

C. Project Description

1. Results from previous NSF funded research

1.1 Ritzwoller

Within the past five years Ritzwoller received three grants from NSF-OPP to perform surface wave and related modeling studies of polar regions; one from NSF-EAR to initiate a collaboration with Mexican scientists that involves data transfer from the Mexican National Seismic Network; one from NSF-OPP to host a workshop on the future of seismic infrastructure on Antarctica; one from NSF-EAR/CMG to host a summer school on mathematical geophysics and uncertainty in geophysical inversion, and most recently one from NSF-EAR to study the crust and upper mantle beneath Tibet.

NSF-OPP-9818498 3/15/99-2/28/02 \$199,984 *Surface Wave Tomography of the Arctic*
NSF-OPP-0136103 5/1/02-4/30/04 \$60,000 *Active Tectonics at the Aleutian-Kamchatka Corner: A Lithospheric Perspective*
NSF-OPP-0125848 7/1/02-6/30/05 \$242,000 *Refinements and Interpretation of Images of the Antarctic Crust and Upper Mantle*
NSF-OPP-0229185 6/1/02-5/31/03 \$52,000 *2003 Workshop: Structure and Evolution of the Antarctic Plate*
NSF-EAR-0207753 8/1/02-7/31/03 \$60,000 *Modeling the Middle American Lithosphere*
NSF-EAR/CMG-0327574 9/1/03-9/1/04 \$180,000 *CMG Training: Summer School on Mathematical Seismology and Uncertainty in Earth Models* (R. Snieder PI)
NSF-EAR-0337622 1/1/04-12/31/06 \$239,000 *Structure of the Tibetan Crust and Upper Mantle and its Geodynamical Implications*

These grants have been used to develop surface wave data sets and inversion methodology on a global scale. They have partially supported several post-docs (e.g., Barmin, Villasenor, Shapiro), two Ph.D. theses (Resovsky, 1997; Vdovin, 1999), one M.S. thesis (James, 1998), three senior honors theses (Landuyt, 2001; Leahy, 2002; D. Smith, 2004), and one current graduate student (S. Smith). Publications resulting from these grants break into the following categories:

Surface wave tomography: Barmin et al. (2001); Ritzwoller et al. (2002); Levshin et al. (2003).
Inversion methodology: James and Ritzwoller (1999); Shapiro and Ritzwoller (2002, 2003a); Shapiro et al. (2003c).
Regional studies: Kamchatka (Levin et al., 2002); Arctic (Levshin et al., 2001); Australian - Antarctic Discordance (Ritzwoller et al., 2003b); South America (Rial and Ritzwoller, 1997; Vdovin et al., 1999); Antarctica (Ritzwoller et al., 2001; Shapiro and Ritzwoller, 2003b); Eurasia (Ritzwoller and Levshin, 1998; Ritzwoller et al., 1998; Villasenor et al., 2001; Ritzwoller et al., 2003a); Pacific (Ritzwoller et al., 2003c).
Surface wave theory: Levshin et al. (1999).

There are four results of the research supported by these grants that are most important for the proposed research. First, there is the data set. As shown in Figure 3 below, the simultaneous inversion of broad-band group velocity measurements with intermediate and long period phase velocity measurements is much more powerful than the use of either data set alone. In oceanic regions, the two data sets help to improve the vertical resolution of the lithosphere from the asthenosphere and reduce ambiguities in estimating radial anisotropy in the uppermost mantle. Second, there is the development of diffraction tomography that models path-length dependent surface wave sensitivity, wave-form healing, and associated diffraction effects which are particularly important for the Pacific because of the long wave paths that cross it. Third, there is the Monte Carlo inversion from which we estimate model uncertainties, which allows us to identify and focus

interpretation on the most robust features of the model. Finally, there is the reformulation of the inverse problem in terms of a thermal model of the mantle that shows promise in improving mantle temperature estimates across the Pacific. Relevant aspects of the data set that we have developed so far, the diffraction tomographic method, the Monte-Carlo inversion method, and the application of thermodynamic constraints on seismic inversions are described in section 3 below.

1.2 Zhong

Zhong has received the following two grants from NSF-EAR:

NSF-EAR-0087567 1/1/2001 - 12/31/2003 \$235,000 *Modeling post-glacial rebound for Earth with 3-D viscoelastic structures*

NSF-EAR-0134939 2/1/2002 - 1/31/2007 \$310,000 *Career: Studies of mantle convection at multiple scales and integration of geophysical fluid dynamics in Geoscience*

The first grant has been used to develop a 3-D spherical shell finite element code for viscoelastic stress relaxation for an Earth with 3-D viscosity structure. We verified the validity of the code by applying it to post-glacial rebounding (PGR) with a realistic ice loading model. We have been investigating the effects of laterally varying viscosity and lithospheric structures on the PGR. So far, one publication (Zhong, Paulson and Wahr, GJI, 2003) and five AGU abstracts have resulted from this project. Two more manuscripts are in preparation (Paulson et al., 2004). The second grant has been used to investigate the effects of plume- plate interaction on the uplift of Hawaiian islands. We have also used the observations at Hawaii to constrain the plume dynamics. We formulated viscoelastic loading problems with a composite rheology (i.e., dislocation creep and diffusion creep) and applied the model to seamount loading to constrain the mantle rheology. We have also developed the capability to model sublithospheric small-scale convection (see Section 3.5). So far, three papers (Zhong and Watts, 2002, van Hunen and Zhong, 2003, Podolefsky et al., 2003) and four AGU abstracts have resulted from this project. These grants support two graduate students (Paulson and Podolefsky).

2. Introduction

Seismology has been conspicuously absent in most investigations of the thermal and dynamical state of the oceanic lithosphere. There have been numerous studies of the seismic wave speeds beneath oceans (e.g., Forsyth, 1975; Yoshii, 1975; Nishimura and Forsyth, 1989; Zhang and Tanimoto, 1992; Zhang et al., 1994; Wen and Anderson, 1997a; Ekström and Dziewonski, 1998; Katzman et al., 1998; Forsyth et al., 1998; Zhang and Lay, 1999; Romanowicz and Gung, 2002; Montagner, 2002), but studies of the temperature structure of the oceanic lithosphere have largely been the province of *geochemical* (e.g., Dick et al., 1984; Klein and Langmuir, 1987; Klein et al., 1988; Johnson et al., 1990; Langmuir et al., 1992; Zhang et al., 1994; Pyle et al., 1995; Niu and Hekinian, 1997; Lecroart et al., 1997; Vlastelic et al., 1999; Keller et al., 2000), *topographic* (e.g., McKenzie, 1967; Parsons and Sclater, 1977; Cazenave, et al., 1986; Davies, 1988; Renkin and Sclater, 1988; Marty and Cazenave, 1989; Colin and Fleitout, 1990; Cazenave and Lago, 1991; Stein and Stein, 1992, 1994 and 1996; Zhang et al., 1994; Kido and Seno, 1994; Carlson and Johnson, 1994; Stunff and Ricard, 1995; Smith and Sandwell, 1997; Wen and Anderson, 1997b; Wessel, 1997; McNutt, 1998), heat flow (e.g., McKenzie, 1967; Parsons and Sclater, 1977; Stein and Stein, 1992, 1994 and 1996; Nagihara et al., 1996), and *gravity* studies (e.g., McKenzie, 1967; Cazenave, et al., 1986; Haxby and Weissel, 1986; McAdoo and Sandwell, 1989; Maia and Diament, 1991; Cazenave et al., 1992; Cazenave et al., 1995; Wen and Anderson, 1997b; McNutt, 1998; Marquart et al., 1999; Marquart, 2001). Global seismology, in particular, has weighed in with little influential information about the cooling history of the oceanic lithosphere. This is largely because the lateral and, perhaps more importantly, the vertical resolution achievable by global models were insufficient to address questions related to thermal structure and evolution of the lithosphere.

