STRUCTURE AND DYNAMICS OF THE CONTINENTAL TECTOSPHERE

by

STEVEN SAMUEL SHAPIRO

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Signature of Author

Department of Earth, Atmospheric, and Planetary Sciences
August 1995

Certified by
Bradford H. Hager
Professor of Geophysics
Thesis Supervisor

Certified by
Thomas H. Jordan
Professor of Geophysics
Thesis Co-Supervisor

Accepted by
Thomas H. Jordan
Department Head
To My Father

With All Of My Love
IN MEMORY OF

ROGER NATHAN METZ
(1938-1991)

Advisor, Supervisor, and Friend
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ABSTRACT

We examine, as a function of depth, the relationship between a tectonic regionalization and upper-mantle shear-wave heterogeneity represented by a recent seismic tomographic model. We perform Monte Carlo simulations that incorporate the spectral properties of both the regions and the seismic signal. Our results indicate that ridges can be readily distinguished from older oceans to a depth of about 200 km. The corresponding platform and shield signature differs significantly (>99% confidence) from that under oceans and orogenic zones to at least 400 km depth.

Results from analogous Monte Carlo simulations reveal that the earth’s gravity variations correlate with surface tectonics no better than they would were the geoid (or gravity field) randomly oriented with respect to the surface. We estimate for the upper mantle a platform and shield signal of $-8 \pm 5$ m and thus conclude that there is little contribution of platforms and shields to the gravity field, consistent with their keels having small density contrasts. We estimate an average value for $\partial \ln \rho / \partial \ln \nu$, within the 140-440 km depth range beneath platforms and shields of $0.035 \pm 0.025$, consistent, at the 1.5$\sigma$ level, with Jordan’s [1988] isopycnic hypothesis.

Through a suite of numerical finite element experiments, we evaluate the relative importance of (1) activation energy (used to define the temperature-dependence of viscosity), (2) compositional buoyancy, and (3) linear or nonlinear rheology in achieving the long-term stability of the continental tectosphere. Stability is assured with a realistic activation energy regardless of the chemical concentration. With lower values of activation energy, compositional buoyancy can significantly influence stability. Compositional buoyancy plays a dual role: It (1) counteracts the thermally-induced density increase and, relatedly, (2) reduces the stress within the boundary layer. With a stress-dependent rheology, this reduction in stress results in an increase in effective viscosity which, in turn, inhibits a greater region of the boundary layer from deforming. The joint application of longevity and gravity constraints allows us to reject all models containing no compositional buoyancy and to predict that the ratio of compositional to thermal buoyancy within the continental tectosphere is approximately unity.

Thesis supervisor: Bradford H. Hager, Professor of Geophysics
Thesis co-supervisor: Thomas H. Jordan; Professor of Geophysics
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CHAPTER 1

INTRODUCTION

The possibility of thermal convection within the earth "has been discussed extensively by geophysicists and geologists, but it appears that the only thing upon which they can be said to agree at present is the desirability of making a quantitative study of it" [Pekeris, 1935].

Sixty years after Pekeris [1935] wrote these words, many basic questions remain concerning the nature of convection within the earth. We will concentrate here on an issue which is directly related to large-scale mantle flow — the structure and dynamics associated with thick continental keels (the continental tectosphere) which translate coherently with the plate during plate motions. This topic represents one of many areas of geophysics which began with, and continues to have, a colorful and controversial image.

Many interleaved lines of inquiry bear on the structure and dynamics of the continental tectosphere. They involve investigation into the chemical and physical properties of the constituents of the upper mantle and their relation to the origin and evolution of the continental tectosphere. In this introduction, after briefly describing some pre-plate tectonic work, we discuss the three most prominent models of the continental tectosphere that were advanced in the late 1970's and early 1980's. We next discuss, in the context of these models, several controversies which began decades ago and exist to this day. We conclude this introduction with an outline of the issues that we address in the remaining chapters of this thesis. Our overall goal is to tightly constrain acceptable models of the continental tectosphere via a multi-pronged quantitative investigation which takes appropriate account of the sensitivi-
ties to the various physical parameters and the uncertainties in the inferences from statistical analyses.

The depth extent and the nature of differences between oceanic and continental tectosphere have been subjects of heated debate during the last three decades. Although Gutenberg [1924] discovered that within the upper 100 km of the mantle, continents were seismically "faster" than ocean basins, it was the combination of seismic, gravity, and heat-flow data that led MacDonald [1963] to first propose that subcontinental and suboceanic materials remain separate and chemically distinct to depths greater than 500 km. MacDonald [1963] used his conclusion to argue against the plausibility of continental drift, asserting that such deep continental "roots" would prevent the translation of continents along the earth's surface.

With the plate tectonics revolution of the latter part of the 1960's came better data and a much improved understanding of geodynamics. By the mid-1970's, the concept that the subcontinental mantle is, on average, seismically faster than the corresponding mantle beneath oceans and orogenic zones, was well documented [Brune and Dorman, 1963; Toksöz and Anderson, 1966; Knopoff, 1972]. However, since most of the relevant seismic studies were based on fundamental-mode surface waves, which have poor resolving power below about 200 km depth, testing the hypothesis of deeper continental structure required the examination of body wave data. But analyses of $ScS_n$ - $ScS_m$ differential travel times provided evidence both for [Sipkin and Jordan, 1975, 1976, 1980] and against [Okal and Anderson, 1975; Anderson, 1979] such deeper continental structure.

Sclater et al. [1980, 1981] proposed that the thickness of the continental thermal boundary layer (TBL) could be explained through an extension to continents of the oceanic plate cooling model [e.g., Parsons and McKenzie, 1978]. Because continents are older than oceans, the continental TBLs would have therefore reached their (asymptotic) thickness of about 150 km. Like the TBL thickness
associated with old oceans, this asymptotic thickness would be maintained by small scale convective instabilities. However, the approximate equality in surface heat flow in old ocean basins and stable continents, combined with the larger concentration of heat-producing agents (U, Th, K) in the continental crust, suggests that the heat flux through the base of the continents is significantly less than that associated with oceans. This heat flow discrepancy indicates that the oceanic and continental thermal profiles must differ significantly, hence old oceans and continents cannot conform to Sclater et al.'s [1980, 1981] model. In response to this conclusion, Sclater et al. [1980, 1981] countered by arguing that hydrothermal circulation in the oceans would more efficiently remove heat from the oceanic crust and that this process could explain the discrepancy. The existence of subcontinental seismic roots and their inferred thermal anomalies, however, would not be compatible with this "unified TBL" model.

Consistent with the hypothesis of continental deep structure, Pollack and Chapman [1977], proposed that the thickness of the continental TBL is not regulated by small-scale basal convective instabilities, but instead increases indefinitely with the square-root of age. Because the continents are older than the oceans, the continental TBL would be thicker in this model, and hence there would be a strong correlation between surface tectonics and TBL thickness. But, such continuous thickening would lead to continental subsidence that is inconsistent with the inferred (near) constancy of continental freeboard [e.g., Wise, 1974]. Further, local Rayleigh number calculations [e.g., Jordan, 1975] suggest that such a thick, uncompensated, TBL would be convectively unstable. Finally, if the continental TBL were stable, after 1 Gy, its thickness would exceed 300 km. As a consequence of such large thermal, and hence density, differences with oceans, continents should be distinguishable from oceans in the long-wavelength geoid. Whether there is such a (significant) ocean-continent signal in the geoid has been debated for decades
Jordan [1975, 1978, 1981a, 1988] created a comprehensive model of the continental tectosphere designed to satisfy the deep structure hypothesis, the continental freeboard evidence, and convective stability requirements, while satisfying the observation that there is no large correlation between continents and the long-wavelength geopotential [e.g., Kaula, 1967]. Jordan [1978] proposed that the continental TBL is ~ 400 km thick and is stabilized by a chemical boundary layer (CBL), consisting primarily of peridotite depleted in major-element basaltic constituents (i.e., \( \text{Al}_2\text{O}_3 \), \( \text{FeO} \), \( \text{CaO} \)) and enriched in large ion lithophile elements relative to the average composition of the upper mantle. The removal of basaltic constituents from a garnet lherzolite (which approximates the mantle mineralogy within the 70-400 km depth interval) would result in a lighter residual rock. By comparing average continental garnet lherzolites (ACGL) obtained from the whole rock analyses of xenoliths from kimberlite pipes with published models of oceanic mantle composition [Ringwood, 1966; Melson et al., 1966], Jordan [1978] showed that the ACGL is depleted in garnet and clinopyroxene and is characterized by a lower \( \text{Fe}/(\text{Fe}+\text{Mg}) \) ratio. The corresponding differences in density of 1.3% ± 0.2% would, under mantle conditions, offset the density imbalance resulting from about a 400°C ocean-continent temperature difference [Jordan, 1978]. At 200 km depth, such a temperature difference is expected [e.g., Pollack and Chapman, 1977]. Through the consideration of normative densities and temperature-sensitive \( \text{Ca}/(\text{Ca}+\text{Mg}) \) ratios, Jordan [1978] inferred that basalt depletion decreases with depth as the oceanic and continental thermal profiles converge, so as to maintain equality between the density of the oceanic and continental material as a function of depth. This isopycnic hypothesis [Jordan, 1988] states that, relative to the oceans, continents have positive buoyancy due to composition that exactly cancels negative
buoyancy due to lower temperatures at every depth between the base of the mechanical boundary layer and the base of the TBL. Continents which satisfy the isopycnic hypothesis would neither generate geoid height anomalies nor modify continental freeboard. In a modification to the isopycnic hypothesis, Forte et al. [1995] proposed that in the mantle above a depth of about 250 km, continents are denser than oceans but below about 250 km, continents are more buoyant than their oceanic counterpart. Explicit in Forte et al.'s [1995] model is a statistically significant (negative) correlation between continents and the long-wavelength geoid.

Jordan [1975] asserted that compositional buoyancy would also stabilize the TBL from convective disruption by reducing the effective Rayleigh number to a value below critical. Because thermal decay and chemical accretion operate on different spatial and temporal scales, Jordan [1988] rejected the notion [e.g., Oxburgh and Parmentier, 1978] that the continental TBL formed primarily by the continuous addition to it of depleted mantle from diapirs originating from subducted oceanic plates. Jordan [1978] instead proposed that the formation of the continental tectosphere in the Archaean was a natural consequence of the Wilson cycle, and that it formed via advective thickening resulting from episodes of compressive orogenesis.

To further evaluate these models of the continental tectosphere, it would be useful to estimate globally the depth extent of the ocean-continent seismic differences. Differential travel-time measurements of $S$, $SS$, and $SSS$ waves [e.g., Grand and Heimberger, 1984a, 1984b, 1985], global tomographic inversions [e.g., Woodhouse and Dziewonski, 1984; Masters et al., 1992; Grand, 1994; Su et al., 1994], and a waveform inversion scheme utilizing surface-wave data [Lerner-Lam and Jordan, 1987] have furnished further support to the continental deep-structure hypothesis. In addition, Lerner-Lam and Jordan [1987] showed that there is a correlation between surface tectonics defined by the six regions of GTR1
[Jordan, 1981b] and the one-way vertical shear-wave travel-times within the 40-400 km depth interval, predicted by the degree-eight tomographic model M84C [Woodhouse and Dziewonski, 1984]. In particular, they found that there is a monotonic progression from seismically "slow" young oceans to "fast" old continents. Other global studies, however, conclude that the seismic properties of cratons do not differ significantly from the global average below about 250 km depth [e.g., Polet and Anderson, 1995].

Additional evidence concerning the nature of the continental tectosphere comes from geochronology. Measured ages of South African diamond inclusions brought to the surface from about 200 km depth [e.g., Richardson et al., 1984; Richardson, 1986; Richardson et al., 1993] provide evidence that some continental tectosphere material is as old as 3.5 billion years. Additionally, recent analyses of rhenium-osmium and other isotope systematics indicate that some Siberian and South African peridotites have been isolated from the convecting mantle for more than three billion years [Walker et al., 1989; Pearson et al., 1995]. The age data further suggest that the whole lithosphere down to a depth of at least 150 km formed contemporaneously with the stabilization of the continental crust [Pearson et al., 1995]. While these age estimates indicate that continents are old, the data are as yet unable to illuminate either the age or the properties of the continental mantle below a depth of about 200 km. Anderson [1989], for example, suggests that seismically fast continental mantle within the 200-400 km depth interval represents the remnants of cold oceanic lithosphere, not old tectosphere.

To have been created in the Archaean and to exist today, the continental tectosphere would have had to withstand for several billion years the disruption caused by the basal tractions associated with a convecting mantle and the double-diffusive instabilities intrinsic to chemically gradated structures such as those proposed by Jordan [1975]. From an analytical analysis of the stability of a constant
viscosity ($\eta = 10^{21}$ Pa s) continental tectosphere described by linear lateral gradients in composition and temperature, Stevenson [1979] found modes of instability with characteristic growth times as short as about 200 My — about an order of magnitude less than that required by the above age constraints. However, if one considers the temperature dependence of viscosity and the fact that continents are cold, a constant viscosity of $\eta = 10^{21}$ Pa s is an unrealistically low estimate. An increase of only one order of magnitude in viscosity would yield characteristic time constants comparable with the age of the earth. In a more recent numerical study, using temperature- and composition-dependent viscosity, Kincaid [1990] concluded that viscosity, not compositional buoyancy, is responsible for achieving long-term stability. However, this study is suspect since Kincaid [1990] used too low a Rayleigh number to describe the earth’s mantle, and hence his model’s flow produced too large a convective stress. Shapiro et al. [1991] further demonstrated that a viscosity increase of a factor of about 20 between the TBL and the surrounding mantle is sufficient to maintain stability, regardless (within reasonable bounds) of the amount of compositional buoyancy and even with basal tractions of about 10-20 times that considered to be appropriate for the earth [e.g., Hager and O’Connell, 1981].

