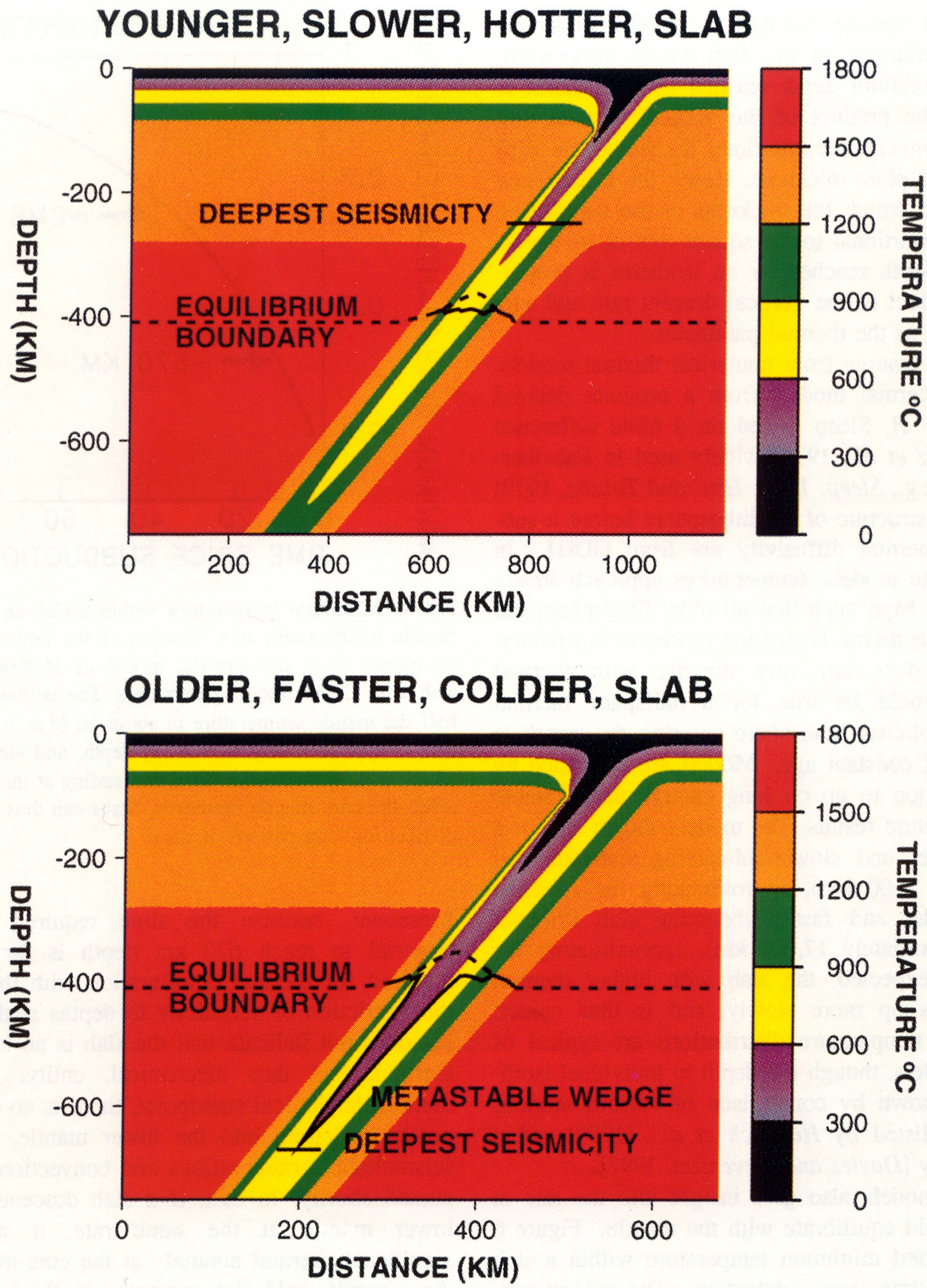

Plate 2. Comparison of thermal structure and predicted regions of metastability for a relatively younger and slower-subducting slab (thermal parameter about 2500 km), which approximates the Aleutian arc, and an older and faster-subducting slab (thermal parameter about 17,000 km) which approximates the Tonga arc. As expected, the slab with higher thermal parameter warms up more slowly, and is thus colder. A metastable wedge forms only for old and fast-subducting slabs which are sufficiently cold that kinetic hindrance prevents phase transformations from keeping pace with the descent rate. For the model parameters used, the metastable wedge is bounded on its sides and bottom by the 600°C isotherm, in those slabs cold enough for it to form. Deep earthquakes are presumed to occur by transformational faulting in the metastable wedges.

Hence these (or other) models' predicted temperatures are probably not accurate to better than about 200°C. The fact that the temperatures from such models predict seismic velocities similar to those inferred from observations (Plate 3) suggests that the models are at least reasonable approximations. As a result, it seems plausible to use such models to explore subduction zone processes.

Implications for Deep Earthquakes

As papers in this volume illustrate, thermal models of subducting slabs are used to study aspects of the subduction process [e.g. *Kirby et al.*, 1996b; *Peacock*, 1996]. The models are used in studies which characterize subduction zones by various parameters, such as the convergence rate and age of the subducting lithosphere, and investigate how processes vary among subduction zones [e.g., *Jarrard*, 1986; *Peacock*, 1992; *Davies and Stevenson*, 1992]. One striking example is the variation in the maximum depth of deep earthquakes (those below 325 km) as a function of thermal parameter (Figure 7). Although earthquakes are restricted to depths shallower than about 680 km, the maximum depth increases with thermal parameter. This observation argues for the maximum depth of earthquakes being controlled by a temperature-dependent mechanism.

No consensus, however, exists about what the thermal control mechanism may be. One possibility is that seismicity is limited by a thermally-controlled strength, such that at higher temperatures the slab is too weak to support seismic failure [*Molnar et al.*, 1979; *Wortel*, 1982; *Wortel and Vlaar*, 1988]. A difficulty, however, is that laboratory results predict that slabs should be strong well below the deepest earthquakes [*Brodholt and Stein*, 1988].