The purpose of this proposal is to bring seismological models more prominently into the discussion about the thermal state and dynamic evolution of the oceanic lithosphere and asthenosphere. This proposal is motivated by the improving resolution of global seismic models of the uppermost mantle, by recent developments in the conversion between seismic speeds and temperatures that have placed the inference of temperatures from seismic speeds on firmer ground, and emerging community initiatives such as Ocean Mantle Dynamics (OMD), Margins, and Ridge 2000 that would benefit from the synoptic view of the oceanic upper mantle that the proposed research would provide.

We will concentrate our efforts principally on the Pacific lithosphere, because of its size and speed of plate motion; the fastest plate yields the best lithospheric age resolution.

There are three major questions that motivate this proposal:

Question 1. What is the temperature structure of the Pacific lithosphere and asthenosphere?

Question 2. What are the dynamical causes of some of the more prominent temperature variations?

In particular, are the dynamics controlled largely by near-surface physical processes (e.g., thermal boundary instabilities), by processes whose genesis lies deeper within the earth (hot-spot plumes, large-scale upwellings), or by interactions between deep-seated and shallow processes (e.g., plumes exciting thermal boundary instabilities)?

Question 3. What complementary information about the structure and dynamics of the lithosphere and asthenosphere is available in radial and azimuthal anisotropy?

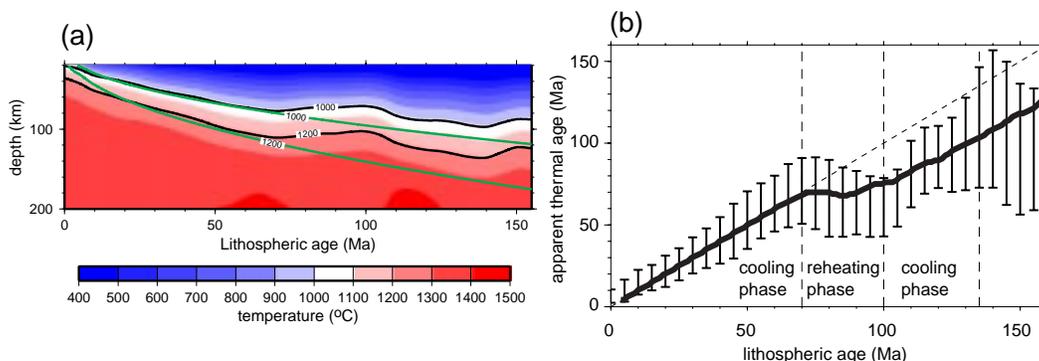


Figure 1: **Preliminary age-averaged temperature model of the Pacific upper mantle.** (a) The temperature model is compared with the prediction from isotherms from a diffusively cooling half-space (green lines). Isotherms in the Pacific are, on average, flat between about 70 and 100 Ma at depths of about 70 and 150 km. (b) Apparent thermal age, τ , is used to summarize lithospheric structure. Here it is averaged in 5 Ma age bins across the Pacific and “error” bars represent the standard deviation within each age range. Two phases of average lithospheric cooling are identified, 0-70 Ma and 100 - 135 Ma, bracketing a phase of in which the Pacific lithosphere undergoes reheating, on average. The thick black line is apparent thermal age, similarly averaged in age bins, computed from a 3-D convection model of thermal boundary instabilities (TBI), with an effective rheological activation energy of 120 kJ/mol. (Further discussion is included in section 3.5.1 below.)

These questions require a seismic model that samples the full range of ages of the Pacific lithosphere and, therefore, must span the entire Pacific. To achieve the vertical resolution needed to resolve high speed lithosphere from the low seismic speeds that underlie it in the asthenosphere, one needs surface wave dispersion measurements that extend to as short of periods as possible, preferably at least to 20 sec period. As described below, we have devoted considerable efforts already to developing a broad-band data set that will allow these questions to be addressed, as well as modeling tools to interpret these data in terms of the shear velocity and temperature structure of the upper mantle.

The questions above are motivated, in part, by a preliminary model of the thermal structure of the Pacific. One of the principal features of this model is illustrated in Figure 1, which shows

that the cooling of the Pacific lithosphere is not continuous, on average, but arrests between about 70 - 100 Ma. Periods of continuous cooling from 0 - 70 Ma and from 100 - 135 Ma, on average, bracket the period of arrested cooling. The effect of this lithospheric reheating in the Central Pacific maximizes at depths from about 70 to 150 km. Further discussion of these results is presented by Ritzwoller et al. (2003) (<http://ciei.colorado.edu/pubs/2003/7.pdf>) and in section 3 below.

This observation has motivated us to propose to investigate the thermal evolution of the Pacific upper mantle, compare with the thermal state of the lithosphere and asthenosphere beneath other oceans, investigate the cause of the arrested cooling that underlies the central Pacific, and explore the genesis of other thermal anomalies that are not so obviously dependent on lithospheric age. Because global seismic methods may image only the large-scale thermal consequences of dynamical processes that are too small to resolve (e.g., plume heating, small-scale convection), geodynamic modeling is crucial to explain the seismic results. Our recent dynamical models of thermal boundary layer instabilities (TBI) and plume heating reveal an intriguing interplay between structures and processes in the lithosphere and uppermost mantle (e.g., Zhong and Watts, 2003; van Hunen and Zhong (2003); van Hunen et al. (2003); Huang et al. (2003)).

The proposed work will take place in four principal research areas.

Purpose 1. We will continue to improve our seismic model of the Pacific lithosphere and asthenosphere. In particular, we will develop a larger data set of fundamental mode group velocities and phase velocities, introduce new measurements of fundamental mode and overtone phase velocities from our collaborators, advance the application of thermodynamic constraints on the model, and continue to improve the ability to estimate azimuthal anisotropy by extending the method of “diffraction tomography” to encompass azimuthal anisotropy.

Purpose 2. We will develop similar seismic and thermal models of the the Atlantic and Indian Oceans, for comparison.

Purpose 3. Assuming that the reheating event in the Central Pacific bears up under further scrutiny – as we have every reason to expect – we will investigate the cause of this reheating through geodynamical simulations of thermal boundary instabilities and their interaction with thermal plumes.

Purpose 4. We will also investigate the cause or causes of other observed thermal anomalies that may be best explained by geographical location or history (e.g., proximity to plumes).

3. Past work

3.1 Group velocity and phase velocity data sets

NSF has supported us in the past to obtain surface wave group velocity dispersion measurements in polar regions. We have, however, also devoted substantial efforts in the last three years to globalizing our data set and, in doing so, have already developed a large data set of fundamental mode group velocity measurements across the Pacific, Indian, and Atlantic Oceans. At present, we have obtained dispersion curves for more than 200,000 paths world-wide. The proposed work will rest, in part, on this existing data set. To date, phase velocity measurements have been graciously donated by Harvard and Utrecht Universities. Our efforts have been devoted to complementing research by other groups and, therefore, we have worked predominantly on improving group velocity coverage globally.