There has also been much discussion during the past two decades about the correlation between the earth’s long-wavelength gravity field and surface tectonics, in particular old continents. This discussion bears directly at the tectosphere issue: for example, central to Jordan’s [1975] model of the tectosphere is that there is no substantial correlation between the long-wavelength gravity field and surface tectonics. Using broad spatial averages over selected areas, Turcotte and McAdoo [1979] concluded that there is no systematic difference in the geoid signal between oceanic and continental regions. But, Souriau and Souriau [1983] demonstrated that there is a significant correlation between the geoid (spherical harmonic degrees $l = 3-12$) and the tectonic regionalization of Okal [1977]. Later, from degree-by-degree cor-
relations \( (l = 2-20) \), Richards and Hager [1988] observed a weak association between geoid lows and shields. Most recently, Forte et al. [1995] reported that the geoid correlates significantly (99% confidence) with an ocean-continent function.

Thus, many issues regarding the structure and dynamics of the continental tectosphere remain unresolved, indicating that the introductory quote from Pekeris [1935] could equally well be applied to the state of the continental tectosphere today. As Pekeris [1935] noted in the context of mantle convection, quantitative study is a necessary step. For the continental tectosphere problem, our approach towards resolving the outstanding conflicts is to be quantitative, both with regard to dynamical model calculations and to the evaluation of the correlations of surface tectonics with, separately, seismic evidence and the geopotential. A key added ingredient is our inclusion of sensitivity analyses for the former investigations and proper error analysis for the latter. As our contribution toward resolution of these issues, we address in this thesis the following fundamental questions: (1) What is the relationship between surface tectonics and shear-wave heterogeneity and how deep does this association extend? (2) What is the relationship between surface tectonics and the long-wavelength geoid (and gravity field) and how does this association constrain the relationship between density and shear-wave velocity? and (3) How dynamically stable is a cold, thick, chemically compensated tectosphere and what mechanisms are important in achieving stability?

In Chapter 2 we address the issue of continental deep structure by investigating statistically the relationship between surface tectonics and shear-wave heterogeneity, and carefully considering uncertainties and significance levels associated with our statistical treatment. Using Su et al.'s [1994] degree-12 shear-wave tomographic model, we analyze quantitatively the relationship between vertical S-wave travel-time anomalies (relative to a radial reference earth model [Dziewonski and Anderson, 1981]) and surface tectonics [Jordan, 1981b]. Because both the
regions and the upper-mantle shear-wave heterogeneity have distinctly red spectra, the application of standard statistical methods can yield inappropriate estimates of uncertainties. Instead, we perform Monte Carlo simulations that incorporate the spectral properties of both the regions and the seismic heterogeneity.

In Chapter 3, we investigate quantitatively the significance of the association between surface tectonics and the long-wavelength geoid [Lerch et al., 1994], referred to the hydrostatic figure of the earth [Nakiboglu, 1982]. Using the techniques developed and applied in Chapter 2, we calculate regional averages of the geoid and of the radial gravity field and estimate their uncertainties. Further, we estimate the contribution of continental deep structure to the geoid by subtracting from the geoid estimates from other contributors: (1) a simplified representation of the upper 120 km based on the oceanic plate cooling model and a uniform 35-km-thick continental crust [Hager, 1983]; (2) the lower mantle [Hager and Clayton, 1989]; (3) slabs [Hager and Clayton, 1989]; and (4) remnant glacial isostatic disequilibrium [Hager et al., 1984]. By combining the upper-mantle shear-wave velocity anomalies associated with platforms and shields (Chapter 2), and the results from this chapter, we estimate and place bounds on the corresponding average values of $\partial \ln \rho / \partial \ln \nu$, within the depth range 140-440 km.

Conclusions drawn from studies of boundary-layer dynamics depend strongly on assumptions concerning the composition, temperature, and stress dependence of viscosity and the magnitude of the basal tractions. In Chapter 4, we design our experiments to model realistic characteristic tractions associated with the buoyancy within the continental tectosphere and the basal tractions resulting from mantle convection. Within this stress regime, we use our modified version of the fully dynamic finite-element program, ConMan [King et al., 1990], to solve numerically the advection-diffusion equations for flow of an incompressible, infinite Prandtl number fluid in a two-dimensional Cartesian domain. We initiate our
experiments with a continental tectosphere (CBL and TBL) and consider separately the effects on the stability of the tectosphere of (1) activation energy (used to define the temperature-dependence of viscosity), (2) compositional buoyancy, and (3) linear or nonlinear rheology.

In Chapter 5, we evaluate each of the above models of the continental tectosphere in the context of our results. In Appendix A, we provide a more detailed description of the Monte Carlo simulations and the statistical analyses presented in Chapters 2 and 3. We also include many additional figures which give a complete representation of the results obtained in these chapters. In Appendix B, we present (1) an analytical approach to investigating the stability of the continental tectosphere, (2) a numerical stability study, which does not include temperature- or stress-dependent viscosity, whose results complement those obtained from the experiments discussed in Chapter 4, (3) additional figures associated with the experiments discussed in Chapter 4, and (4) a discussion of the effect of boundary conditions, geometry, and grid resolution on the results given in Chapter 4. Because we will submit Chapters 2, 3, and 4 for journal publication, we present them in this thesis in preprint form.
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CHAPTER 2

SURFACE TECTONICS AND UPPER-MANTLE SHEAR-WAVE HETEROGENEITY

Abstract. The relationship between surface tectonics and upper-mantle seismic heterogeneity is central to our understanding of upper mantle structure, particularly the continental tectosphere. To establish the significance of coupling between surface tectonics and the seismic signature of the continental tectosphere, we examine, as a function of depth, the relationship between a tectonic regionalization and upper-mantle shear-wave travel-time anomalies predicted by a recent seismic tomographic model. Standard statistical analyses are based on the assumption that the relevant data have white spectra. Because both the regions and the upper-mantle shear-wave heterogeneity have distinctly red spectra, the application of such statistical methods yields inappropriate estimates of uncertainties. Instead, we perform Monte Carlo simulations that incorporate the spectral properties of both the regions and the seismic heterogeneity. Our results indicate that (1) ridges (young oceans) can be readily distinguished by their seismic signature from older oceans to a depth of about 200 km and (2) the platform and shield signature differs significantly (> 99% confidence) from that under oceans and orogenic zones to at least 400 km depth.

INTRODUCTION

The depth extent and the nature of differences between oceans and continents have been subjects of heated debate during the last three decades. Analyses of differences in travel times through continents and oceans of nearly vertically propagating seismic shear waves (ScS) provided evidence both for [Sipkin and
Jordan, 1975, 1976, 1980] and against [Okal and Anderson, 1975; Anderson, 1979] deep (= 400 km) continental structure. Lerner-Lam and Jordan [1987] used the degree-eight tomographic model M84C [Woodhouse and Dziewonski, 1984] to investigate the relationship between shear-wave travel-time anomalies and the global tectonic regionalization GTR1 [Jordan, 1981]. From the mean values of anomalies associated with the six tectonic regions, they showed that, within the 40-400 km depth interval, oceans and continents are seismically distinct. In particular, they found that mean travel time anomalies decrease monotonically with age from young oceans to old continents. However, Polet and Anderson [1995] examined two more recent tomographic models [Zhang and Tanimoto, 1993; Grand, 1994] and proposed that at about the one standard deviation level, seismic velocities corresponding with old cratons do not differ from the global average below a depth of about 250 km. Additionally, they found no significant correlation between high velocity seismic anomalies and young cratons at any depth.

Using the recent global seismic tomographic model, S12_WM13 (a successor to M84C) [Su et al., 1994], we analyze quantitatively the relationship between vertical S-wave travel-time anomalies (relative to PREM [Dziewonski and Anderson, 1981]) and surface tectonics by considering separate depth intervals, calculating regional averages and estimating their uncertainties, and discussing the significance of the surface projections of the tomographic model. In addition, we use this tomographic-tectonics study to present a better approach for estimating uncertainties when analyzing data defined on a sphere. This method is particularly useful when the data do not have white spectra.

We select S12_WM13 because (1) it is a global model unlike, for example, Grand [1994] and (2) it has a higher resolution (spherical-harmonic degree-12, represented radially by 13 Chebyshev polynomials) than other global models (e.g., SH.10c.17 [Masters et al., 1992], MDLSH [Tanimoto, 1990], M84C [Woodhouse
and Dziewonski, 1984]. Although Zhang and Tanimoto [1992, 1993] provide a degree-36 global model, the applicability of this model is suspect—the magnitude of the heterogeneity at depths greater than about 100 km is much lower than that given in S12_WM13 and in Grand's [1994] tomographic model and yields predictions of SS and SS-S travel-time residuals which differ significantly from those observed [Su et al., 1992]. To test the sensitivity of our results to our selection of the tomographic model, we compare our conclusions (see Appendix A) with those we obtain using SH.10c.17 [Masters et al., 1992] (spherical-harmonic degree 10, 11 radial layers).

**TECTONIC REGIONALIZATION AND INVERSION**

To represent the surface tectonics, we use GTR1. GTR1 contains six regions defined on the earth's surface on a grid with 5° x 5° cells. This grid spacing gives about six points per wavelength at degree 12 and thus provides an appropriate spatial sampling. The three oceanic regions (including marginal basins) are categorized by crustal age, whereas the continental classifications are based on their tectonic behavior during the Phanerozoic. Other regionalizations from Okal [1977] and Mauk [1977] contain seven and 20 regions, respectively, and are classified differently from both GTR1 and each other. Okal's [1977] model is discretized on a 10° x 15° to 15° x 15° grid, but its major limitation is in the accuracy of its designation of regions, which appears worse than would be expected given this rather coarse discretization. For example, Okal [1977] labels the entire continent of Antarctica a shield, whereas a significant fraction (= 1/3) is orogenic in nature. Okal [1977] also classifies some islands (e.g., Iceland and Great Britain) as shields. Misidentifications such as these can have a profound effect on results from associated data projections. Due to the resolution limit associated with our degree-12 seismic tomographic model, the finer resolution inherent in Mauk's [1977] re-
Regionalization does not offer us any advantage. In fact, as we will show, within no 100-km depth interval can we even distinguish the six (much larger) regions represented in GTR1. Through a representative projection, we demonstrate that (1) the results we obtain using Mauk's [1977] regionalization and GTR1 are equivalent and (2) that Mauk's [1977] finer classification of continental geology does not result in a significantly improved fit with the seismology (Appendix A).

By combining regions of GTR1, we can construct other, coarser regionalizations which, in some cases, represent the limit of our resolution in comparing with travel-time anomalies. For example, by consolidating young oceans (A; 0-25 My), intermediate-age oceans (B; 25-100 My), and old oceans (C; >100 My), into one region, and Phanerozoic orogenic zones (Q), Phanerozoic platforms (P), and Precambrian shields and platforms (S), into another region, we can create a two-component (ocean-continent) tectonic regionalization (ABC, QPS). In general, a tectonic regionalization containing \( N \) distinct regions can be described by \( N \) functions, \( R_n (n = 1, N) \), each having unit value over its region and zero elsewhere.

We expand each \( R_n \) in spherical harmonics, omitting degree zero from our analysis because we are interested only in lateral variations in the seismic structure. With coefficient \( R_n^{lm} \) representing the \((l,m)\) harmonic of region \( n \) and coefficient \( d_n^{lm} \) representing the \((l,m)\) harmonic of the shear-wave travel-time anomalies, we use a least-squares approach to solve \( R_n^{lm} \tau_n = d_n^{lm} \) (summation convention implied here and below) for the regional averages, \( \tau_n \), of the data. We include the additional constraint that \( A_n \tau_n = 0 \), where \( A_n \) represents the surface area spanned by region \( n \). This constraint ensures that the \( \tau_n \) have a zero (weighted) average, as, by definition, do the shear-wave travel-time anomalies. Our weighted-least-squares solution can be written as \( \tau = [R^TWR]^{-1}R^TWd \), where the values \( R_n^{lm} \) and \( A_n \) are the elements of the matrix \( R \), \( W \) is a diagonal weight matrix constructed from the diagonal covariance matrix associated with S12_WM13, \( \tau_n \) are the elements of the vector \( \tau \).
and \( d^m \) and zero constitute the vector \( d \). Obviously, there are non-zero off-diagonal elements in \( W \), however they are inaccessible. We apply a large weight to the constraint; results from our inversions are insensitive to the value of this weight so long as it is not less than ten times the maximum weight associated with any datum nor so large (\( > 10^6 \) times the maximum weight) that the inversion becomes numerically unstable. As a criterion for the success of the model in fitting the data, we use the percent fractional difference in the prefit and postfit chi-squares. The percent variance reduction associated with each inversion is thus defined by 

\[
100 \left[ 1 - \frac{\chi^2_{\text{post}}}{\chi^2_{\text{pre}}} \right].
\]

SURFACE TECTONICS AND DEEP SEISMIC STRUCTURE

We project separately onto GTR1 one-way travel-time perturbations from a suite of adjacent 100 km thick intervals, beginning at 40 km and extending to 1540 km depth. Using a method similar to that presented by Souriau and Souriau [1983], we use an Euler angle approach to choose 10,000 orientations randomly from a uniform distribution over the sphere and then correspondingly rotate the sphere on which the surface tectonics are defined ("tectonic sphere") with respect to the sphere on which the data are defined ("data sphere"). From each such set of random rotations, we estimate the significance level in the variance reduction associated with the actual projection. Results from these Monte Carlo simulations (Figure 2.1) demonstrate the implausibility that the association between the tomography within the 40-440 km depth range and the surface tectonics is fortuitous. Consider these four 100-km intervals. In each case, the actual orientation of the tectonic sphere is the one which yields the maximum variance reduction, except within the 40-140 km depth interval, where the actual orientation is within a few degrees of the orientation associated with the maximum variance reduction. For that interval, the variance reduction associated with the actual orientation ("actual
variance reduction") is within 0.02% of the maximum variance reduction obtained from the 10,000 random rotations of the tectonic sphere. The actual variance reduction and the maximum variance reduction from the simulations diverge within the 440-540 km depth interval. Within each of the next five depth intervals, the actual variance reduction indicates a significant (better than 95% confidence level) coherence between the tomography and our ocean-continent function. However, with this analysis we are unable (at any depth) to address directly the issue of separating a real tectonic signal in the tomography from one caused by the smoothness constraint inherent in the use of Chebyshev polynomials to model radial variations in shear-wave velocities. Specifically, an imperfect crustal correction combined with this smoothness constraint could cause such a (crustal) signal to be projected well into the mantle. Nonetheless, the very strong results obtained for the first four layers, compared to the simulations, suggest that these constraints are not the main contributors to these results. (See Ritzwoller and Lavely [1995] for a detailed comparison of several tomographic models, including a brief discussion of their sensitivity to different models of the crust.)