A second possibility is that faulting occurs by brittle fracture as for shallow earthquakes. Although high pressures would normally suppress fracture, it may occur once slabs become hot enough that water released by decomposition of hydrous minerals reduces effective stress on fossil faults formed before subduction [*Raleigh*, 1967; *Meade and Jeanloz*, 1991; *Silver et al.*, 1995]. Because shallow earthquakes in oceanic lithosphere only occur where the temperature is less than approximately 800°C [*Wiens and Stein*, 1983; *Chen and Molnar*, 1983], similar temperature control would be expected in the slab [*Stein*, 1995]. A possible problem with this dehydration embrittlement model is that hydrothermal circulation would not be expected to bring water to depths of more than a few km in oceanic plates [e.g., *Stein et al.*, 1995], whereas large earthquakes would require water in the cold interior of slabs. Moreover, it is unclear whether the hydrated minerals could survive to these depths [*Ulmer et al.*, 1994].

In the third hypothesis, deep earthquakes result from solid state phase changes, primarily that in which olivine transforms to a denser spinel structure [*Kirby*, 1987; *Green and Burnley*, 1989; *Kirby et al.*, 1991]. Thus deep seismicity occurs only in the depth range of the mantle transition zone, where phase changes should occur in downgoing slabs. Because the rate of the phase transformation depends exponentially on temperature, then in fast-subducting cold slabs the transformation cannot keep pace with the descent, and metastable olivine should persist below the equilibrium phase boundary [*Sung and Burns*, 1976ab; *Rubie and Ross*, 1994; *Kirby et al.*, 1996a] (Plate 2). Deep earthquakes are assumed to occur in the metastable material by a shear instability, known as transformational faulting, observed in the laboratory when metastable materials under stress undergo strongly exothermic reactions that isochemically transform one phase into a denser form. This mechanism resolves the objection traditionally raised to phase-change models for deep earthquakes, because the resulting motion would be slip on a fault rather than an implosion, in accord with seismological observations [*Kawakatsu*, 1991].