3.2 Diffraction tomography: isotropic and anisotropic

Our recent dispersion maps result from “diffraction tomography”, described by Ritzwoller et al. (2002). The method is based on a physical model of the surface wave Fresnel zone rather than on ray-theory with ad-hoc regularization (Gaussian tomography) such as the method documented by Barmin *et al.* (2001). Diffraction tomography accounts for path-length dependent sensitivity, waveform healing and associated diffraction effects, and provides a more accurate assessment of spatially

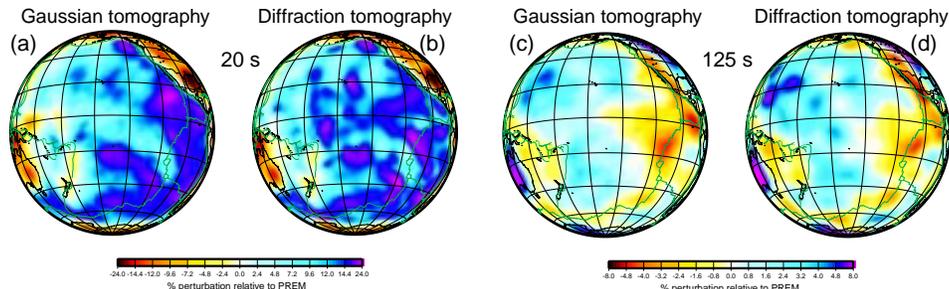


Figure 2: **Importance of diffraction tomography for the Pacific.** Demonstration of the importance of diffraction tomography for the Pacific. The example is for Rayleigh wave group velocity tomography. (a) and (c) Results from Gaussian tomography (ray theory with ad-hoc Gaussian smoothing). (b) and (d) Results from diffraction tomography. (a) and (b) 20 sec maps. (c) and (d) 125 sec maps. Diffraction tomography becomes increasingly important as periods grow because Fresnel zones widen with period. Even at 20 s period, however, diffraction tomography and Gaussian tomography yield substantially different results for the Pacific because wave paths are, on average, very long across the Pacific.

variable resolution than traditional tomographic methods. It is inspired by the work of Woodhouse and Girnius (1982), Wielandt (1987), Snieder and Romanowicz (1988), Yomogida (1992), Friederich (1999), Dahlen *et al.* (2000), Nolet and Dahlen (2000), Zhao *et al.* (2000), Spetzler *et al.* (2001), Yoshizawa and Kennett (2002) and others. (A more complete bibliography is given by Ritzwoller *et al.*, 2002). Figure 2 demonstrates the particular importance of using diffraction tomography for the Pacific where path lengths tend to be very long and Fresnel zones are wide, even at periods as short as 20 sec. As discussed below, diffraction effects may be important in estimating azimuthal anisotropy, too.

3.3 Monte-Carlo inversion: seismic and temperature parameterizations

We have used the method of Shapiro and Ritzwoller (2002) to estimate a shear velocity model on a $2^\circ \times 2^\circ$ grid globally to a depth of about 300 km. Views of the model can be interactively generated at <http://ciei.colorado.edu/~nshapiro/MODEL>.

The goal of the inversion procedure is to estimate the range of models that fit the dispersion maps subject to the uncertainties in the maps together with a priori information. The procedure culminates in a resampling of model space using a Monte Carlo method to produce a radially anisotropic v_s model and uncertainties equal to the half-width of the corridor of acceptable models. Although Monte Carlo and related model-based sampling methods have a long history in seismology (e.g., Levshin *et al.*, 1966; Keilis-Borok and Yanovskaya, 1967; Press, 1968; Calcagnile *et al.*, 1982; Lomax and Snieder, 1994; Shapiro *et al.*, 1997), we are unaware of studies that have performed them on a global scale. Consequently, global scale models have typically lacked uncertainty information.

An example for a point in the Pacific is shown in Figure 3, which also illustrates the importance of simultaneously inverting group and phase velocities. An example of a vertical slice across the Pacific is shown in Figure 4b. The most robust features of the resulting model are those that appear in every member of the ensemble of acceptable models. We identify these features in Figure 4 by encircling them with black lines. We refer to these features as “persistent” and concentrate interpretation on these features.

The model shown at the point in Figure 3 and the slice in Figure 4b is constructed using the seismic parameterization in the mantle. Some of the considerable variability in the lithosphere, seen in Figure 4b for example, is questionable on physical grounds. To address this issue, we have reformulated the inversion by replacing the seismic parameterization with a temperature parameterization that is based on the thermal model shown in Figure 5 in which a conductive lithospheric layer overlies a convective mantle. The inverse problem then reduces to estimating two thermal parameters, the apparent thermal age (τ) of the lithosphere and the potential temperature

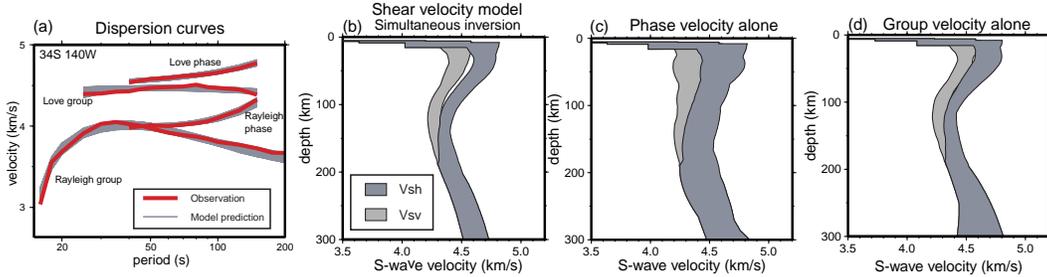


Figure 3: **Improved vertical resolution of joint inversion of group and phase velocity curves.** Example of the Monte-Carlo inversion using the seismic parameterization for a point in the Pacific (34°S, 140°W). (a) Observed dispersion curves are plotted as the thick red line and dispersion curves from the ensemble of acceptable models shown in (b) are plotted as grey lines. (b) Corridors defined by the ensemble of acceptable models that fit group and phase velocities simultaneously. (c) Same as (b), but only phase velocities are used in the inversion. (d) Same as (b), but only group velocities are used in the inversion. Simultaneous inversion of group and phase velocities greatly improves the vertical resolution of the model.

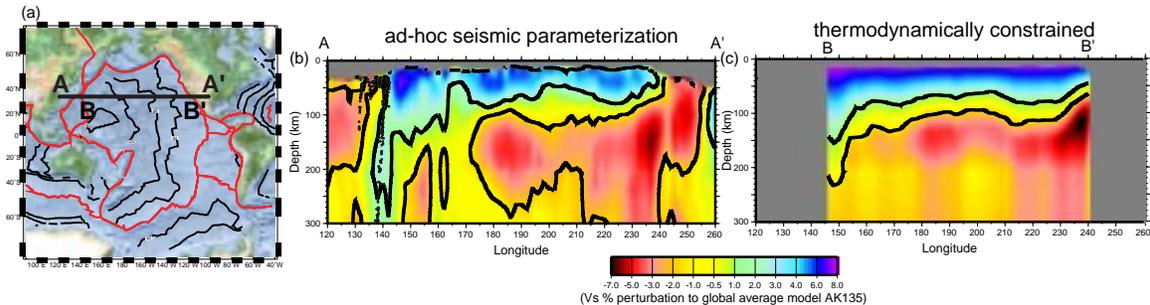


Figure 4: **Vertical slices across the Pacific using the seismic and temperature parameterizations.** (a) Reference map for (b) and (c). (b) - (c) Vertical slice across the Pacific comparing (b) the model using the ad-hoc seismic parameterization with (c) the thermodynamically constrained model which exists only in oceanic regions. Persistent features of the models are encircled with black lines.

of the asthenosphere (T_p). The imposition of physical constraints on the inversion reformulates the inversion as a hypothesis test. We hypothesize a physical model and determine if the data and other a priori constraints can be satisfied by the model. The example shown in Figure 4c illustrates that some of the variability displayed by the model based on the seismic parameterization is inconsistent with the physical constraints imposed by the thermal model. In addition, the data can, be fit across much of the profile shown in Figure 4 with the thermal parameterization. In regions where the data are not fit well by the thermal parameterization, the problem seems to be related to errors in the input crustal model, in particular crustal thickness.