Using a 1% F-test criterion, we identify from the set of regionalizations obtainable from combinations of the six components of GTR1 the regionalizations which represent the resolution limit of each projection. For the first interval (40-140 km) through the fifth (440-540 km), \((A,b,C_P,Q,S)\), \((A,b,C_Q,P_S)\), \((A b,C,Q,P_S)\), \((A b,C,Q,P_S)\), and \((A b,C,Q,P_S)\), respectively, represent the regionalizations with the largest number of distinct regions that yield statistically improved results. As indicated by the selections of these regionalizations, we can discern differences in the seismic properties between young, mid-age, and old oceans through the 140-240 km depth interval, below which these regions are indistinguishable from one another. In other words, we estimate that ridges extend to about 200 km. Shields represent either a discrete region or are grouped with plat-
forms. Platforms, when not linked with shields, are placed, separately, with old oceans and orogenic zones. Below 540 km, the two-region ocean-continent function \(ABC, QPS\) represents the resolution limit of the projections.

**PARAMETER UNCERTAINITIES AND PROJECTIONS**

Estimates of parameter uncertainties depend on assumptions that we make about the data. In particular, because the tomography and the regions have power spectra that are dominated by low-frequency terms (see Figure 2.2 for a typical example), we obtain different uncertainty estimates depending on whether we assume that the tomography is a member of a population of models defined by (1) its total power or (2) its power spectrum. Assumption (1) could correspond to a white-noise spectrum and would lead to parameter uncertainties consistent with the statistical uncertainties that we obtain from our inversions. However, the distinctly red spectra of both the regions and the shear-wave heterogeneity lead, in general, to larger fluctuations in the correlations and hence to larger uncertainties than would white spectra. We assume that the earth's tectonics and its shear-wave heterogeneity are samples from distributions with correspondingly red spectra. Hence to estimate uncertainties, we use the standard deviation in the values of the \(\tau^*\) obtained from a large set of trials, each based on a random rotation (see above) of the tectonic sphere with respect to the data residual sphere \(d_{res}^{lm} = d^{lm} - R_s^{lm} \tau_s\); thus each \(\tau^*_s\) is determined with respect to \(d_{res}^{lm}\) as \(\tau_s\) was with respect to \(d^{lm}\). We assume that \(d_{res}^{lm}\) represents the noise part of \(d^{lm}\), with the signal components being contained entirely in \(R_s^{lm} \tau_s\). Figure 2.3, as a representative example, shows the relevant histograms corresponding to the 240-340 km depth interval. Table 2.1 contains the regional averages and corresponding uncertainties (based on assumption (2)) associated with the upper five depth intervals. Uncertainty estimates based on assumption (1) are smaller by factors of between about 1.25 and 2.0.
The regional averages (Table 2.1) decrease monotonically with increasing tectonic age, except within the 40-140 km depth interval, where the old ocean's cold lithosphere apparently contributes to its platform-like average. The old-ocean average changes sign between the 140-240 and 240-340 km depth intervals, presumably as a result of the diminishing effect of plate cooling. The young-ocean average decreases monotonically, whereas the intermediate-age ocean average appears to peak between 140 and 340 km. Since S12_WM13 is parameterized radially with 13 orthonormal Chebyshev polynomials from the core-mantle boundary to 24.4 km below the surface, our 100-km thick layers somewhat "over resolve" the seismic model.

The one-way shear-wave travel-time anomalies associated with the 40-140 km, 140-240 km, and 240-340 km depth intervals are shown in Figures 2.4a, 2.5a, and 2.6a, respectively. Figures 2.4b, 2.5b, and 2.6b display the corresponding projections of these seismic anomalies onto regionalizations (A, B, CP, Q, S), (A, B, CQ, PS), (ABC, Q, PS). Subtracting projections from data yields plots of the residual seismic signal (Figures 2.4c, 2.5c, and 2.6c). Figure 2.4 illustrates well the age-dependent progression of anomalous seismic velocity beneath the Pacific ocean. The shear-wave velocity is much slower than average beneath young oceans and increases in speed as the oceanic plate ages. Further, Figures 2.4c, 2.5c, and 2.6c indicate that the region beneath the Atlantic ocean is roughly 1% seismically "faster" and the region under the Pacific ocean is roughly 1% seismically "slower" than our model predicts. Given the small uncertainties of the parameter estimates (Table 2.1), this difference appears significant. The surface expression of the shear-wave heterogeneity within the 240-340 km depth interval (Figure 2.6) illustrates clearly that the link with oceans and continents extends through this depth interval.
The greatest disagreement between the tomography and our model occurs in Northeastern Africa, near the Afar triple junction. This "bull's eye" is seen clearly down through the 340-440 km depth interval and suggests that the upper-mantle beneath this rift zone more closely resembles mantle beneath oceans or Phanerozoic orogenic zones than it does mantle associated with platforms and shields.

**DISCUSSION**

Analyzing the spatial relationship between two data fields defined on the surface of a sphere is a process which is at the heart of many studies of the earth. When the relevant data are described by non-white spectra, standard statistical methods based on the assumption of white noise are inappropriate. Monte Carlo simulations, which incorporate the spectral properties of the relevant data, are better suited to obtain estimates of uncertainties of spatial averages. These simulations can be based on the random rotation of one data sphere with respect to the other, or can be accomplished by assigning to one data field random values of spherical harmonic coefficients while maintaining the power spectrum of the data (see Appendix A).

Our analysis indicates that there is a very significant correlation between surface tectonics and the upper-mantle shear-wave heterogeneity predicted by S12_WM13, as is summarized by the regional averages and uncertainties listed in Table 2.1. We calculate the chi-square per degree of freedom for each row to estimate the probability that these results would occur as a sample from a normal distribution. For every depth interval, the corresponding tables of the chi-square distribution show that these chi-square values will occur in a random sample with probabilities that each are less than 0.1%.

The results from our study are generally consistent with those of Lerner-Lam and Jordan [1987]. Through our statistical treatment, we evaluate our ability
to distinguish tectonic provinces in the shear-wave travel-time anomalies as a function of depth. Most importantly, our estimates of parameter uncertainties allow us to assess the significance of the differences in regional averages. When we repeat Lerner-Lam and Jordan's [1987] study using the 40-400 km depth interval and apply our statistical machinery, we find that the six regions of GTR1 are not all distinguishable. Using a 1% F-test, we find conclude that only the four regions — A, B, CQ, and PS — are discernible from travel-time anomalies within this depth interval.

Our results are inconsistent with those from the recent study of Polet and Anderson [1995]. We use (1) different statistical techniques, (2) different tomographic models with different resolutions and coverage, and (3) different regionalizations from the ones that they used. Although the information given by Polet and Anderson [1995] is insufficient to allow us to pinpoint the contributions of each of these differences to the overall disparity in results, we can make some general comments. First, the error bars accompanying their average velocity anomalies, and hence their conclusions, appear to be the standard deviation of the individual velocity anomalies associated with each tectonic region and depth interval, rather than the (appropriate) standard deviation of the mean. Second, as discussed above, Zhang and Tanimoto's [1992, 1993] model is inconsistent with observations of SS and SS-S travel-time residuals [Su et al., 1992]. Further, Grand's [1994] model has a limited geographic coverage, primarily centered on the Atlantic hemisphere, and averages are biased due to this incomplete coverage. Because Grand's [1994] model does not include some major cratonic regions (e.g., Australia and Antarctica), the regional averages are weighted more heavily by those locations that are included; moreover, regions associated with the periphery of the model's coverage, and also the model's sensitivity to these regions (e.g., Western Africa), could introduce further distortions. Within the 240-340 km depth interval, oceans are
slow and continents are fast (Table 2.1); but, relative to these global regional averages, the Atlantic is anomalously fast and Africa is anomalously slow (Figure 2.6c). Hence the dichotomy between oceanic and continental shear-wave velocities is diluted in Grand's [1994] model by these anomalous signals and extrapolations to global regional averages are misleading.

In summary, robust error estimation is essential for many fields of geophysical inference. Underestimating errors by assuming an inappropriate error spectrum may lead to interpretations of differences that are not real. But overestimating errors can lead to the mistake of concluding that real differences are insignificant. We have demonstrated one approach to error estimation that allows us to conclude that in S12_WM13, the expression of high seismic velocities beneath platforms and shields extends to at least 400 km depth.

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REFERENCES

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Table 2.1. S12_WM13 ($l = 1-12$): Regional averages and statistical standard errors corresponding to one-way S-wave travel-time anomalies. Statistical standard errors are from Monte Carlo simulations.

<table>
<thead>
<tr>
<th>(km)</th>
<th>$\tau_A$ (%)</th>
<th>$\tau_B$ (%)</th>
<th>$\tau_C$ (%)</th>
<th>$\tau_Q$ (%)</th>
<th>$\tau_P$ (%)</th>
<th>$\tau_S$ (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>40-140</td>
<td>2.7 ± 0.3</td>
<td>0.8 ± 0.2</td>
<td>-1.4 ± 0.3</td>
<td>-0.1 ± 0.3</td>
<td>-1.4 ± 0.3</td>
<td>-4.2 ± 0.6</td>
</tr>
<tr>
<td>140-240</td>
<td>1.8 ± 0.3</td>
<td>0.9 ± 0.1</td>
<td>-0.4 ± 0.1</td>
<td>-0.4 ± 0.1</td>
<td>-2.3 ± 0.2</td>
<td>-2.3 ± 0.2</td>
</tr>
<tr>
<td>240-340</td>
<td>0.7 ± 0.1</td>
<td>0.7 ± 0.1</td>
<td>0.7 ± 0.1</td>
<td>-0.7 ± 0.2</td>
<td>-1.6 ± 0.2</td>
<td>-1.6 ± 0.2</td>
</tr>
<tr>
<td>340-440</td>
<td>0.4 ± 0.1</td>
<td>0.4 ± 0.1</td>
<td>0.4 ± 0.1</td>
<td>-0.4 ± 0.2</td>
<td>-1.0 ± 0.2</td>
<td>-1.0 ± 0.2</td>
</tr>
<tr>
<td>440-540</td>
<td>0.2 ± 0.1</td>
<td>0.2 ± 0.1</td>
<td>0.2 ± 0.1</td>
<td>-0.2 ± 0.2</td>
<td>-0.2 ± 0.2</td>
<td>-1.0 ± 0.3</td>
</tr>
</tbody>
</table>
FIGURE CAPTIONS

Fig. 2.1. S12_WM13 (l = 1-12): One-way S-wave travel-time anomalies associated with 100 km-thick depth intervals projected onto GTR1. Open circles indicate the variance reductions associated with the actual orientation of the tectonic sphere. Asterisks represent the maximum variance reduction achieved from 10,000 random rotations of the tectonic sphere (see text for description of random distribution). The horizontal line bisecting each open circle indicates the interval's thickness. From bottom to top, the three regions of gray scale correspond to confidence levels of <75%, 75-95%, and 95-99% as determined from the Monte Carlo simulations.

Fig. 2.2. Global root-mean-square amplitudes, $rms(l) = \left[ \sum_{n} (d_{nm}^{\text{in}})^2 / (2l + 1) \right]^{1/2}$, as a function of $l$, of the travel-time anomalies associated with the 240-340 km depth interval (in percent) and analogous values for regions ABC, Q, and PS. S12_WM13 is represented by a line, ABC by a dash-dotted line, Q by a dotted line, and PS by a dashed line.

Fig. 2.3. Histograms of parameter values (a) $\tau_{ABC}$, (b) $\tau_{Q}$, (c) $\tau_{PS}$, obtained from projections of the 240-340 km data-residual sphere onto 10,000 random orientations of the tectonic sphere. Histograms are centered on the parameter values given in Table 2.1. Gaussian distributions, determined by the standard deviation, mean, and area of each histogram, are superposed. (d) Histogram of variance reductions obtained from these projections of the data onto 10,000 random orientations of the tectonic sphere. The variance reduction corresponding to the actual orientation of the tectonic sphere is indicated with an arrow.

Fig. 2.4. (a) One-way S-wave travel-time anomalies from S12_WM13 (l = 1-12) associated with the 40-140 km depth interval, (b) Projection of (a) onto (A, B, CP,
\(Q, S\), and (c) Residual: \((a) - (b)\). Positive contour lines are dashed and the zero contour line is thick. The contour interval is 1% of the mean travel time.

Fig. 2.5. (a) One-way \(S\)-wave travel-time anomalies from S12_WM13 \((l = 1-12)\) associated with the 140-240 km depth interval, (b) Projection of (a) onto \((A, B, CQ, PS)\), and (c) Residual: \((a) - (b)\). Positive contour lines are dashed and the zero contour line is thick. The contour interval is 1% of the mean travel time.