The metastability hypothesis makes several predictions generally consistent with various observations. The idea that deep earthquakes occur by a failure mechanism different from that for shallow and intermediate earthquakes is tempting, because seismicity as a function of depth has a minimum at about 350 km and then increases, suggesting deep earthquakes form a distinct population. The idea of deep earthquakes due to phase changes explains why these earthquakes coincide with the 400-700 depth range of the transition zone, where these phase changes are expected. In particular, it explains why deep earthquakes cease at the base of the transition zone, because phase changes associated with formation of the lower mantle mineral assemblage are endothermic and thus should not cause transformational faulting. This idea is significant for mantle dynamics, in that although the simplest explanation for the cessation of seismicity near 670 km is that slabs do not descend into the lower mantle, seismological observations are interpreted as indicating that some do [e.g., *Fischer et al.*, 1988; *Van der Hilst et al.*, 1991; *Fukao et al.*, 1992; *Van der Hilst*, 1995]. Such slab penetration is also predicted by mantle convection modeling [e.g., *Christensen and Yuen*, 1984].

Metastability may explain the observation (Figure 7) that deep seismicity occurs only for slabs with thermal parameter greater than 5000 km. Rapid deepening is hard

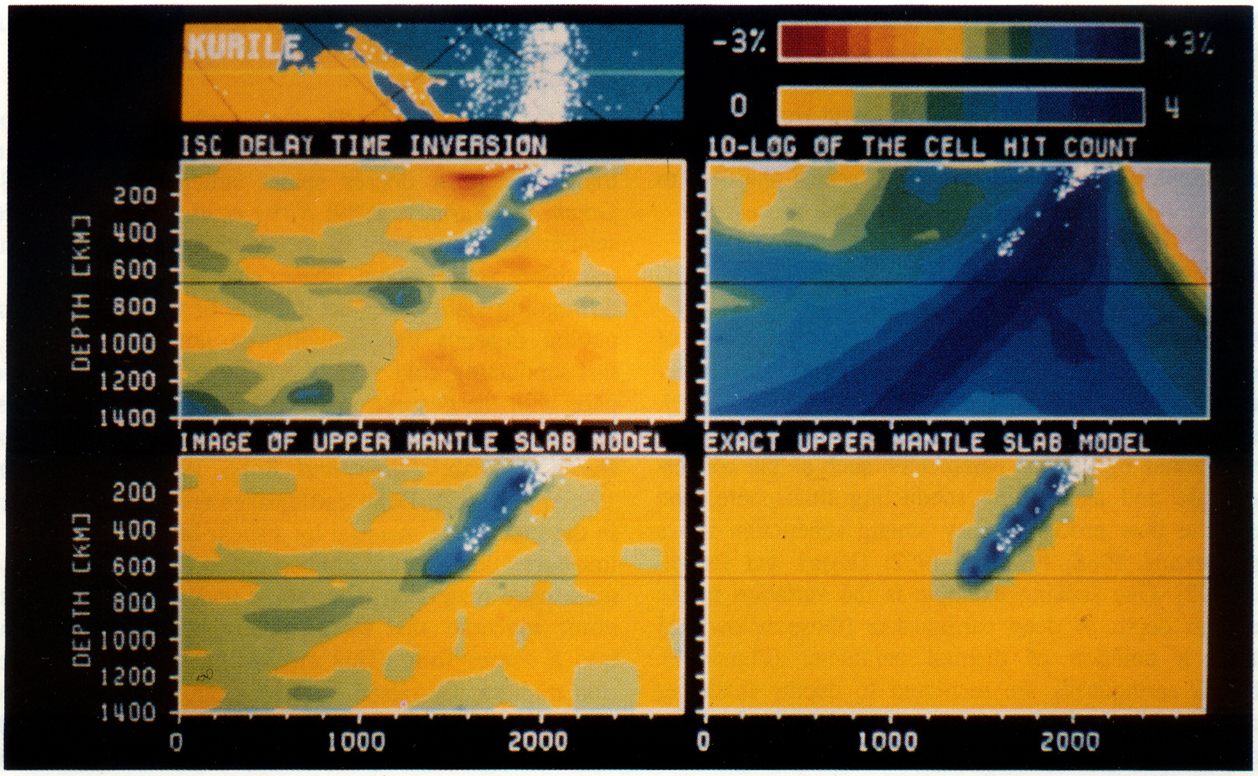

Plate 3. Comparison of seismic (P-wave) tomographic image of the subducting Kurile slab obtained from inversion of International Seismological Center delay time data (*center left*) to the image (*lower left*) predicted for a slab thermal model. The seismic velocity anomaly predicted by the thermal model (*lower right*) is imaged by a simulated tomographic study with the same seismic ray path sampling as used for the data. The ray path sampling is shown by the hit count (*center right*), the number of rays sampling each cell used in the inversion. White dots indicate earthquake hypocenters. As a result of ray geometry