We acknowledge that interconversion between shear velocity and temperature is hazardous, but believe that the uncertainties can be estimated and propagated in the inversion. Shapiro and Ritzwoller (2003a) discuss the interconversion in detail as well as the inherent uncertainties.

3.4 Preliminary results for the Pacific and other oceans

Dispersion maps across the Pacific, such as those in Figure 2, vary with lithospheric age. Although in a general sense dispersion velocities at intermediate and long periods increase with age, there is a detailed structure to this overall trend that contains information about the thermal state of the Pacific upper mantle. For example, intermediate period surface wave speeds increase approximately linearly with square-root of age until about 70 Ma, flatten between 70 Ma and 100 Ma, and then increase again until 125-135 Ma, on average, as shown in Figure 6. The exact patterns contained in the dispersion maps vary with period and wave type. These age trends are robust to data

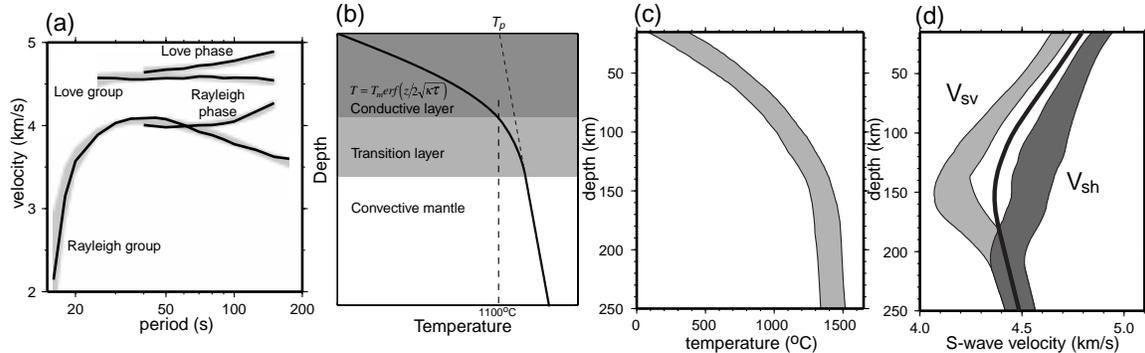


Figure 5: Example of the application of the temperature parameterization at a location in the Central Pacific (14°N, 200°E). (a) The four observed dispersion curves at the location are plotted with black lines. (b) The temperature parameterization is based on a thermal model in which an error function, which represents temperatures in the lithosphere, is underlain by an adiabatic gradient in the convective mantle (asthenosphere), joined smoothly by a transition region. The unknown in the conductive layer is the apparent thermal age, τ , and the unknown in the underlying asthenosphere is the potential temperature, T_p . (c) & (d) Inversion results. The ensemble of acceptable temperature models in the uppermost mantle is shown in (c). The ensemble of seismic models is displayed in (d), where the light grey-shaded envelope is V_{sv} and the dark grey-shaded envelope is V_{sh} . The thick black line is the median of the ensemble of isotropic shear velocities, V_s . This example demonstrates the unusual anisotropy in the Central Pacific⁶ in which the bifurcation between V_{sv} and V_{sh} grows with depth, maximizing here at about 140 km. Predictions from the ensemble of acceptable models to the four observed dispersion curves are shown as grey lines in (a).

subsetting, arbitrary choice of a broad range of damping, and the simultaneous estimation of azimuthal anisotropy.

The information in the dispersion maps is, ultimately, summarized in the V_s and temperature models. A particularly simple summary is given by the apparent thermal age, τ , which represents lithospheric temperatures. Preliminary estimates of apparent thermal age across the Pacific are shown in Figures 7 and 1 together with how τ deviates from lithospheric age. Prior to 70 Ma, the apparent thermal age trends with lithospheric age, on average, although there are regional variations. A systematic difference between the apparent thermal age of the lithosphere and lithospheric age sets-on at about 70 Ma across the entire Pacific. The central Pacific appears thermally younger than its true age. Beyond 100 Ma until about 135 Ma apparent thermal age again trends with lithospheric age, on average. This suggests that the cooling of the Pacific lithosphere progresses in two stages, from 0 - 70 Ma and from 100 Ma - 135 Ma, on average, but cooling is arrested from 70 - 100 Ma presumably by a process or processes that reheat the lithosphere in this age range. The reheating appears to maximize at depths from 70 to 150 km. As Figure 8 shows, the reheating of the lithosphere between 70 - 100 Ma appears to be a unique feature of the Pacific. This work is presented in greater detail in a submitted manuscript (<http://ciei.colorado.edu/pubs/2003/7.pdf>) where we argue that thermal boundary layer instabilities (TBI) may play a role in reheating the lithosphere in the Central Pacific.

Other seismic velocity and temperature variations are also apparent in our model, but are not correlated with lithospheric age. Examples of isochronous variability at 100 km depth are presented in Figure 9. The physical mechanism that causes these variations may be different than the processes that reheat the lithosphere in the Central Pacific. These processes may include variations in the conditions of formation along the ridge, direct thermo-mechanical erosion of the lithosphere by thermal plumes, heating due to large scale upwellings, variability in the interaction between TBI and thermal plumes, and non-stationarity of the physical processes that heat the Central Pacific lithosphere.

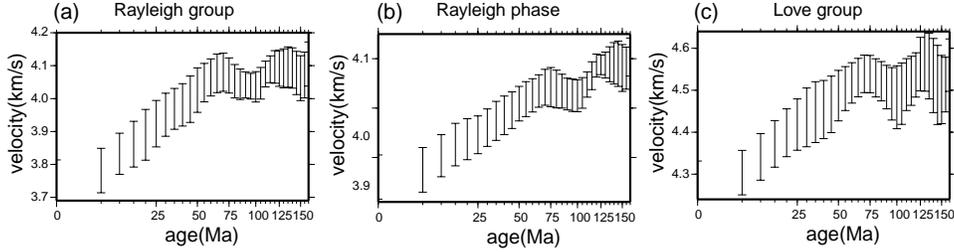


Figure 6: **Surface wave speed versus square-root of age.** Average speed for 50 sec Rayleigh wave group and phase speed and Love wave group speed plotted versus the square-root of age. The error bars display the standard deviation within each 5 Ma age range. These surface wave speeds increase linearly with square-root of age until about 70 Ma, flatten between 70 Ma and 100 Ma, and then increase again until 125-135 Ma, on average.