Fig. 2.6. (a) One-way \(S\)-wave travel-time anomalies from S12_WM13 \((l = 1-12)\) associated with the 240-340 km depth interval, (b) Projection of (a) onto \((ABC, Q, PS)\), and (c) Residual: \((a) - (b)\). Positive contour lines are dashed and the zero contour line is thick. The contour interval is 1% of the mean travel time.
Figure 2.1
Figure 2.2
Figure 2.3
Figure 2.4
CHAPTER 3

THE CONTINENTAL TECTOSPHERE AND THE
LONG-WAVELENGTH GRAVITY FIELD

Abstract. To estimate the average density contrast associated with the continental tectosphere, we separately project the long-wavelength non-hydrostatic geoid and the radial component of the gravity field onto a tectonic regionalization. Because both the regionalization and the geoid have distinctly red spectra, we do not use conventional statistical analyses based on assumptions of white spectra. Rather, we utilize a Monte Carlo approach which incorporates the spectral properties of these two fields. Results from these simulations reveal that the earth's gravity variations correlate with surface tectonics no better than they would were the geoid (or gravity field) randomly oriented with respect to the surface. The average geoid anomaly and perturbation to the gravitational acceleration over platforms and shields are $-13 \pm 11$ m and $-9 \pm 7$ mgal, respectively. After removing from the geoid the contributions associated with (1) a simple model of the upper 120 km, (2) the lower mantle, (3) slabs, and (4) remnant glacial isostatic disequilibrium, we estimate for the upper mantle a platform and shield signal of $-8 \pm 5$ m. Thus, we conclude that there is little contribution of platforms and shields to the gravity field, consistent with their keels having small density contrasts. Using this platform and shield signal and previous estimates of upper-mantle shear-wave travel-time perturbations, we find that the average value of $\partial \ln \rho / \partial \ln v_s$ within the 140-440 km depth range is about $0.035 \pm 0.025$. A continental tectosphere with an isopycnic (equal-density) structure ($\partial \ln \rho / \partial \ln v_s = 0$) enforced by compositional variations, is consistent, at the 1.5σ level, with this result. However, without compositional buoyancy, the continental tectosphere would have an average value of $\partial \ln \rho / \partial \ln v_s$. 
of $\approx 0.25$ which, because it exceeds our estimate by almost $9\sigma$, appears inconsistent with the average value of geoid anomalies associated with platforms and shields.

**INTRODUCTION**

Motivated by seismological evidence [e.g., *Sipkin and Jordan, 1975*] and the lack of a strong correlation between continents and the long-wavelength geoid [e.g., *Kaula, 1967*, *Jordan, 1975*] proposed that continents are (1) characterized by thick ($>400$ km) thermal boundary layers (TBLs) which translate coherently during lateral plate motions, (2) stabilized against small-scale convective disruption by gradients in composition, and (3) not observable in the long-wavelength gravity field. The simple plate cooling model, which enjoys much success in describing the structure of oceanic TBLs, can not be extended to explain thicker continental TBLs [*Jordan, 1981a*]. Instead, *Jordan, 1981a* postulated that the thick continental TBL (continental tectosphere) was formed during the Archaean through continental collisions and has been stabilized against convective disruption by the compositional buoyancy provided by the depletion of basaltic constituents incurred during these collisions. The isopycnic (equal-density) hypothesis [*Jordan, 1988*] predicts that the compositional and thermal effects on density exactly cancel at every depth between the base of the mechanical boundary layer and the base of the TBL. Such a structure would be neutrally buoyant with respect to neighboring oceanic mantle and would not be visible in the long-wavelength gravity field.

There has thus been much discussion during the past two decades about the correlation between the earth's long-wavelength gravity field and surface tectonics. In particular, the significance of a continental contribution to undulations of the geoid is a question whose answer is still disputed. Using broad spatial averages over selected areas, *Turcotte and McAdoo, 1979* concluded that there is no sys-
tematic difference in the geoid signal between oceanic and continental regions. But, Souriau and Souriau [1983] demonstrated that there is a significant correlation between the geoid (spherical harmonic degrees $l = 3-12$) and the tectonic regionalization of Okal [1977]. More recently, from degree-by-degree correlations ($l = 2-20$), Richards and Hager [1988] observed a weak association between geoid lows and shields. On the other hand, Forte et al. [1995] reported that the geoid correlates significantly (99% confidence) with an ocean-continent function. Were there a significant ocean-continent signal, the continental tectosphere might have a substantial density anomaly associated with it, and might therefore be expected to play an active role in the large-scale structure of mantle convection.

We investigate quantitatively the significance of the association between the six-region global tectonic regionalization GTR1 [Jordan, 1981b] and the geoid, GEM-T3 [Lerch et al., 1994], referred to the hydrostatic figure of the earth [Nakiboglu, 1982]. Although we will use GTR1 (and coarser regionalizations created by combining some of these regions; see Chapter 2) for the bulk of this study, we will compare our results with those obtained using Mauk's [1977] and Okal's [1977] tectonic regionalizations. Because the geoid spectrum is red and because the longest wavelengths are likely dominated by the effects of density contrasts in the lower mantle [Hager et al., 1985], we also investigate the relationship between GTR1 and the radial component of the earth's gravity field. The gravity field at spherical harmonic degree $l$ is proportional to $l$ times the geoid anomaly at degree $l$, so the gravity field deemphasizes long-wavelength variations.

We calculate regional averages of the geoid and the gravity field and estimate their uncertainties. Further, we estimate the contribution of the continental tectosphere to the geoid by subtracting from the geoid estimates from other contributions. By combining the upper-mantle shear-wave travel-time anomalies associated with platforms and shields (Chapter 2) and the results from this study, we es-
timate and place bounds on the average value of \( \partial \ln \rho / \partial \ln v \), within the depth range 140-440 km and compare our estimate with Jordan's [1988] isopycnic hypothesis.

**INVERSION**

GTR1 contains six regions defined on the earth's surface on a grid with 5° x 5° cells. The three oceanic regions (including marginal basins) are categorized by crustal age: 0-25 My (A), 25-100 My (B), and >100 My (C). The continental regions are classified by their tectonic behavior: Phanerozoic orogenic zones (Q), Phanerozoic platforms (P), and Precambrian shields and platforms (S). In general, a tectonic regionalization containing \( N \) distinct regions can be described by \( N \) functions, \( R_n (n = 1, N) \), each having unit value over its region and zero elsewhere. In this analysis, we combine regions P and S into one region (PS). Because we are interested in geoid variations with length scales characteristic of continents, we consider only harmonic degrees 2-12. The specific value for the upper limit was chosen to parallel the seismic study presented in Chapter 2.

With coefficient \( R_n^{lm} \) representing the \((l,m)\) harmonic of region \( n \) and coefficient \( d^{lm} \) representing the \((l,m)\) harmonic of the geoid-height or gravity field, we use a weighted least-squares approach to solve \( R_n^{lm} \gamma_n = d^{lm} \) (summation convention implied here and below) for the regional averages, \( \gamma_n \), of the data (see Chapter 2, for a more detailed description of the methodology). Ideally, we would like to use the full covariance matrix in our analysis but, since it is unavailable to us, we instead assume that each GEM-T3 coefficient is uncorrelated with the others and known to the same accuracy as the others. Hence, we use the identity matrix as our weight matrix, except for the arbitrarily chosen (1000 times) larger weight applied to the constraint that \( A_n \gamma_n = 0 \). Here \( A_n \) represents the surface area spanned by region \( n \). Our results are insensitive to the weight applied to the constraint within
the range of about 10 to 10^6. The upper bound arises from a numerical limitation in our inversion.

**STATISTICAL ANALYSIS**

As discussed in more detail in Chapter 2, because neither the geoid nor the regionalization have white spectra, we do not use common statistical estimates of uncertainties. In fact, their spectra are quite red, implying that uncertainties based on the assumption of white spectra will be substantially smaller than the actual uncertainties. Through the use of Monte Carlo techniques, we incorporate the spectral properties of these fields in our estimates of uncertainties. For each of 10,000 trials, we (1) randomly select Euler angle triples (from a parent distribution in which all pole orientations are equally probable) and then rigidly rotate the sphere on which the data residuals \( d^\text{im}_\text{res} = d^\text{im} - R^\text{im}_n \gamma_n \) are defined with respect to the sphere on which the surface tectonics are defined ("tectonic sphere"), (2) combine the rotated data residuals \( d^\text{im}_\text{res} \) ("\( \sim \)" means rotated) with the correlated data \( \bar{d}^\text{im} = \bar{d}^\text{im}_\text{res} + R^\text{im}_n \gamma_n \), and (3) project \( \bar{d}^\text{im} \) onto \((A, B, C, Q, PS)\). We take as the uncertainties in the parameter estimates, the standard deviations in the values of the regional averages obtained from these simulations. Because the correlated signal is added to the rotated data residual before projecting the composite, the resulting histograms of parameter values (see Figure 3.1 for a representative sample) are centered approximately on the parameter values associated with the actual orientation of the tectonic sphere.

As a criterion for the success of the model in fitting the data, we use the percent fractional difference in the prefit and postfit chi-squares. The percent variance reduction associated with each inversion is thus defined by

\[
100 \left[ 1 - \left( \frac{\chi^2_{\text{post}}}{\chi^2_{\text{pre}}} \right) \right].
\]

By randomly rotating the sphere on which the data are defined ("data sphere") with
respect to the tectonic sphere, we estimate significance levels in the variance reduction associated with each projection.

**PROJECTIONS**

Table 3.1 shows the regional averages and their corresponding statistical standard errors obtained by separately projecting the geoid and the radial component of the gravity field onto \((A, B, C, Q, PS)\). With the geoid, only regions \((C)\) and \((PS)\) have averages which are larger than their standard errors. However, the significance of these averages is only slightly above the one-standard-deviation level. For example, with 95\% (2\σ) confidence, the geoid signature associated with platforms and shields is in the range -35 to +9 m, a rather broad constraint. Figure 3.1f further demonstrates the weak association between the long-wavelength geoid and \((A, B, C, Q, PS)\): the actual orientation of the geoid explains more of the variance than only about one third of the random orientations of the data sphere. Using the long-wavelength radial component of the gravity field yields a similar null result (Figure 3.2). From Figure 3.3a, it is clear that continents (as well as oceans) are associated with both positive and negative geoid anomalies (e.g., Australia, Africa, and South America - positive; Antarctica and North America - negative). Consequently, very little of the long-wavelength non-hydrostatic geoid can be explained simply in terms of surface tectonics (Figure 3.3b). The magnitude of the geoid signal that is uncorrelated with \((A, B, C, Q, PS)\) (Figure 3.3c) is essentially the same as that of the geoid anomalies themselves, given by GEM-T3.

To attempt to isolate the tectosphere’s contribution to the long-wavelength geoid, we subtract from the geoid the effects of previously modeled components: (1) a simplified representation of the upper 120 km based on the oceanic plate cooling model and a uniform 35-km-thick continental crust [Hager, 1983]; (2) the lower mantle [Hager and Clayton, 1989]; (3) slabs [Hager and Clayton, 1989]; and
(4) Remnant glacial isostatic disequilibrium [Hager et al., 1984]. Separately projecting each of these four contributions to the model geoid onto \((A, B, C, Q, PS)\) yields the results given in Tables 3.2 and 3.3. Our resulting model (residual) geoid, TECT-1, provides an estimate of the contributions to the geoid of the upper mantle structure below 120 km depth, excluding subducted slabs. For TECT-1, 

\[
\gamma_{ps} = -8 \pm 5 \text{ m (Table 3.2),}
\]

where \(\gamma\) denotes a regional average. As with GEM-T3, TECT-1 is associated with both continental geoid highs and lows (Figure 3.4a) and hence the projection of TECT-1 onto \((A, B, C, Q, PS)\) results in a very weak signal (Figure 3.4b). TECT-1 and the component of TECT-1 which is uncorrelated with \((A, B, C, Q, PS)\) (Figure 3.4c) are nearly the same.

The projections of TECT-1 separately onto \((A, B, C, Q, PS), (ABC, QPS)\), and \((ABCQ, PS)\) lead to reductions in variance that are listed in Table 3.3. From the percent of random trials that yield smaller variance reduction than that of the actual orientation (confidence level), it is clear that the geoid signal represented by TECT-1 is, among these choices, best represented by the two-region regionalization: \((ABCQ, PS)\). Although the projection of TECT-1 onto \((ABCQ, PS)\) results in a variance reduction of only about 2.6%, this value exceeds those obtained from 85% of the projections associated with random rotations of the data sphere. This result is consistent with the 1.5\(\sigma\) result associated with the platform and shield signal represented in TECT-1, but contrasts markedly with the results for the five-region grouping \((A, B, C, Q, PS)\), where the actual orientation of the data sphere explains more of the variance than only 35% of the random orientations. This apparent discrepancy arises because random orientations of the other tectonic regions can "lock on" to regional features in the geoid such as those associated with subduction zones, providing a better fit to the synthetic geoids globally, but not in regions spanned by the projection of \(PS\).
ESTIMATE OF $\frac{\partial \ln \rho}{\partial \ln v_s}$

The isostatic geoid height anomaly, $\delta N$, associated with static density anomalies can be calculated for each lateral location from:

$$\delta N = -\frac{2 \pi G}{g} \int \Delta \rho(z) z dz$$

where $G$ is the universal gravitational constant, $g$ is the acceleration due to gravity, and $\Delta \rho(z)$ is the anomalous density at depth $z$. The integration extends from the surface to the assumed depth of compensation. Assuming that $\frac{\partial \ln \rho}{\partial \ln v_s} = \text{constant}$ within a specified depth interval, we may write the scaling between fractional perturbations in density and shear-wave velocity as:

$$\frac{\Delta \rho}{\bar{\rho}} = -\left( \frac{\partial \ln \rho}{\partial \ln v_s} \right) \left( \frac{\Delta \tau}{\bar{\tau}} \right)$$

where $\bar{\rho}$ is obtained, for example, from the radial earth model PREM [Dziewonski and Anderson, 1981], and the fractional perturbations in shear wave velocity $\Delta v_s/\bar{v}_s$ are equal to the negative of the fractional travel-time perturbations $\Delta \tau/\bar{\tau}$, for small perturbations. We base the subsequent calculation on a depth of compensation of 440 km. Below this depth, we assume that there is no platform and shield contribution to the geoid, as there is no significant distinction at such depths between the shear-wave signal beneath platforms and shields and the global average (Chapter 2).

Given regional averages of one-way shear-wave travel-time perturbations associated with 100-km-thick intervals between 140-440 km depth, we can express the integral of the depth-dependent density anomaly as the sum of the anomalies associated with the various layers. Using the travel-time perturbations given in Chapter 2 (reproduced in Table 3.4) and $\delta N_{ps} \equiv \gamma_{ps} = -8 \pm 5$ m, we find that for platform and shields, the average value of $\frac{\partial \ln \rho}{\partial \ln v_s}$ is about $0.035 \pm 0.025$. (This estimate of standard error is based only on that of $\delta N_{ps}$. The uncertainties associated with the regionally averaged travel-time perturbations have a much
smaller effect on the value of $\partial \ln p / \partial \ln v$, than the uncertainty associated with the
geoid and were therefore ignored.)