and noise added to the synthetic data, the exact model of the slab gives a somewhat distorted image (*lower left*), showing how the model would appear in such a tomographic study. The fact that the image of the slab model and the tomographic result (*center left*) are similar suggests that the slab model is a reasonable description of the major features of the actual slab. A similar conclusion emerges from the observation that the tomographic result also resembles parts of the model image which are resolution artifacts not present in the original model. These artifacts, generally of low amplitude, cause the slab to appear to broaden, shallow in dip, or "finger". [*Spakman et al.*, 1989].

to explain if seismicity is controlled directly by temperature, because temperatures vary smoothly as a function of thermal parameter. It is easier to explain as a consequence of metastability, as illustrated by the wedge-shaped region of predicted metastability in Plate 2. This region is delineated above by the equilibrium boundary for the olivine → spinel transition, which is elevated in the cold slab, and below by the 99% transformation contour where almost all metastable olivine has transformed. The younger, slower-subducting, and hence warmer slab is hot enough that phase transformation keeps pace with subduction, essentially no metastable wedge forms, transformational faulting should not occur, and deep earthquakes are not expected. In contrast, the older faster-subducting slab is cold enough that phase transformation cannot keep pace with subduction, a distinct metastable wedge forms, and deep earthquakes are expected. For the model parameters used, the wedge is bounded on its sides and bottom by the 600°C isotherm, so the occurrence of deep earthquakes is temperature-controlled in slabs cold enough for a metastable wedge to form.

Problem: a Fault too big?

Recent large deep earthquakes illustrate a potential difficulty with all three hypotheses for deep earthquakes. In these hypotheses, earthquakes should be restricted to the portion of the slab cooler than 600-800°C, which ther-