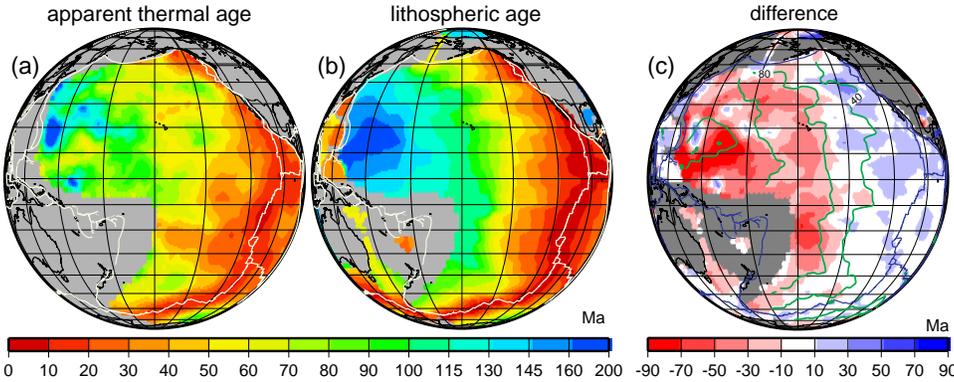


Figure 7: **Apparent thermal age.** (a) Preliminary estimate of apparent thermal age, τ , across the Pacific. (b) Lithospheric age across the Pacific (Mueller et al., 1997). (c) Local difference between the apparent thermal age and the lithospheric age. In Figure 1a, apparent thermal age, averaged in 5 Ma lithospheric age bins across the Pacific, is plotted versus lithospheric age. The pattern suggests two stages of cooling of the Pacific lithosphere bracketing a period between 70 - 100 Ma in which cooling is arrested.

3.5 Preliminary models of the interaction between the mantle and lithosphere

There are three dynamical interactions between the lithosphere and asthenosphere that are relevant to understanding aspects of the seismically inferred temperature structure beneath the Pacific: (1) spontaneous thermal boundary layer instabilities (TBI) also called sub-lithospheric small scale convection (SSC) caused by the cooling of oceanic lithosphere, (2) TBI precipitated by heating from below, and (3) thermo-mechanical erosion of oceanic lithosphere due to the dynamic interaction between lithosphere and ascending thermal plumes. Our preliminary studies of scenarios 1 and 3 demonstrate the intriguing dynamic interplay between lithospheric structure and mantle dynamics.

3.5.1 Thermal boundary layer instabilities (TBI)

TBI was originally proposed to explain the reduced subsidence of sea floor topography and the constancy of heat flux in relatively old sea floor [Richter, 1973; Richter and Parsons, 1975; Parsons and McKenzie, 1978], which deviate from the predictions of the half-space cooling (HSC) model [Parsons and Sclater, 1977; Lister et al., 1991]. TBI initiates near the base of oceanic lithosphere in a thermo-mechanical transition region where a strong lithosphere with a super-adiabatic temperature gradient gradually changes into a weak asthenosphere in which the temperature follows the mantle adiabat. In this transition region, the super-adiabatic temperature gradient and relatively low viscosity cause thermal boundary layer instabilities to develop. TBI has been proposed to cause

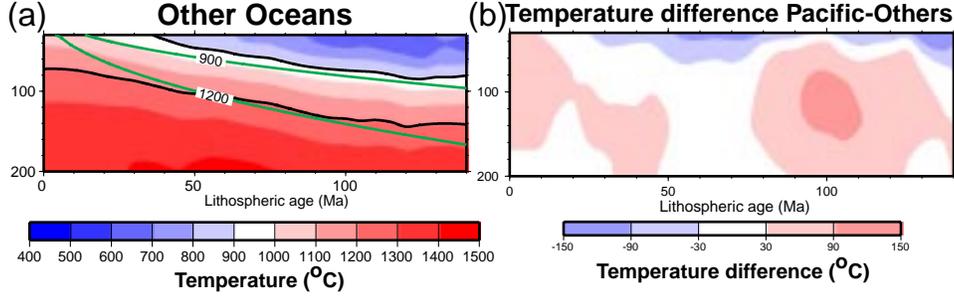


Figure 8: **Age-average upper mantle temperature in other oceans and comparison with the Pacific.** (a) Preliminary estimate of upper mantle temperatures averaged versus age in oceans other than the Pacific. Similar to Figure 1a, but the average is taken over the Atlantic and Indian Oceans whose isotherms more closely follow that of a diffusively cooling half-space (green lines). (b) Difference between Pacific mantle temperatures and those from the other oceans. A Pacific lithospheric temperature excess of more than 100°C is seen to start at about 75 Ma and maximize at about 100 Ma.

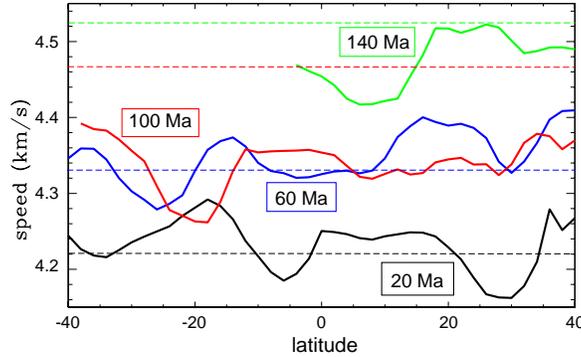


Figure 9: **Isochronous shear wave speeds at 100 km depth beneath the Pacific.** Shear velocity at 100 km depth plotted as a function of latitude (parallel to isochrons) in the Pacific. (Solid Lines) Observed velocities for four age ranges: (black) 20 ± 5 Ma, (blue) 60 ± 5 Ma, (red) 100 ± 5 Ma, (green) 140 ± 5 Ma. (Dashed Horizontal Lines) Predictions from a diffusive half-space model are color coded the same as the observations.

short-wavelength (~ 300 km) geoid undulations in the Pacific [Buck and Parmentier, 1986; Haxby and Weissel, 1986] and intermediate-scale structures in the Pacific upper mantle [Katzman et al., 1998].

Both numerical and laboratory studies suggest that the onset of TBI is controlled by mantle rheology; i.e., activation energy E and asthenospheric viscosity η_0 [Davaille and Jaupart, 1994; Buck and Parmentier, 1986; Yuen and Fleitout, 1985; Korenaga and Jordan, 2003; Huang et al., 2003]. These studies indicate that η_0 must be relatively small (e.g., 10^{19} Pa s) in order for TBI to produce constant heat flux [Davaille and Jaupart, 1994; Korenaga and Jordan, 2002]. These studies did not, however, consider the shear forces of plate motion and layered mantle viscosity structure [Hager and Richards, 1989] which may be important for the time evolution of TBI.

We have begun to study TBI by formulating both 2-D and 3-D mantle convection models with Newtonian rheology [Huang et al., 2003; van Hunen et al., 2003]. Our 2-D models have not included plate shear, but have helped us determine the scaling relations for the onset times of TBI on E and η_0 [Huang et al., 2003]. Our 3-D models are the first of this type of study of TBI with plate motion [van Hunen et al., 2003]. The effects of plate motion on onset times seem to be secondary. However, our 3D models have revealed important information about how the time evolution of TBI is related to thermal structure. Figure 10 presents an example the average temperature structure that results from the 3D simulation, which included plate motion, a relatively realistic temperature-dependent rheology, and a layered viscosity with a weak asthenosphere only in the top 410 km.