**DISCUSSION**

How likely are the results in Table 3.1 to occur as a sample from a normal
distribution? To address this question, we calculate the chi-square per degree of
freedom for each of the two rows. These values will occur in a sample from a
normal distribution with a probability greater than 50% and 70%, respectively, for
the geoid and the gravity analyses. Further, none of the projections based on (1)
the non-hydrostatic geoid, (2) the radial component of the gravity field, or (3) our
model geoid, TECT-1, yields a platform and shield signal which is significant at a
level exceeding about $1.6\sigma$.

This conclusion differs substantially from the highly significant (99% con-
fidence) correlation, reported by Forte et al. [1995], between an ocean-continent
function and the non-hydrostatic long-wavelength ($l = 2-8$) geoid. However, a cor-
relation coefficient ($r$) between different fields defined on a sphere is only mean-
ful (subject to tests of significance) for fields with non-white spectra if correlation
coefficients are determined separately for each spherical harmonic degree of interest
[Eckhardt, 1984]. Given the appropriate number of degrees of freedom associated
with the correlation, one can nonetheless estimate the confidence level correspond-
ing to the assumption that the true correlation is zero. Therefore, we estimate the
effective number of degrees of freedom in Forte et al.'s [1995] analysis and, using
this value, estimate the probability that the correlation they obtained is significantly
different from zero.

Under the conditions outlined above, we can estimate the effective number
of degrees of freedom using Student's $t$ distribution. For uncorrelated fields, the
quantity $t = r \left[ v/(1 - r^2) \right]^{1/2}$ can be described by Student's $t$ distribution with $v$ de-
degrees of freedom [e.g., Cramer, 1946; see, too, O'Connell, 1971]. We create 10,000 degree-eight fields, each with the same spectral properties as the non-hydrostatic geoid, by randomly selecting coefficients from a uniform distribution and then scaling them degree-by-degree so that the power spectrum of each "synthetic" field matches that of the geoid. From these synthetic fields and an ocean-continent function derived from GTR1, we generate a collection of 10,000 correlation coefficients (Figure 3.5a). We then estimate $\nu$ by minimizing the chi-square in the fit of Student's distribution to this set of correlation coefficients (Figure 3.5b). Figures 3.5c, 3.5d, and 3.5e demonstrate the sensitivity of the fits to the value of $\nu$. As shown, values of $\nu$ which differ from the selected value ($\nu = 30$) by even five degrees of freedom, noticeably degrade the fit.

The correlation coefficient corresponding to GEM-T3 ($l = 2-8$) is -0.18. However, using an ocean-continent function derived from Mauk's [1977] $5^\circ \times 5^\circ$ tectonic regionalization, we obtain Forte et al.'s [1995] value -0.28. Simulations like those described above yield 31 as the estimate of the effective number of degrees of freedom. The significance levels of the correlations associated with the GTR1 and Mauk [1977] ocean-continent functions are, respectively, about 85% and 95%. The dominant degree-two term in the geoid governs the correlation and highlights a difficulty associated with attaching significance to the correlations between such fields. For example, if one considers only degrees $l = 3-8$, the significance levels of the correlations associated with the GTR1 and Mauk [1977] ocean-continent functions reduce to about 55% and 60%, respectively, and hence indicate insignificant correlations.

Our conclusion also differs substantially from that of Souriau and Souriau [1983] who, also using a Monte Carlo scheme based on random rotations of the data sphere with respect to the tectonic sphere, found that the non-hydrostatic geoid ($l = 3-12$) correlates significantly (at the 95% confidence level) with surface tecton-
ics defined by Okal [1977]. The close geoid-tectonic association presented by Souriau and Souriau [1983] is partially related to the fact that Okal's [1977] regionalization includes subduction zones; the association between the geoid and this regionalization is a result of the strong geoid-slab correlation [e.g., Hager, 1984]. Unlike in our study, Souriau and Souriau [1983] perform their projections in the spatial rather than in the spherical harmonic domain. After reproducing their results, we repeat the suite of projections in the spherical-harmonic domain. We find that the correlation between the long-wavelength geoid ($l = 3-12$) and Okal's [1977] regionalization is significant at about the 98% confidence level, slightly higher than Souriau and Souriau's [1983] result of about 95% from a spatial-domain analysis. However, when we substitute a slab-residual model geoid [Hager and Clayton, 1989] for GEM-T3, we find that the significance reduces to about 50%, indicating that the signal observed by Souriau and Souriau [1983] is largely due to the correlation between slabs and the regionalization.

The isopycnic hypothesis [Jordan, 1988] predicts a value of zero for $\partial \ln \rho / \partial \ln v_z$. This value is within $1.5\sigma$ of our estimate and indicates that at this level of significance, the isopycnic hypothesis is consistent with the average geoid anomaly associated with platforms and shields. On the other hand, we can estimate the value of $\partial \ln \rho / \partial \ln v_z$ by considering only thermal effects on density:

$$\frac{\partial \ln \rho}{\partial \ln v_z} \approx \frac{(1/\rho)(\delta \rho / \delta T)}{(1/v_z)(\delta v_z / \delta T)}$$

Using a coefficient of volume expansion of $3 \times 10^{-5}$ K$^{-1}$, we make two estimates: (1) $\partial \ln \rho / \partial \ln v_z \approx 0.23$, using $\delta v_z / \delta T \approx -0.6$ m s$^{-1}$ K$^{-1}$ from McNutt and Judge [1990] and an average upper-mantle shear velocity of $v_z = 4.5$ km s$^{-1}$, and (2) $\partial \ln \rho / \partial \ln v_z = 0.27$, using $(1/v_z)(\delta v_z / \delta T) = -1.1 \times 10^4$ K$^{-1}$ from Nataf and Ricard [1995]. The average of these estimates is inconsistent at about the 9$\sigma$ level with the value of $\partial \ln \rho / \partial \ln v_z$ that we estimate for the continental tectosphere. Hence, our
analysis indicates that a simple conversion of shear-wave velocity to density via temperature dependence is inappropriate for the continental tectosphere and that one must consider compositional effects.

In summary, to obtain realistic estimates of correlations between data fields defined on a sphere requires that one consider the spectra of the data fields so that the number of degrees of freedom can be determined appropriately. Our analysis demonstrates that the relationship between the long-wavelength geoid and the ocean-continent function is tenuous. The large difference in correlation that we obtain with different ocean-continent functions further illustrates the insignificance of the relationship. From error estimates which account for the redness in the geoid, gravity field, and tectonic regionalization spectra, we conclude that neither the geoid nor the radial component of the gravity field has a platform and shield signal which differs significantly ($> 1.5\sigma$) from zero. Additionally, by considering regionally averaged shear-wave travel-time anomalies together with our model of the continental tectosphere's contribution to the geoid, we find that $\partial \ln \rho / \partial \ln v_s$ is about 0.035 ± 0.025, consistent at the 1.5\sigma level with Jordan's [1988] isopycnic hypothesis.

Acknowledgments. We thank P. Puster and G. Masters for computer code and T. A. Herring, P. Puster, W. L. Rodi, and M. Simons for helpful discussions. Figures 3.3 and 3.4 were created using the Generic Mapping Tools system [Wessel and Smith, 1991].
REFERENCES


Table 3.1. GEM-T3 ($l = 2-12$): Regional averages and statistical standard errors corresponding to the geoid and to perturbations to the radial component of the gravity field.

<table>
<thead>
<tr>
<th>Geoid (m)</th>
<th>$\gamma_A$</th>
<th>$\gamma_B$</th>
<th>$\gamma_C$</th>
<th>$\gamma_Q$</th>
<th>$\gamma_{PS}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Geoid (m)</td>
<td>0 $\pm$ 19</td>
<td>3 $\pm$ 10</td>
<td>29 $\pm$ 26</td>
<td>-11 $\pm$ 21</td>
<td>-13 $\pm$ 11</td>
</tr>
<tr>
<td>Gravity (mgal)</td>
<td>2 $\pm$ 11</td>
<td>0 $\pm$ 6</td>
<td>7 $\pm$ 14</td>
<td>1 $\pm$ 10</td>
<td>-9 $\pm$ 7</td>
</tr>
</tbody>
</table>

Table 3.2. Regional averages and statistical standard errors, from projections onto $A, B, C, Q, PS$), corresponding to contributions to the geoid from five model geoids - each representing a separate contribution to the geoid. The bottom two (below the double line) represent projections of TECT-1, separately, onto $(ABC, QPS)$ and $(ABCQ, PS)$.

<table>
<thead>
<tr>
<th>Geoid Contributors</th>
<th>$\gamma_A$ (m)</th>
<th>$\gamma_B$ (m)</th>
<th>$\gamma_C$ (m)</th>
<th>$\gamma_Q$ (m)</th>
<th>$\gamma_{PS}$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper 120 km</td>
<td>4.3 $\pm$ 0.4</td>
<td>-3.4 $\pm$ 0.2</td>
<td>-6.7 $\pm$ 0.4</td>
<td>4.1 $\pm$ 0.3</td>
<td>3.5 $\pm$ 0.3</td>
</tr>
<tr>
<td>Lower Mantle</td>
<td>-1 $\pm$ 11</td>
<td>8 $\pm$ 6</td>
<td>29 $\pm$ 16</td>
<td>-29 $\pm$ 12</td>
<td>0 $\pm$ 7</td>
</tr>
<tr>
<td>Slabs</td>
<td>-9 $\pm$ 8</td>
<td>-5 $\pm$ 4</td>
<td>-2 $\pm$ 10</td>
<td>16 $\pm$ 8</td>
<td>-3 $\pm$ 5</td>
</tr>
<tr>
<td>Post-Glacial Rebound</td>
<td>2 $\pm$ 1</td>
<td>1.6 $\pm$ 0.7</td>
<td>1 $\pm$ 1</td>
<td>-0.4 $\pm$ 0.9</td>
<td>-6 $\pm$ 1</td>
</tr>
<tr>
<td>TECT-1</td>
<td>2 $\pm$ 7</td>
<td>1 $\pm$ 4</td>
<td>9 $\pm$ 8</td>
<td>-3 $\pm$ 6</td>
<td>-8 $\pm$ 5</td>
</tr>
<tr>
<td>TECT-1/(ABC, QPS)</td>
<td>3 $\pm$ 2</td>
<td>3 $\pm$ 2</td>
<td>3 $\pm$ 2</td>
<td>-5 $\pm$ 4</td>
<td>-5 $\pm$ 4</td>
</tr>
<tr>
<td>TECT-1/(ABCQ, PS)</td>
<td>2 $\pm$ 1</td>
<td>2 $\pm$ 1</td>
<td>2 $\pm$ 1</td>
<td>2 $\pm$ 1</td>
<td>-8 $\pm$ 5</td>
</tr>
</tbody>
</table>
Table 3.3. Variance reductions and the corresponding confidence levels associated with the projection onto different groups of tectonic regions of five model geoids — each representing a separate contribution to the geoid. Confidence level represents the percent of random trials that yield a smaller variance reduction than that of the actual orientation of each geoid contributor.

<table>
<thead>
<tr>
<th>Geoid Contributor</th>
<th>Projection</th>
<th>Var. Red. (%)</th>
<th>Confidence (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper 120 km</td>
<td>A, B, C, Q, PS</td>
<td>87</td>
<td>100</td>
</tr>
<tr>
<td>Lower Mantle</td>
<td>A, B, C, Q, PS</td>
<td>18</td>
<td>89</td>
</tr>
<tr>
<td>Slabs</td>
<td>A, B, C, Q, PS</td>
<td>10</td>
<td>69</td>
</tr>
<tr>
<td>Post-Glacial Rebound</td>
<td>A, B, C, Q, PS</td>
<td>30</td>
<td>100</td>
</tr>
<tr>
<td>TECT-1</td>
<td>A, B, C, Q, PS</td>
<td>4.3</td>
<td>35</td>
</tr>
<tr>
<td>TECT-1</td>
<td>ABC, QPS</td>
<td>3.0</td>
<td>78</td>
</tr>
<tr>
<td>TECT-1</td>
<td>ABCQ, PS</td>
<td>2.6</td>
<td>85</td>
</tr>
</tbody>
</table>

Table 3.4. S12_WM13 [Su et al., 1994] (l = 1-12): Platform and shield averages and uncertainties corresponding to one-way S-wave travel-time anomalies (Chapter 2).

<table>
<thead>
<tr>
<th>Depth Interval (km)</th>
<th>((\Delta t/\overline{t})_p) (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>140-240</td>
<td>-2.3 ± 0.2</td>
</tr>
<tr>
<td>240-340</td>
<td>-1.6 ± 0.2</td>
</tr>
<tr>
<td>340-440</td>
<td>-1.0 ± 0.2</td>
</tr>
</tbody>
</table>
FIGURE CAPTIONS

Fig. 3.1. Non-hydrostatic geoid ($l = 2\cdots12$): Histograms of parameter values (a) $\gamma_A$, (b) $\gamma_B$, (c) $\gamma_C$, (d) $\gamma_Q$, (e) $\gamma_{PS}$ obtained from projections onto the tectonic sphere of the correlated data combined with 10,000 random orientations of the data residual sphere, characterized by $\bar{d}^{im}$ (see text). Gaussian distributions, determined by the standard deviation, mean, and area of each histogram, are superposed. (f) Histogram of variance reduction obtained from projections onto the tectonic sphere of 10,000 random orientations of the data sphere characterized by $d^{im}$. The shaded and unshaded arrows indicate the variance reductions associated with the actual orientation and the maximum variance reduction, respectively.

Fig. 3.2. Radial component of the earth's gravity field ($l = 2\cdots12$): Histogram of variance reduction resulting from 10,000 random rotations of the data sphere with respect to the tectonic sphere. The shaded and unshaded arrows indicate the variance reductions associated with the actual orientation and the maximum variance reduction, respectively.