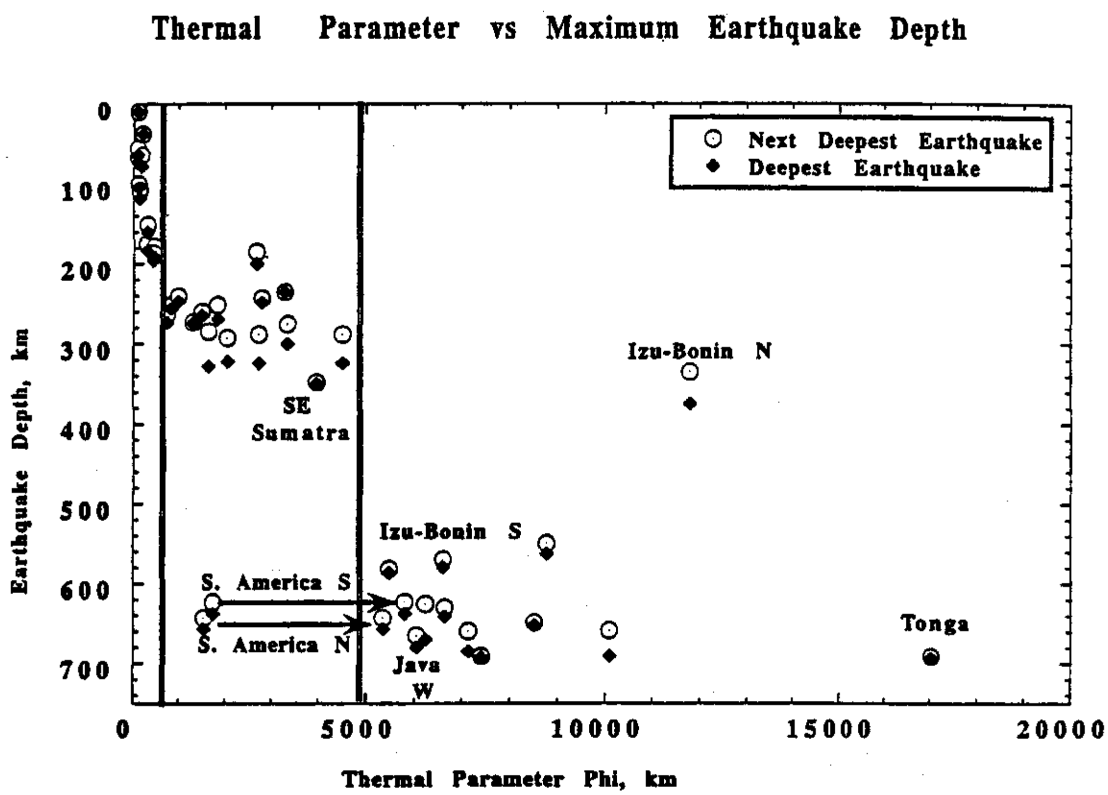

Fig. 7. Plot of earthquake depths for different subduction zones [*Kirby et al.*, 1996a]. The plot is a function of thermal parameter, the product of vertical descent rate and lithospheric age. For simple thermal models, the maximum depth to an isotherm should vary with the thermal parameter. Hence if deep earthquakes were directly temperature limited, their maximum depth should be a smooth function of thermal parameter. Instead, the maximum depths seem divided into a group with thermal parameter less than about 5000 km, which do not have deep earthquakes, and a group with greater thermal parameters that do. This abrupt change is consistent with the predictions of the thermo-kinetic model in Plate 2 where deep earthquakes result from phase changes in metastable olivine, and so occur only in slabs where significant metastability is expected.

mal models predict should be a wedge narrowing to less than about 10 km at depths greater than 600 km. The fault areas of recent deep earthquakes, however, exceed the predicted wedge dimensions. A magnitude 7.6 earthquake beneath Tonga in 1994 had an usually large aftershock sequence, which defined a 50 x 65 km fault zone [*Wiens et al.*, 1994]. Only a few months later, the largest deep earthquake instrumentally recorded occurred beneath Bolivia. Analyses of data for this earthquake indicate a near-horizontal fault area 30-50 km on a side [*Kikuchi and Kanamori*, 1994; *Silver et al.*, 1995]. Hence although this interpretation is non-unique [*Chen*, 1995], large fault zones appear to cut across the predicted narrow wedge of material below 600-800°C.

The idea of a temperature-controlled process is hard to abandon, however, because deep earthquakes occur only in those slabs which are relatively colder. Thus the fault areas may indicate slab temperature structures more complicated than the simple models of essentially parallel isotherms in undeformed slabs. The high seismic energy release below about 600 km [e.g., *Richter*, 1979], earthquake mechanisms [*Lundgren and Giardini*, 1994], images of slabs from seismic tomography [e.g., *Van der Hilst et al.*, 1991], and convection models [e.g., *Kincaid and Olson*, 1987; *Tao and O'Connell*, 1993] suggest that slabs deform due to interaction with the 670 km discontinuity, presumably because a major change in physical properties occurs at the base of the transition zone.