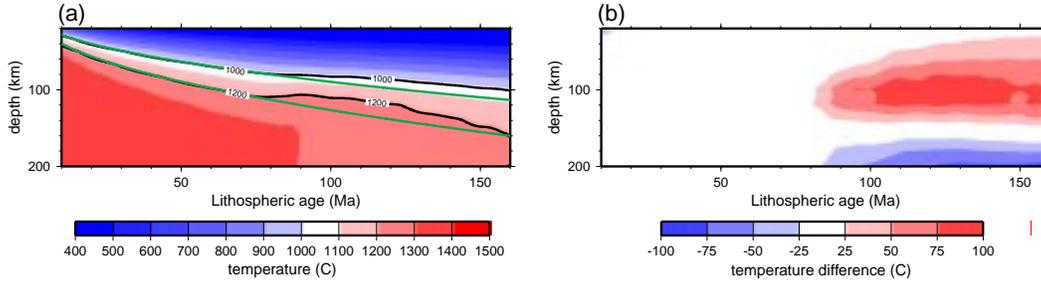


Figure 10: **Thermal structure versus lithospheric age for a 3D simulation of TBI.** (a) Temperatures from the 3D simulation of thermal boundary layer instabilities are averaged parallel to the ridge at each depth and plotted versus lithospheric age. The green lines are isotherms from the half-space cooling model as in Figure 1a. (b) The difference between the temperatures from the simulation of TBI with the HSC model. Reds imply that temperatures in the TBI simulation are warmer than in the HSC model.

A detailed description of the 3D simulation is beyond space limitations. TBI initiates in the middle of the box (corresponding to ~ 70 Ma) and continues to operate with varying vigor over the rest of the box. The effects of TBI on the thermal structure of lithosphere and asthenosphere is seen in both Figure 10 and in the apparent thermal age estimated for the 3D simulation and compared with the seismic estimate for the Pacific (Fig. 1b). The modeled apparent thermal age shows remarkable similarity to the seismic observations, in particular the significantly reduced thermal age between 70 Ma and 100 Ma and then increased thermal age after 100 Ma (Fig. 1). The large change in thermal age at 70 Ma is caused by the initial burst of TBI which tends to be very vigorous [van Hunen et al., 2003]. After 100 Ma, although TBI continues to operate, the reheated lithosphere actually cools with time.

This model of TBI shows the interesting dynamical interplay between rheology, thermal structure, and dynamics that may be essential to understanding our seismic observations of upper mantle structure. The principal remaining questions concern the rheological and thermal conditions that are necessary to reproduce the seismic observations and the effects of plume heating.

3.5.2 Thermo-mechanical erosion of lithosphere caused by ascending plumes

Spontaneous TBI does not involve forces other than the buoyancy of the lithosphere itself. However, other buoyancy forces, including thermal plumes, may also contribute to the evolution of lithospheric structure. This is particularly relevant in the Pacific where a super-plume (or a system of plumes) has been proposed to explain the uplift of residual sea floor topography [e.g., McNutt, 1996]. The dynamic origin of the super-plume is probably related to thermo-chemical convection in the lower mantle [e.g., Tackley, 1997; Davaille, 2000]. The plume structure appears to be consistent with seismic observations of radial anisotropy [Ekstrom and Dziewonski, 1997] and Q [Romanowicz and Gung, 2002]. It is possible, however, that TBI itself may not necessarily lead to reduced thermal subsidence of old sea floors [Davies, 1988].

A super-plume may provide heating that could influence shallow mantle thermal structure. The questions we address are how this heat is transmitted to the lithosphere and how it affects the thermal structure of the lithosphere and asthenosphere. The effects of a super-plume on the thermal structure of the upper mantle and lithosphere are not yet well understood partially because of our incomplete understanding of the origin of super-plumes. However, thermo-mechanical erosion of lithosphere due to plume-plate interaction or other forms of reheating has been studied extensively in the context of understanding the origin of the Hawaiian swell topography [e.g., Detrick and Crough, 1978; Olson, 1990; Ribe and Christensen, 1994, 1999; Moore et al., 1998; Zhong and Watts, 2002]. To first order, the effects of a super-plume can be informed by models of plume-plate interaction for Hawaii, even though the plume is much smaller than that of a super-plume. We have studied the Hawaiian plume to demonstrate the general features of 3-D plume-plate interaction

[Zhong and Watts, 2002; van Hunen and Zhong, 2003] similar to studies by Ribe and Christensen [1994] and Moore et al. [1998]. Upon reaching the bottom of the lithosphere, the ascending plume pushes aside (or erodes) the bottom portion of the lithosphere over a region significantly larger than the plume. In general, the erosion caused by the plume increases the temperature of the shallow mantle and lithosphere without decreasing the asthenospheric temperature, which is different from TBI and may provide a way to discriminate between these scenarios. Our results show that the amount of thermo-mechanical erosion of the lithosphere is dependent critically on the strength of the lithosphere (i.e., rheology) and also on the temperature anomalies of the plume.

4. Proposed work

4.1 Continue seismic data base and model development

We propose to continue to develop the surface wave data base across the Pacific and other oceans including developing our own data base of phase velocity measurements, to introduce new data contributed by collaborators (Trampert, Woodhouse, van Heijst) including information from mantle overtones, to continue the study of azimuthal anisotropy informed by modifications to the theory to include spatially extended sensitivity kernels, and to continue to advance the application of thermodynamic constraints on the seismic inversion.

4.1.1 Further data base development

The data distribution for Rayleigh wave group velocities remains far from uniform across the Pacific. The situation is worse for Love waves. For paths crossing the Pacific, Love wave group velocities are difficult to measure below about 25 sec period and above about 100 sec. This is not ideal, good distribution of both Rayleigh and Love waves is needed to resolve variations in the strength of radial anisotropy (RA) in the upper mantle. This is particularly important because errors in RA estimates can obscure thermal anomalies.

Although we have been working on the current data set for several years, much improvement is still possible. particularly given the introduction of ocean island stations over the past few years and when data from the Ocean Mantle Dynamics initiative become available. We propose to improve data coverage by continuing to obtain Rayleigh and Love wave group velocities as we have in the past, but we will also make phase velocity measurements on the cleaned waveforms that emerge from our standard data processing procedure. In our standard procedure, every Rayleigh and Love wave group velocity dispersion measurement involves human interaction to choose the frequency band of measurement, identify and circumvent spectral holes, and separate the fundamental mode from other signals and noise. This results in shorter period measurements than can be obtained automatically and a very high quality data set. Output from this procedure is a waveform cleaned from “noise”. We will use this data set as the basis for fundamental mode phase velocity measurements, to complement data provided by colleagues. At present, the phase velocities in our data set have been graciously donated by Harvard (Ekström et al., 1997) and Utrecht (Trampert and Woodhouse, 1995) Universities.

Jeannot Trampert from Utrecht has recently expanded his global phase velocity data set by about a factor of 10. We will collaborate with Dr. Trampert in combining this large phase velocity data set with our group and phase velocities world-wide.