Fig. 3.3. (a) Non-hydrostatic geoid (GEM-T3, $l = 2\cdots12$), (b) Projection of (a) onto $(A, B, C, Q, PS)$, and (c) Residual: (a) - (b). Negative contour lines are dashed and the zero contour line is thick. The contour interval is 10 m.

Fig. 3.4. (a) TECT-1 ($l = 2\cdots12$), (b) Projection of (a) onto $(A, B, C, Q, PS)$, and (c) Residual: (a) - (b). Negative contour lines are dashed and the zero contour line is thick. The contour interval is 10 m.

Fig. 3.5. (a) Histogram of correlations ($r$) between an ocean-continent function derived from GTR1 and 10,000 synthetic degree-eight fields each with the same spectral properties as the non-hydrostatic geoid. The shaded and unshaded arrows
indicate the variance reductions associated with the actual geoid and the maximum variance reduction, respectively. (b) Chi-square, calculated from the fit of Student’s $t$ distribution with the set of $t$'s calculated from $t = r\left[\frac{\nu}{\left(1 - r^2\right)}\right]^{1/2}$, plotted as a function of the number of degrees of freedom ($\nu$). The minimum value of chi-square corresponds with $\nu = 30$. Histogram of values of $t$ with Student’s $t$ distribution with $\nu$ degrees of freedom superposed: (c) $\nu = 25$, (d) $\nu = 30$, and (e) $\nu = 35$. 
Figure 3.1
Figure 3.2
Figure 3.3
Figure 3.5
CHAPTER 4

DYNAMICS OF THE CONTINENTAL TECTOSPHERE

Abstract. Data relevant to continental deep structure suggest that continental cratons overlie thick, viscous, thermal (TBL) and chemical (CBL) boundary layers, where the CBLs are intrinsically buoyant because they are depleted in basaltic constituents. If, as proposed by Jordan [1988], the continental tectosphere formed in the Archaean, then a continental tectosphere must be able to survive immersed in a convecting mantle for several billion years. In addition, since platforms and shields correlate weakly with the observed geoid (Chapter 3), acceptable models of the continental tectosphere must also satisfy this gravity constraint. We investigate the roles of different parameters in attaining long-term stability using a fully dynamic finite-element program operating within a two-dimensional Cartesian domain. We initiate our experiments with a tectosphere (CBL and TBL) immersed in a region of uniform composition, temperature, and viscosity, and consider separately the effects on the stability of the tectosphere of (1) activation energy (used to define the temperature-dependence of viscosity), (2) compositional buoyancy, and (3) linear or nonlinear rheology. The large lateral thermal gradients required to match oceanic and continental values initiate the dominant instability: a "drip" which develops at the side of the tectosphere and moves to beneath its center. High activation energies and, to a lesser extent, high background viscosity values restrict the amount of such entrainable material and the rate at which it is entrained. Compositional buoyancy does not significantly change the flow pattern. Rather, compositional buoyancy slows the destruction process somewhat and reduces the stress within the tectosphere. With a non-Newtonian rheology, this reduction in stress helps stiffen the tectosphere. In these experiments, we find that dynamical systems which ade-
quately model the present ocean-continent structures have activation energy $E^* \geq 180 \text{ KJ mole}^{-1}$ — a value about one third the estimate of activation energy for olivine, $E^* \approx 520 \text{ KJ mole}^{-1}$. Although for $E^* \approx 520 \text{ KJ mole}^{-1}$, compositional buoyancy is not required for the tectosphere to survive, the joint application of longevity and gravity constraints allows us to reject all models containing no compositional buoyancy and to predict that the ratio of compositional to thermal buoyancy within the continental tectosphere is approximately unity.

INTRODUCTION

The weak association between platforms and shields and long wavelength geoid height anomalies (Chapter 3), inferred (near) constancy of continental freeboard [e.g., Wise, 1974], and local Rayleigh number calculations [e.g., Jordan, 1975] suggest that the continental tectosphere did not form simply by conductive cooling. Jordan [1975, 1978] proposed that the thick, presumably cold continental chemical boundary layer (CBL) inferred from the analysis of seismic data [see, for example, the more recent studies by Masters et al., 1992; Grand, 1994; Su et al., 1994] are stabilized against convective disruption by compositional variations that yield neutral buoyancy (i.e., the continental tectosphere and neighboring oceanic material have the same density profile). These compositional variations have been attributed to a depletion of the denser basaltic constituents (garnet and clinopyroxene) in the continental tectosphere relative to that from the source of mid-ocean-ridge volcanism [Jordan, 1978].

Monte Carlo simulations indicate that the strong association between platforms and shields and upper-mantle shear-wave anomalies is not simply fortuitous (Chapter 2). This global relationship supports the hypothesis [Jordan, 1975] that these thick CBLs translate coherently with continental plate motions. Combining this conjecture with the measured ages of South African diamond inclusions [e.g.,
Richardson et al., 1984] and the more recent ages obtained from rhenium-osmium and other isotope systematics [Walker et al., 1989; Pearson et al., 1995] leads to the supposition that the continental tectosphere can remain intact in the convecting mantle for times in excess of a billion years [Jordan, 1988].

If the tectosphere is to survive in a convecting mantle, it must be stable both to double diffusive instabilities resulting from compositional buoyancy [e.g., Stevenson, 1979] and to tractions from the convecting mantle in which it is immersed. In this chapter, we investigate the former. In particular, we seek to determine under what circumstances compositional buoyancy might be important in allowing the tectosphere to survive. We begin by investigating the contributions of various properties of a thick, cold, CBL towards its long-term stability. In particular, we consider separately different (1) values of activation energy, (2) values of compositional buoyancy, and (3) dependence of rheology on stress. We then select those sets of parameters which both produce long-term CBL stability and create density distributions which yield geoid height anomalies consistent with those observed over platforms and shields.

STRESS REGIME

Consider the characteristic convective and buoyancy tractions associated with the continental tectosphere. The basal tractions, \( \tau_c \), associated with convection in the mantle are estimated to be in the few bar range [e.g., Hager and O'Connell, 1981]. The tractions, \( \tau_b \), associated with buoyancy forces within the tectosphere depend on the buoyancy ratio, \( B \), defined by \( B = \delta \rho / \hat{\rho} \alpha \delta T \) where \( \delta \rho \) is the difference in normative density, \( \hat{\rho} \), defined as the density at a standard temperature and pressure, \( \alpha \) is the coefficient of volume expansion, and \( \delta T \) is the difference in temperature. Here the base values are such that the initial density profile within the continental tectosphere would match that in the oceanic tectosphere for \( B \)
1; i.e., composition changes, reflected by $\delta \hat{\rho}$, would just compensate for temperature changes, $\delta T$, relative to the background temperature. Thus, with $B = 1$, $\tau_b$ vanishes. Assuming that $\hat{\rho} \approx 3500 \text{ kg m}^{-3}$, $\alpha = 3 \times 10^{-5} \text{ C}^{-1}$, $\delta T \approx 300 \text{ C}$, and the continental tectosphere is about 400 km thick, we find for $B = 0$, where composition has no effect on density that, $\tau_b \approx 1000 \text{ bars}$. For $B = 1.5$, on the other hand, there is over-compensation, with $\delta \hat{\rho}$ being larger in magnitude than is required for neutral buoyancy, and the resultant buoyancy stress is about 500 bars. (In our numerical models (shown below), however, we obtain characteristic values for $\tau_b$ which are about a factor of ten smaller than those we estimated above. This discrepancy arises because in our models, the lithostatic pressure is balanced by shear stresses, rather than normal stresses, and the characteristic thickness of our tectosphere is about ten times smaller than the width of our experimental domain.)

We address the stability of a CBL exposed to these convective and buoyancy stresses while not explicitly including all the effects of a convecting mantle. In particular, we suppress both hot plumes and small scale convection beneath oceanic regions. Modeling numerically a domain the size of the earth with a realistic Rayleigh number, and temperature and stress-dependent viscosity is computationally intensive even in two-dimensions because of the high spatial and temporal resolution required to accurately model the flow globally. However, by immersing our tectosphere in a hot (and hence reduced viscosity) isothermal and isochemical environment, we obtain basal tractions in the few bar range.

**Numerical Formulation**

We use a double-diffusive version of the finite-element program, *ConMan* [King et al., 1990], to solve numerically the advection-diffusion equations for flow of an incompressible, infinite Prandtl number fluid in a two-dimensional Cartesian
domain. With two fields affecting density: temperature \((T)\) and composition \((C)\), the relevant (dimensionless) equations are those of momentum balance

\[
\nabla \cdot (\eta \nabla \mathbf{u}) = \nabla p - Ra_T(T + BC) \hat{z}
\]

continuity

\[
\nabla \cdot \mathbf{u} = 0
\]

conservation of energy (with no internal heating)

\[
\frac{\partial T}{\partial t} = - \mathbf{u} \cdot \nabla T + \nabla^2 T
\]

and its compositional analog

\[
\frac{\partial C}{\partial t} = - \mathbf{u} \cdot \nabla C + \frac{1}{Le} \nabla^2 C
\]

where \(\eta\) is dynamic viscosity, \(\mathbf{u}\) velocity, \(p\) pressure, \(t\) time, \(Le\) Lewis number (ratio of thermal to compositional diffusivity), and \(\hat{z}\) the unit vector in the direction of increasing depth. The thermal Rayleigh number is \(Ra_T = \alpha g \rho \Delta T d^3 / \kappa T \eta_i\) where \(\alpha\) is the coefficient of volume expansion, \(g\) the acceleration due to gravity, \(\rho\) the density, \(\Delta T\) the difference between the temperature at the bottom and that at the top of the domain of depth \(d\), \(\kappa\) the thermal diffusivity, and \(\eta_i\) a reference value of dynamic viscosity (arbitrarily) defined to be the viscosity corresponding to (dimensionless) \(T = 1\), where dimensionless \(T\) varies from zero to one. Unlike thermal gradients, which evolve through the diffusion of heat over geologic time scales, compositional gradients are essentially unaffected by solid-state diffusion; hence \(Le\) is effectively infinite. Due to numerical constraints, however, we are limited to \(Le \leq 100\) [Brooks, 1981].

We define a dimensionless, temperature-dependent Newtonian viscosity, \(\eta_N\), using an Arrhenius law [e.g., King, 1990]:

\[
\eta_N(T) = \eta_i \exp \left( \frac{E^*}{T + T_{off}} - \frac{E^*}{1 + T_{off}} \right)
\]

where \(E^*\) is the activation energy, \(T_{off}\) is the offset for dimensionless surface temperature required for (dimensionless) \(T = 0\) to correspond to the \(\approx 300\) K surface
temperature of the earth. The background viscosity, $\eta_i$, implicitly accounts for pressure variations, including the effects of phase changes. A generic form for non-Newtonian viscosity is $\eta_{NN} = \eta_i (\tau_0/\tau)^{n-1}$, where $\tau$ is the second invariant of the stress tensor, $\tau_0$ is the stress value when the Newtonian and non-Newtonian viscosities are equal, and $n$, the power law exponent, is unity for Newtonian flow and three for our non-Newtonian experiments. Since strain rates are cumulative, an effective viscosity, $\eta_{eff}$, that in the limit of low stress yields Newtonian creep and in the limit of high stress yields non-Newtonian creep, can be written as $\eta_{eff} = [\eta_i^{-1} + \eta_{NN}^{-1}]^{-1}$. We use a value of $\tau_0 = 10$ bars to define the transition between Newtonian and non-Newtonian rheology and we fix the maximum dimensionless viscosity at $\eta_{cut}/\eta_i$. Our results are insensitive to $\eta_{cut}$ for $\eta_{cut} \geq 10^5$ (see Appendix B).

**Experimental Design and Parameters**

We solve our system of equations in a two-dimensional Cartesian domain using a 76 x 38 grid of square elements (see Appendix B for a discussion of the sensitivity of the results to grid resolution, aspect ratio, and symmetry). We begin all of our experiments with an "oceanic" TBL of 100 km and a "continental" TBL of 400 km thickness above a region of uniform temperature (Figure 4.1). We apply the following boundary conditions along the top ($z = 0$):

\[
\begin{align*}
    u_x &= 0 \\
    u_z &= 0 \\
    T &= 0 \ (0^\circ C) \\
    \frac{\partial C}{\partial z} &= 0
\end{align*}
\]

along the bottom ($z = 1.0 \ (760 \ km)$),
\[ \tau_x = 0 \]
\[ u_x = 0 \]
\[ \frac{\partial T}{\partial z} = 0 \]
\[ \frac{\partial C}{\partial z} = 0 \]

along the sides \((x = 0, 2.0 \text{ (1520 km)})\),

\[ \tau_x = 0 \]
\[ u_x = 0 \]
\[ \frac{\partial T}{\partial x} = 0 \]
\[ \frac{\partial C}{\partial x} = 0 \]

and beneath the oceanic TBL \((x \geq 800 \text{ km}; z = 100 \text{ km})\): \(T = 1 \text{ (1300°C)}\). See Appendix B for a discussion of the choice of boundary conditions.

Our initial temperature field (Figure 4.1) contains two adjoining TBLs of different thicknesses, one each representing an ocean and a continent. We obtain our initial temperature field by solving the conduction equation subjected to the above temperature boundary conditions and a fixed temperature, \(T = 1300°C\), along the base \((z = 400 \text{ km})\) and side \((x \leq 800 \text{ km})\) of the continental TBL. We remove these supplementary boundary conditions once the (initial) temperature field is formed.

The compositional field represents the degree of basalt depletion with respect to the oceanic average, with a higher value of \(C\) representing a larger amount of depletion and yielding a lower normative density. Following Jordan's [1988] hypothesis that, within the continental tectosphere, contours of constant composi-
tion are parallel to isotherms, we create an initial (dimensionless) composition field (Figure 4.1) that is a function of the initial (dimensionless) temperature field:

\[ C = 1 - T, \quad 0 \leq T \leq T_{CBL} \]

\[ C = 0, \quad T > T_{CBL} \]

where \( T_{CBL} = 0.9 \) \( (1170^\circ C) \) defines the temperature at the base of the CBL. It is not essential that \( T_{CBL} \) have this particular value. Somewhat larger values lead to initial (conductive) thickening of the TBL while somewhat smaller values lead to the lower part of the TBL dropping off into the fluid below (see Appendix B). We use sample buoyancy ratios of 0, 1, and 1.5, where \( B = 0 \) corresponds to density unaffected by composition and \( B = 1 \) to Jordan's [1988] isopycnic hypothesis.