Kirby et al. [1995] hence suggest that the slab in the region of the Bolivian earthquake has a complex thermal structure because of variations in the age of the subducting plate over time and thickening due to slab deformation, causing a widened cold "pod". Large deep earthquakes could occur in this region, either due to metastability or another temperature-controlled process. The real geometry is presumably more complicated and varies within and among slabs. For the Tonga earthquake, characterized by many more aftershocks than usual for deep earthquakes, a different mechanism seems required, such as aftershocks by ductile faulting in the relatively cold spinel-rich region outside the wedge, perhaps triggered by a large transformational-faulting main shock in the wedge [*Kirby et al.*, 1996a].

Complex and variable deep slab thermal structure is plausible for several reasons. Although simple thermal models vary only slowly along strike for a given slab, deep seismicity is quite variable. Deep seismicity has distinct clusters and gaps where later large earthquakes can occur (as for the Bolivian earthquake) [*Kirby et al.*, 1995]. Tomographic images of deep slabs vary along strike and show more complexity [*Van der Hilst et al.*, 1991; *Fukao et al.*, 1992; *Engdahl et al.*, 1995; *Van der Hilst*, 1995] than simple thermal models predict [*Spakman et al.*, 1989]. As noted earlier, some of this complexity may result from deformation at the 670 km discontinuity. Moreover, in addition to mechanical perturbations to the slab, some of this variability may reflect metastability, because latent heat release would perturb thermal structure [*Daessler and Yuen*, 1993; *Kirby et al.*, 1996a]. These variations in both temperature and metastability would cause complex density variations, and thus affect slab stresses and driving forces [*Kirby et al.*, 1996a]. If wedges were large and continuous enough, their buoyancy might deflect the slab toward the horizontal [*Silver et al.*, 1995], as observed in some cases.

PROSPECTS

Ideas about the thermal structure of the oceanic lithosphere before it subducts, and the resulting thermal structure of slabs, seem poised for refinement. Both have been based on fairly simple models, as reviewed here. The models are surprisingly successful at describing the basic observed phenomena with a small number of parameters. This general success is striking, given that the models are two dimensional, include only the simplest thermal effects, and use temperature-, depth-, and pressure-independent physical properties.

The challenge is less how to pose more complex models, than to test and constrain them. Many features of the data illustrate the need for improved models. In particular, variations in ocean depth and heat flow about their mean values as functions of age, and the misfit to geoid data at fracture zones, illustrate the need for models in which depth and heat flow depend on more than age-dependent thermal structure. For this application, a best-fitting age-dependent thermal model can be used as a reference model to identify regions differing from the average at that age, and thus estimate the magnitude of the perturbing process [*Stein and Stein*, 1993]. Much needs to be done to characterize average lithosphere, and to investigate variations about the average. Our sense is that the primary deviations from halfspace cooling are thermal, and hence described on average by a plate model, whereas secondary regional deviations reflect temperature and pressure variations, perhaps due to both asthenospheric flow and local temperature variations. An important issue is the cause of variations in ridge crest depth [e.g., *Calcagno and Cazenave*, 1994] which provide differing initial conditions. We expect that models will continue to be posed and tested, hopefully by explicit numerical comparison to data [*Stein and Stein*, 1994b].

Similarly, the variation along subduction zones in phenomena including velocity structure and seismicity distribution illustrate the need for better thermal models. Because these variations can be attributed to effects not included in current models, such as three-dimensional slab geometry, slab mineralogy, and the variation of properties like thermal diffusivity with temperature and pressure, improved models are being suggested and it will be interesting to see which prove most successful.

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