As shown in Figure 3, the combined use of fundamental mode group and phase velocity measurements provides substantial improvement in the vertical resolution of the uppermost mantle. This improvement is needed to resolve the temperatures in the lithosphere from the underlying asthenosphere. Nevertheless, resolution below about 250 km remains poor. To improve resolution below 200 km, our collaboration with Jeannot Trampert has expanded to include John Woodhouse and Hendrik van Heijst so that we can combine the overtone measurements of van Heijst and Woodhouse (1999) with the fundamental mode group and phase velocity measurements. The overtone measurements are distributed very inhomogeneously, but are most numerous across the northern

Pacific. We will construct all of the dispersion maps (fundamental and overtone, Rayleigh and Love) at CU-Boulder and perform the inversion using an extension of our Monte-Carlo inversion method. The combination of data sets with different intrinsic resolutions presents problems that we can address by performing the inversion iteratively, first for the largest scale structures across the entire upper mantle and transition zone, followed by the introduction of smaller scale structures in the uppermost mantle alone using only the highest resolution data which are preferentially sensitive to shallow structures.

4.1.2 Thermal model of the Pacific lithosphere

The new data set described in section 4.1.1 will provide powerful new constraints on the thermal and RA structure of the oceanic upper mantle. At least in the northern Pacific, the overtone measurements will improve our ability to determine how deeply RA extends into the mantle. The resulting model, however, will still be prone to contamination from errors in a priori information, such as the starting crustal model. We propose to continue to develop and apply thermodynamic constraints, such as those presented in Figure 5, to help improve the uppermost mantle model and to test our understanding of the thermal state of the lithosphere and asthenosphere.

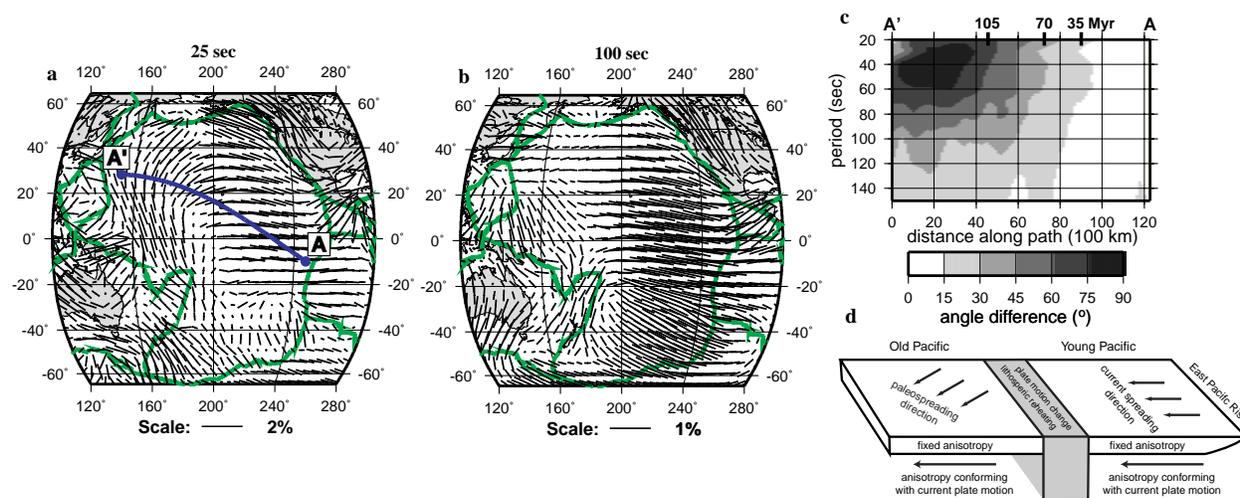


Figure 11: **Preliminary results on Pacific azimuthal anisotropy.** (a) - (b) 2ψ fast axes for the 25 sec and 100 sec Rayleigh wave group velocity. (c) Angular difference between fast axis directions and present day plate motions (HS3-NUVEL1A, Gripp and Gordon, 2002) along profile A – A' shown in (a). Best agreement is in the young Pacific at all periods, and largest differences are at the shorter periods in the old Pacific. (d) Cartoon interpretation of the azimuthal anisotropy observations in which asthenospheric anisotropy conforms with present day plate motions at all ages, but anisotropy is approximately fixed in the shallow lithosphere agreeing better with paleo-spreading directions than current plate motions in the old Pacific.

4.1.3 Azimuthal anisotropy (AA)

Maps of surface wave azimuthal anisotropy (AA) have appeared recently from several groups (e.g., Montagner and Guillot, 2000; Becker et al., 2003; Trampert and Woodhouse, 2003). These maps, however, are all for fairly long period phase velocities. The group velocity maps that we are constructing extend to much shorter periods which provides improved resolution of the lithosphere. Some preliminary results on AA for group velocities across the Pacific are summarized in Figure 11. As with phase velocities, AA for the long period Rayleigh wave group velocities, which are most sensitive to the asthenosphere, agree fairly well with current plate motion directions across most of the Pacific. At short periods, however, agreement with plate motion is confined to the

eastern Pacific. At ages beyond about 40 Ma, fast axis directions of waves sensitive predominantly to the lithosphere are rotated in a direction more similar to the paleo-spreading direction than the current plate motion direction. This apparent shift in fast axis directions in the Central Pacific may be due to a change in plate motion direction circa 40 Ma. A disruption of the mineralogical fabric in the lithosphere due to dynamical processes such as plume heating or TBI may also play a role, however.

Before addressing the cause of the disparity between fast-axis directions in the lithosphere and asthenosphere we propose first to improve tomographic capabilities by extending diffraction tomography to the estimation of AA across the Pacific. Preliminary results show that extended Fresnel zones affects estimates of AA in some places. Although AA is interesting in itself as a key to preferred-crystal orientation in the uppermost mantle, in terms of the goals of this proposal what may be most important is to ensure that AA does not bias the isotropic part of the dispersion maps. Simultaneous inversion of isotropic and azimuthally anisotropic velocities is desired. Final results await the conversion of the AA part of our tomographic code to incorporate diffraction effects.

4.2 Modeling lithosphere - upper mantle interactions

The proposed work on mantle dynamics is aimed at understanding the dynamic interplay between oceanic lithosphere and upper mantle. In particular, we are interested in how the two-stage cooling inferred for the Pacific lithosphere may happen and what controls it. By modeling relevant dynamic processes for mantle-lithosphere interaction over a large parameter space and comparing the model predictions with seismic inferences and other observations including sea floor topography, we will seek to constrain the dynamics and rheology of the mantle. We will consider mainly the two end-member processes that were discussed earlier: spontaneous, self-generated thermal boundary instabilities (TBI) and thermo-mechanical erosion due to thermal plumes.

We will employ 3-D Cartesian models similar to those used in Zhong and Watts [2002] and van Hunen et al. [2003] to study lithosphere-asthenosphere interactions. We will use both Newtonian and non-Newtonian rheologies to explore a sufficiently large parameter space.

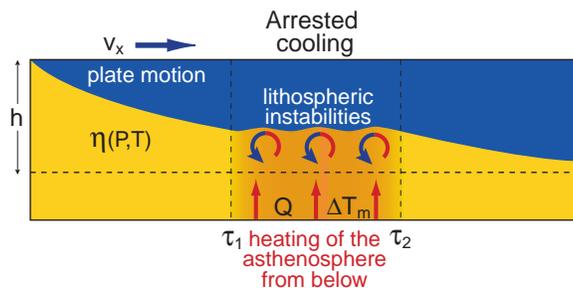


Figure 12: **Alternative processes for lithospheric heating.** Shown here are parameters describing a generalized physical model that will be explored to explain the punctuated cooling history of the central Pacific lithosphere. By varying the model parameters, diffusive heating, thermo-mechanical erosion by ascending plumes, and small-scale lithospheric instabilities that are either self-generated or precipitated by plume heating from below can be simulated. The parameters of the model include plate speed v_x , viscosity $\eta(P, T)$ which depends on activation energy E and reference viscosity η_0 , lower thermal boundary conditions including heat input Q and temperature anomaly ΔT_m between ages τ_1 and τ_2 , and plate thickness h for purely diffusive models.