As background viscosity profiles \( (\eta_b(z)) \), we use the models HGPA [Hager and Richards, 1989] and NLO [Nakada and Lambeck, 1989] (Figure 4.2). We base our selection of activation energies on Ashby and Verrall's [1977] estimate for dry olivine of \( E^* = E_{ref}^* = 522 \) KJ mole\(^{-1}\); we use \( E^* = E_{ref}^* \), \( E_{ref}^*/3 \), and \( E_{ref}^*/9 \) to explore the sensitivity of tectosphere stability to activation energy (Tables 4.1 and 4.2). Figure 4.2 shows, for each activation energy used, the corresponding initial mid-oceanic and mid-continental Newtonian viscosity profiles. In all of our numerical experiments, we use standard (constant) values for the following quantities: \( \alpha = 3 \times 10^{-5} \) \( ^\circ C^{-1} \), \( g = 9.8 \) m s\(^{-2} \), \( \rho = 3.5 \times 10^3 \) kg m\(^{-3} \), and \( \kappa_T = 10^4 \) m\(^2\) s\(^{-1}\).

**BUOYANCY CONTRAST**

One can begin to analyze the stability of a particular thermal and compositional structure by calculating the continent-ocean buoyancy contrast, i.e., by calculating at each depth the difference in density between the average mid-continental (\( \chi \leq 200 \) km) value and the laterally averaged value (Figure 4.1):

\[ \frac{\delta \rho(z)}{\rho} \equiv \alpha \Delta T \left[ \langle T + BC \rangle_{\chi \leq 200 \text{ km}} - \langle T + BC \rangle_{\chi} \right] \] (4.2)
For example, a negative value of $\delta \rho(z)/\rho$ indicates that at the depth $z$, the center of the tectosphere is lighter, on average, than the average of the ocean and continent values. Considering $\delta \rho/\rho$ as a function of depth allows one to predict whether a structure might remain near its initial configuration. As one can see from the plots of $\delta \rho(z)/\rho$ for $B = 0$ and $B = 1$ in Figure 4.1, a structure with $B = 1$ has a greater chance for survival than a structure with $B = 0$ because, with $B = 0$ the mid-continent is, on average, much denser than the surrounding material. Of course, this plot neither tells us which structure, if any, is stable nor how an unstable structure might disintegrate. In the subsequent discussion, we take 1000 My as the characteristic time at which to assess stability. By this time the CBL has either been destroyed or the fluid flow is sufficiently regular that one can reliably make predictions concerning long-term stability.

COMPOSITIONAL BUOYANCY: EFFECT ON BOUNDARY LAYER STABILITY

To illustrate the effect on boundary-layer stability of composition-induced buoyancy, we consider in detail the evolution of our model for two cases that differ only in the value of $B$. Since an acceptable model of the continental tectosphere must satisfy both the longevity and the gravity constraints, at 1000 My, we calculate the depth to the base of the CBL, $z_{CBL}$, and estimate the associated geoid height anomaly, $\delta N$ (Tables 4.1 and 4.2). In the first example, we take $E_{ref}^*/3$, $B = 0$, and $n = 3$, which leads to very rapid destruction of the continental CBL. With $B = 0$, composition is simply a tracer field — it has no effect on the motion of the fluid. At 1000 My, the CBL is essentially gone, having been washed away by the flow driven by the lateral variations in density (Figures 4.3 and 4.4). The continent-ocean buoyancy contrast, $\delta \rho(z)/\rho$, decreases in magnitude and, as the continental CBL disappears, becomes non-zero only in shallow depths.
The average geoid height anomaly associated with the beginning of the experiment is much greater in amplitude than those observed over platforms and shields (Chapter 3) (Figure 4.5a). Of course, as the source of the density contrast (the continental TBL) disappears, the average geoid height anomaly decreases accordingly. Similarly, the initial dynamic topography associated with the continental CBL is unreasonably large when compared with the earth's continental freeboard [e.g., Wise, 1974] (Figure 4.5b). The initial viscosity and stress fields (Figure 4.5c) show that the transition between Newtonian and non-Newtonian rheology occurs near the base of the tectosphere — much of the flow occurs in the Newtonian regime although a substantial amount occurs in regions of high stress (Figures 4.6a and 4.6b). At 1000 My, the tectosphere is gone, so the size of the region of high stress is significantly reduced, as is the region of high viscosity (Figure 4.5d). Subtracting the initial from the final composition (and temperature) fields illustrates which regions have gained or lost composition (and heated up or cooled down) and by how much (Figure 4.5e).

By plotting the area \( A \) of the -0.1 composition difference contour (normalized to the combined area of the initial oceanic and continental CBLs) as a function of time (Figure 4.5f), we see that most of the composition is removed from the continental CBL within the first 50 My. After this time, the rate of removal diminishes dramatically, decreasing slowly to zero at about 600 My. After the initially weak (i.e., easily deformed) material is quickly washed away, the remnant continental CBL weakens gradually as a result of the combination of the hot material flowing along its base and the strong temperature-dependence of the material's viscosity. We interpret this temporal pattern as indicating the presence of two instabilities: a mechanical mode \((t < 50 \text{ My})\) and a thermal ablation mode \((t > 50 \text{ My})\). The change in the depth to the base of the CBL, \( \Delta z_{CBL}(t) \equiv z_{CBL}(0) - z_{CBL}(t) \), expressed as a percentage of the initial CBL thickness \((z_{CBL}(0))\), varies with time in
a manner similar to $A(t)$ (Figure 4.5f). Such a correspondence is reassuring since both of these quantities were devised independently to estimate, as a function of time, the condition of the CBL.

The conductive heat flux through the continental surface (Figure 4.5g) increases by only a factor of about two (10 mW m$^{-2}$) throughout the calculation. The advective heat flux through the base of the domain is more variable. The intervals of large fluxes coincide with part of the tectosphere's TBL falling off and sinking through the base of the domain. Such occurrences are episodic. Note that the time interval between flux measurements is too crude to fully resolve the peaks in the flux-time curve.

When we repeat the above experiment with $B = 1$, the results differ significantly. The CBL remains largely intact after 1000 My, although its initially horizontal base becomes oriented at an angle of about 45° (Figures 4.7 and 4.8). The function $\delta \rho(z)/\rho$ changes shape, gradually forming a profile which indicates that a cold dense upper section of the continental CBL is being partially supported from below by a light CBL — opposite to our initial condition where the TBL extends beneath the CBL causing the small positive deviation from zero in $\delta \rho(z)/\rho$. This dense upper region is caused by the conductive cooling of the upper section of the tectosphere — with composition effectively unable to diffusive, the isopycnic condition is no longer satisfied in this region. Throughout the experiment, both the average geoid height anomaly and the dynamic topography associated with the continental CBL have geophysically reasonable values (Figures 4.9a and 4.9b).

Due to the addition of compositional buoyancy, the second invariant of the stress is reduced relative to the above experiment ($B = 0$). This lower stress yields a correspondingly higher viscosity field (Figures 4.6c, 4.7d, 4.9c, and 4.9d) which leads to a more stable CBL. The changes in the composition and temperature fields after 1000 My are confined to a much smaller area than in the previous experiment.
(Figures 4.5e and 4.9e). Again, the initially weak material is removed quickly from the tectosphere, although with the stabilizing effect of compositional buoyancy more material remains (Figure 4.9f). Further, the higher viscosity due to the lower stress than in the $B = 0$ case inhibits the thermal ablation process so that it is barely observable in $A(t)$. Here, $A(t)$ and $\Delta z_{CBL}(t)$ diverge at about 200 My. After this time, $A(t)$ is essentially constant whereas $\Delta z_{CBL}(t)$ decreases slightly indicating that the CBL is thickening. From Figure 4.8 one can see that this thickening occurs in the center of the continental CBL as a consequence of the cold "blobs" detaching from the continental TBL. The plot of heat flux through the base of the domain (Figure 4.9g) shows some of these instabilities. The heat flux through the top of the tectosphere, however, does not vary noticeably with time, holding at a value of about 20 mW m$^{-2}$.

**Bounds on $E^*$, $B$, and $\eta(z)$ from Longevity and Geoid Constraints**

To display the results of our numerical experiments and to evaluate the relative effects of (1) activation energy, (2) compositional buoyancy, and (3) rheology on tectosphere stability and the resulting geoid signal, we plot the depth to the base of the CBL and the average geoid height anomaly over the continental tectosphere, at $t = 1000$ My, for each experiment listed in Tables 4.1 and 4.2 (Figure 4.10). We do not plot results for our experiments with $E^*/9$, which were so unstable that the tectosphere was removed in $t < 1,000$ My, and the calculations terminated. In both the Newtonian and non-Newtonian experiments, using an activation energy corresponding to dry olivine of 522 KJ mole$^{-1}$ [Ashby and Verrall, 1977] assures stability — regardless of the amount (within reasonable limits) of compositional buoyancy present. However, not all of these experiments produce geoid height anomalies consistent with those observed for platforms and shields (Chapter 3). In fact, by comparing the average geoid height anomalies resulting from the $E^* = E^*_{ref}$ ex-
periments (triangles in Figure 4.10), one can see that simply satisfying the isopyc-
nic hypothesis is insufficient to ensure a geoid height anomaly consistent with plat-
forms and shields — one must also consider the dynamics of the flow. For exam-
ple, the Newtonian experiments with initial $B = 1.5$, not $B = 1$, satisfy the geoid
constraint. From the corresponding, more realistic, experiments with stress-depen-
dent rheology, we estimate that the initial $B$ is likely between about 0.9 and 1.3,
depending on the background viscosity profile (Figure 4.10b). (We refer to the
"initial" $B$ because after 1,000 My, both the temperature and composition are modi-
ﬁed and the local value of $B$ changes. For example, for the experiment shown in
Figure 4.8, at 1000 My, for $z \leq 400$ km, $\delta \rho > 0$ implies $B < 1$, while for $z \geq 400$
km, $\delta \rho < 0$ implies $B > 1$).

With an activation energy one third the above value ($E_{\text{ref}}^*/3$) and a stress-
dependent rheology, compositional buoyancy plays a major role in stabilizing the
continental tectosphere (compare the open with the shaded squares in Figure 4.10b;
compare also Figures 4.3, 4.4, 4.7, and 4.8). After 1000 My, the CBLs with $B = 1$,
are roughly 200 km thicker than those containing no compositional buoyancy.
Compositional buoyancy has a similar effect on stability for the corresponding
Newtonian experiments except for those characterized with an activation energy of
$E_{\text{ref}}^*/3$. In this case, $B = 1$ results in CBLs which are only about 50 km thicker
than the corresponding CBLs which contain no compositional buoyancy. By com-
paring the Newtonian with the non-Newtonian experiments, we see that, in general,
compositional buoyancy plays a larger stabilizing role with a non-Newtonian rheol-
ogy. As discussed in the previous section, compositional buoyancy reduces the
stress within the boundary layer, which in conjunction with the stress dependence
of viscosity, causes an increase in viscosity which stabilizes the CBL.

The two background viscosity models yield continental tectosphere thick-
nesses which are always within about 50 km of each other. In most, but not all,
cases, the model with a background viscosity described by NLO is the more stable one.

**DISCUSSION**

Our transition stress value is based on a small grain size — about one mm [Ashby and Verrall, 1977]. As grain size increases, the transition stress decreases, so our choice of $\tau_0 = 10$ bars is a conservative one in the following sense: If we had assumed a larger grain size such that $\tau_0 = 1$ bar, the fluid within and beneath the tectosphere would have been more viscous and the tectosphere would have been more likely to survive.

Conclusions drawn from studies of boundary layer dynamics depend strongly on assumptions concerning the temperature and stress dependence of viscosity and the magnitude of $\tau_c$. From an analytical analysis of the stability of a constant viscosity ($\eta = 10^{21}$ Pa s) continental tectosphere described by linear gradients in composition and temperature, Stevenson [1979] found modes of instability with characteristic growth times as short as about 200 My — at least an order of magnitude less than that required by the above age constraints. If one considers the temperature dependence of viscosity and the fact that continents are cold, a constant viscosity of $\eta = 10^{21}$ Pa s is an unrealistically low estimate. An increase of only one order of magnitude will yield characteristic time constants comparable with the age of the earth. In addition, the temperature dependence of viscosity will help stabilize the instabilities driven by the large lateral variations in temperature. As shown above, even with a high background viscosity, a strong temperature dependence is required for the tectosphere to endure.

In the high-stress regime, where $\tau_c \gg$ a few bars, Kincaid [1990] concluded that viscosity, not compositional buoyancy, is responsible for achieving long-term stability. Shapiro et al. [1991] further demonstrated that with $\tau_c \approx 60$
bars — 10-20 times that considered to be appropriate for the earth [e.g., Hager and O'Connell, 1981], a viscosity increase of a factor of about 20 between the TBL and the surrounding mantle is sufficient to maintain stability (see Appendix B), regardless of (within reasonable bounds) the amount of compositional buoyancy. These studies may overestimate the role of tractions from mantle convection because, due to numerical constraints, they were carried out at a Rayleigh number lower than appropriate for the earth. Higher Rayleigh numbers result in thinner boundary layers, and lower stresses.

In the (realistic) low-stress regime, we find that destruction is achieved in two ways: (1) through a mechanical removal of material and (2) via a thermal ablation process in conjunction with the mechanical process. The (initially) weak regions of the tectosphere are washed away quickly by convective processes; the remaining material is removed more slowly. As the base warms through conduction, it becomes weaker due to viscosity's inverse dependence on temperature. Then, in its weakened state, it is swept away by the convection currents. To estimate the viscosity required to prevent this mechanical removal, we consider the viscosity corresponding with this "ablation front". For the experiments which yield CBLs which are stable over a billion year time scale (Figure 4.10), the viscosity corresponding to the edge of the ablation front is approximately $10^{20.5} \text{ Pa s}$ (for a representative example, see Figure 4.11) indicating that a viscosity contrast of about ten between the tectosphere and the mantle is sufficient to maintain stability — similar to the result from the high-stress analysis.