In general, our models will include four independent parameters (Figure 12): (1) heating from a super-plume or a system of plumes which can be characterized as either a temperature anomaly ΔT_m or heat flow anomaly Q at the bottom boundary and their horizontal extent (τ_1, τ_2), (2) activation energy E , and (3) reference viscosity η_0 in the asthenosphere and the mantle below. Models with $\Delta T_m = Q = 0$ can be used to investigate spontaneous TBI, such as the 3-D simulation from which Figure 10 derives. However, for models with basal heating from plumes, TBI may still

take place depending on rheological parameters. In fact, it remains unclear how a super-plume may enhance or hinder the development of TBI. While plume heating may enhance TBI by reducing the asthenospheric viscosity, the large-scale pressure associated with the upwelling may also reduce the thickness of the potentially unstable part of the lithosphere, thus hindering the development of TBI. To study the interaction between a super-plume and TBI will be one of the major goals of our studies.

By varying these four model parameters ($\Delta T_m, Q, E, \eta_0$), we can encompass three different dynamical scenarios: (1) predominantly spontaneous, self-generated TBI, (2) predominantly plume heating without TBI, and (3) TBI in the presence of plume heating. We recognize that plume heating models may be more relevant to the Pacific lithosphere (e.g., Romanowicz and Gung, 2002).

4.2.1. Spontaneous thermal boundary layer instabilities (TBI)

For self-generated TBI models, we will set $\Delta T_m = Q = 0$ and systematically vary rheological parameters including E and η_0 in the asthenosphere and the mantle below (i.e., the thickness of asthenosphere) and the power-law exponents (i.e., non-Newtonian) to explore how TBI initiates and evolves in 3-D models as oceanic lithosphere ages. For each model, we will determine (1) the dependence of lithospheric and upper mantle thermal structure and thermal anomalies on lithospheric age, (2) surface heat flux, and (3) surface dynamic topography including that from lithospheric cooling. In synthesizing these models, we will determine the parameters that control both the increase in lithospheric temperature following the onset of TBI and the convective structure of TBI.

When possible (e.g., for onset time and temperature anomalies of TBI), we will develop scaling laws to delineate the parameter dependences and compare our results to previous 2-D and 3-D models with no plate motion [e.g., Davaille and Jaupart, 1994; Korenaga and Jordan, 2003; Huang et al., 2003]. One important aspect of our studies is to understand the two-stage cooling, as observed for the Pacific (Figure 1a). For the TBI models to work, we anticipate that activation energy E and asthenospheric thickness will play a critical role (e.g., Figure 10). The temperature anomalies produced by TBI are inversely proportional to E . When E is too large, the TBI may not produce sufficiently large temperature anomalies to affect the lithospheric structure. Non-Newtonian rheology would increase the temperature anomalies of TBI for a given E . We anticipate that the observed two-stage cooling may pose important constraints on the mantle rheology including activation energy E .

4.2.2. Effects of plume heating

Plume heating of the upper mantle from, for example, hotspots, is expected to be substantial in the Pacific. Figure 13 illustrates how much of the western Pacific has resided close to one or more hotspots during its paleo-trajectory. The effects of thermal plumes may be important both to explain the time-average thermal evolution of the Pacific and to explain isochronous thermal anomalies such as those seen in Figure 9.

For plume heating dynamical models, we will systematically vary the amount of plume heating Q as well as E and η_0 . For each model, in addition to the dependence of the thermal structure of the lithosphere and the upper mantle on lithospheric age, surface heat flux, and dynamic topography, we will also determine whether TBI will occur. In synthesizing these models, two questions are particularly interesting.

Question 1. How does the plume heating erode and reheat the lithosphere and what controls the efficiency of this process? The key to this question is to understanding the effects of rheological parameters on the thermo-mechanical erosion. The stronger the lithosphere, the more difficult it is for the reheating and erosion to occur. A special case is a static plate model in which the plate is infinitely rigid and reheating can only take place through the inefficient process of heat conduction. For relatively large E and η_0 , the TBI may not happen and the lithospheric reheating must rely

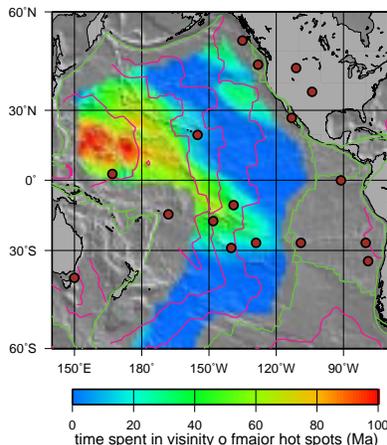


Figure 13: **Past proximity to hotspots.** Map showing the duration different parts of the Pacific plate have spent in the vicinity of major hotspots (within 1000 km radius) in units of Ma. Hotspots are shown with red circles (Richards et al., 1988). We assumed that hotspots remained fixed during last 180 Ma and used the "backtrack" algorithm and the stage poles for the Pacific plate from Wessel and Kroenke (1997) to compute geographic paleo-trajectories for each point within the Pacific plate. The green lines denote plate boundaries, the magenta lines are isochrons of lithospheric age in increments of 35 Ma.

only on the plume flow to mechanically erode the lithosphere. When E and η_0 are sufficiently smaller, TBI may happen to aid the reheating process. We will determine whether plume heating and its associated upwelling flow field are sufficient to reheat the lithosphere without TBI.

Question 2. Does plume heating enhance or hinder TBI? Because plume heating will very likely cause the reheating or the arrested cooling of lithosphere (its efficiency depends on the rheological parameters), essential to our studies will be the physical mechanism to restart the lithospheric cooling after the lithosphere passes the plumes. If plume heating enhances TBI, it is quite possible that the lithospheric reheating may continue indefinitely even after the lithosphere passes the plumes. In this scenario, we may have to resort to the mechanism outlined in section 4.2.1 (i.e., the cooling of the asthenosphere from TBI and the resulting increase in viscosity) to diminish TBI and restart the lithospheric cooling. This may happen only for certain rheological parameters. We will address this second question by carefully comparing models with the same rheological parameters but different amounts of plume heating, including models in section 4.2.1.

4.2.3. Distinguishing between the alternatives

We will develop diagnostic measures to distinguish between competing models. The spontaneous TBI model and the plume heating model may differ at least in the following two respects that may help to distinguish between them. (1) Dynamic topography. Plume heating should produce significant long-wavelength topography above the plume, while TBI itself may not result in significant variations in topography [Davies, 1988; Huang and Zhong, 2002]. (2) Vertical temperature variations in the upper mantle. Plume heating should increase the temperature in the lithosphere and upper mantle, while TBI may increase lithospheric temperature but reduce the upper mantle temperature (i.e., redistributing the heat vertically). They may also result in very different partial melting.

We will investigate the trade-offs between different parameters. It is likely that within a certain range of parameters, different combinations of parameters may produce similar results. It is important to identify the possible trade-offs. In particular, we will examine the trade-offs between the thickness of asthenosphere, E and η_0 on halting TBI and restarting lithospheric cooling.