Fleitout and Yuen [1984] demonstrate that the combination of pressure- and temperature-dependent viscosity can help to stabilize a thick thermal boundary layer from convective disturbance. The temperature-dependence of viscosity stabilizes the cold (shallow) part of the TBL while the pressure-dependence of viscosity stabilizes the warmer (deeper) part. Our numerical representation of the continental
tectosphere is stable even without requiring the stabilizing effects of pressure-dependent viscosity. If we were to incorporate such an effect, compositional buoyancy would presumably be an even less important stabilizing agent. Heat fluxes generated by our flow models are within a factor of two or so with those estimated by Fleitout and Yuen [1984]

Forte et al. [1995] suggested that the buoyancy profile beneath continents reverses in sign at about 250 km depth. Specifically, Forte et al. [1995] proposed that the negative buoyancy in the upper 250 km of the subcontinental mantle is partially supported by underlying lighter material. Interestingly, some of our acceptable models exhibit this same buoyancy reversal (see, for example, Figure 4.8). As we discussed above, this buoyancy reversal is caused by the conductive cooling of the tectosphere. This cooling is dependent on the convecting system in which it is placed and on the boundary conditions that we impose along the base of the domain. In fact, the existence of a buoyancy reversal is very sensitive to several of our assumptions with some of our acceptable models showing this reversal, and others not (see, for example, Figure B.21). Hence, our study can not determine whether such a reversal actually exists.

In summary, the joint application of longevity and gravity constraints allows us to evaluate the importance of specific properties of a continental tectosphere in the low-stress regime. Flow models characterized by the activation energy for dry olivine, 522 KJ mole\(^{-1}\), yield stable boundary layers, even with no compositional buoyancy present. However, activation energies, say ten-fold smaller, are much too low; they lead to a rapid (of order 10 My) destruction of the tectosphere. Even with an activation energy about 20% less than that estimated for olivine, temperature-dependent viscosity is sufficient to assure stability (Figure 4.10). With lower values of activation energy, stability can be achieved with compositional buoyancy. Compositional buoyancy plays a dual role within a thermal (and chemical) bound-
ary layer: It (1) reduces the stress within the boundary layer and (2) counteracts the thermally-induced density increase. With a stress-dependent rheology, this reduction in stress results in an increase in viscosity which, in turn, inhibits a greater region of the boundary layer from deforming. Given the weak sensitivity of our results to background viscosity (within the $10^{19}$ - $10^{20}$ Pa s range), we can not constrain this quantity further than has been done by other studies [e.g., Hager and Richards, 1989; Nakada and Lambeck, 1989].

Although not essential for the survival of the tectosphere, the geoid constraint indicates that $B \approx 1$. We infer that for the tectosphere to have formed by advective thickening, it was essential that $B$ was then about unity. Formation via continental collisions requires that the material which now constitutes the tectosphere, was, at the time of tectosphere formation, ductile enough to form the tectosphere, but subsequently on a rapid time scale became viscous enough to avoid destruction. Stress-dependent viscosity may provide such a mechanism. If the material left behind were approximately isopycnic, the high stresses associated with the detachment of the thermal lithosphere would be rapidly reduced as the thermal lithosphere sank [Houseman et al., 1981].

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Table 4.1: Newtonian rheology: Experimental parameters and resulting (i) geoid height anomalies ($\delta N$) over the continent, and (ii) depth of the base of the continental CBL ($z_{CBL}$). Both quantities are calculated at 1000 My. We keep all entries in each column to the same place for ease of reading even though not all digits are significant. We do not list the results with $E'^* = 80$ KJ mole$^{-1}$ because these calculations result in such a rapid destruction of the tectosphere that assigning a thickness is meaningless. See Figure 4.10 for a graphical representation of this table.

<table>
<thead>
<tr>
<th>$E'^*$ (KJ mole$^{-1}$)</th>
<th>$B$</th>
<th>$\eta_h(z)$</th>
<th>$\delta N$ (m)</th>
<th>$z_{CBL}$ (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>174</td>
<td>0.0</td>
<td>HGPA</td>
<td>$-37 \pm 28$</td>
<td>216</td>
</tr>
<tr>
<td></td>
<td></td>
<td>NLO</td>
<td>$-62 \pm 40$</td>
<td>250</td>
</tr>
<tr>
<td></td>
<td>1.0</td>
<td>HGPA</td>
<td>$5 \pm 3$</td>
<td>284</td>
</tr>
<tr>
<td></td>
<td></td>
<td>NLO</td>
<td>$-8 \pm 5$</td>
<td>287</td>
</tr>
<tr>
<td></td>
<td>1.5</td>
<td>HGPA</td>
<td>$14 \pm 8$</td>
<td>297</td>
</tr>
<tr>
<td></td>
<td></td>
<td>NLO</td>
<td>$21 \pm 10$</td>
<td>302</td>
</tr>
<tr>
<td>522</td>
<td>0.0</td>
<td>HGPA</td>
<td>$-131 \pm 86$</td>
<td>317</td>
</tr>
<tr>
<td></td>
<td></td>
<td>NLO</td>
<td>$-119 \pm 70$</td>
<td>309</td>
</tr>
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<td></td>
<td>1.0</td>
<td>HGPA</td>
<td>$-53 \pm 34$</td>
<td>346</td>
</tr>
<tr>
<td></td>
<td></td>
<td>NLO</td>
<td>$-42 \pm 20$</td>
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</tr>
<tr>
<td></td>
<td>1.5</td>
<td>HGPA</td>
<td>$-13 \pm 8$</td>
<td>346</td>
</tr>
<tr>
<td></td>
<td></td>
<td>NLO</td>
<td>$-3 \pm 5$</td>
<td>346</td>
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Table 4.2: Non-Newtonian rheology: Experimental parameters and resulting (i) geoid height anomalies ($\delta N$) over the continent, and (ii) depth of the base of the continental CBL ($z_{CBL}$). Both quantities are calculated at 1000 My. We keep all entries in each column to the same place for ease of reading even though not all digits are significant. We do not list the results with $E^* = 80$ KJ model$^{-1}$ because these calculations result in such a rapid destruction of the tectosphere that assigning a thickness is meaningless. See Figure 4.10 for a graphical representation of this table.

<table>
<thead>
<tr>
<th>$E^*$ (KJ mole$^{-1}$)</th>
<th>$B$</th>
<th>$\eta_b(z)$</th>
<th>$\delta N$ (m)</th>
<th>$z_{CBL}$ (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>174</td>
<td>0.0</td>
<td>HGPA</td>
<td>-10 ± 5</td>
<td>92</td>
</tr>
<tr>
<td></td>
<td></td>
<td>NLO</td>
<td>-17 ± 11</td>
<td>138</td>
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<tr>
<td></td>
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<td></td>
<td></td>
<td>NLO</td>
<td>-2 ± 5</td>
<td>320</td>
</tr>
<tr>
<td></td>
<td>1.5</td>
<td>HGPA</td>
<td>-2 ± 1</td>
<td>289</td>
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<td></td>
<td></td>
<td>NLO</td>
<td>10 ± 6</td>
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</tr>
<tr>
<td>522</td>
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<td>HGPA</td>
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<td>HGPA</td>
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<td></td>
<td></td>
<td>NLO</td>
<td>-19 ± 11</td>
<td>348</td>
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<tr>
<td></td>
<td>1.5</td>
<td>HGPA</td>
<td>21 ± 10</td>
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<td></td>
<td></td>
<td>NLO</td>
<td>14 ± 8</td>
<td>348</td>
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</table>
FIGURE CAPTIONS

Fig. 4.1. Contours of initial composition (left) and temperature (mirrored - right). Composition is contoured in increments of 0.1, with $C = 1$ at the top and $C = 0.1$ at the base of the CBL. Temperature is contoured in increments of 0.1, with $T = 0$ at the top and $T = 1$ at the base of the TBL. The $C = T = 0.1$ contours are thick. Fluctuations in the initial temperature field are due to the superposition on the temperature field of a white noise perturbation with a zero mean and a peak-to-peak amplitude of 0.01. Composition is determined from the unperturbed temperature field according to equation 4.1. The temperature at the base of the CBL, $T_{CBL}$, is 0.9 ($1170$ °C). The center frame displays, as a function of depth, the difference between the average mid-continental ($x \leq 200$ km) density and the average lateral density (see equation 4.2) for $B = 0$ (dashed line), 1 (thin line), and 1.5 (thick line).

Fig. 4.2. Newtonian viscosity profiles based on background viscosities (a) NLO (thick line) and (b) HGPA (thick line), and evaluated using the initial temperature field of the mid-continent and three activation energies: $E^*_{ref}$ (thin lines), $E^*_{ref}/3$ (dashed lines), and $E^*_{ref}/9$ (dash-dotted lines), where $E^*_{ref} = 522$ KJ mole$^{-1}$. The dotted lines represent the viscosity profile corresponding to the initial mid-oceanic temperature profile. Note: we do not model NLO's 100-fold increase in viscosity at 670 km depth — its effect on our experiments is insignificant (see Appendix B).

Fig. 4.3. Parameters: $E^* = E^*_{ref}/3$, $B = 0$, $n = 3$, $\eta_b(z) = NLO$. Four equitemporal frames: $t = (a) 0$, (b) 50, (c) 100, and (d) 150 My. Each frame contains (left) contours of composition (purely a tracer field having no effect on the dynamics of the fluid) with superposed velocity arrows, (center) the difference, as a function of depth, between the average mid-continental ($x \leq 200$ km) density and the average
lateral density (see equation 4.2), and (mirrored - right) contours of temperature with superposed velocity arrows. Contour levels as in Figure 4.1.

Fig. 4.4. Parameters as in Figure 4.3. Four equitemporal frames: $i = (a) 250$, $(b) 500$, (c) 750, and (d) 1000 My. Frames as described in Figure 4.3.

Fig. 4.5. Parameters as in Figure 4.3. (a) Geoid height anomalies ($\delta N$) at $t = 0$ (dotted line), 10 (dashed line), 100 (thin solid line), and 1000 (thick solid line) My. There is little difference between $\delta N$ at $t = 0$ and 10 My so the dotted and dashed lines appear superposed. (b) Dynamic topography ($h$) at $t = 0$ (dotted line), 10 (dashed line), 100 (thin solid line), and 1000 (thick solid line) My. There is little difference between $h$ at $t = 0$ and 10 My so the dotted and dashed lines appear superposed. (c) Viscosity field ($\eta$) with superposed velocity arrows (left) and second invariant of the stress tensor ($\tau(II)$) (mirrored - right) at $t = 0$. Viscosity contours are spaced by factors of 100, with the thick line representing the lowest contour level ($10^{20}$ Pa s). For the stress field, the thick line represents the lowest contour level (five bars) and each succeeding contour indicates a stress value a factor of two larger than that for the immediately preceding contour. (d) Viscosity field ($\eta$) with superposed velocity arrows (left) and second invariant of the stress tensor ($\tau(II)$) (mirrored - right) at $t = 1000$ My. Contour intervals as in (c). (e) Initial composition field subtracted from the composition field at $t = 1000$ My (left) and initial temperature field subtracted from the temperature field at $t = 1000$ My (mirrored - right). Dimensionless contours are spaced in increments of 0.1 with dashed lines representing a loss of composition/temperature and thick solid lines representing a gain. The zero contours are shown with thin solid lines. (f) Area ($A$) of the -0.1 difference contour, normalized by the area of the initial oceanic and continental CBL, representing a loss of composition (thin line, open circles), and the change ($\Delta z_{CBL}$) in the depth of the base of the continental CBL ($C = 0.1$) at 1000 My ex-
pressed as a percentage of the initial depth (thick line, asterisks). The depth of the base of the continental CBL is estimated from the median depth of the $C = 0.1$ contour within the continental CBL. (g) Conductive heat flux ($Q$) through the surface of the continent (thin line, open circles) and advective heat flux through the base of the domain (thick line, asterisks).

Fig. 4.6. (a) Parameters: $E' = E_{ref}^*/3$, $B = 0$, $n = 3$, $\eta_b(z) = \text{NLO}$ (Figure 4.3). Viscosity field ($\eta$) with superposed velocity arrows (left) and second invariant of the stress tensor ($\tau(II)$) (mirrored - right) at $t = 0$. Velocity scaling is clipped at $v_{cut} = 1$ cm/yr to exhibit lower values more clearly. Contour intervals as in (4.5c). (b) Viscosity field ($\eta$) with superposed velocity arrows (left) and second invariant of the stress tensor ($\tau(II)$) (mirrored - right) at $t = 1000$ My. Contour intervals as in (4.5c), Velocity scaling as in (a). (c) Parameters: $E' = E_{ref}^*/3$, $B = 1.0$, $n = 3$, $\eta_b(z) = \text{NLO}$ (Figure 4.7). Viscosity field ($\eta$) with superposed velocity arrows (left) and second invariant of the stress tensor ($\tau(II)$) (mirrored - right) at $t = 0$. Velocity scaling as in (a). Contour intervals as in (4.5c). (b) Viscosity field ($\eta$) with superposed velocity arrows (left) and second invariant of the stress tensor ($\tau(II)$) (mirrored - right) at $t = 1000$ My. Contour intervals as in Figure 4.5c, Velocity scaling as in (a).

Fig. 4.7. Parameters: $E' = E_{ref}^*/3$, $B = 1.0$, $n = 3$, $\eta_b(z) = \text{NLO}$. Four equitemporal frames: $t = (a) 0$, (b) 50, (c) 100, and (d) 150 My. Frames as described in Figure 4.3.

Fig. 4.8. Parameters as in Figure 4.7. Four equitemporal frames: $t = (a) 250$, (b) 500, (c) 750, and (d) 1000 My. Frames as described in Figure 4.3.

Fig. 4.9. Parameters as in Figure 4.7. Frames as described in Figure 4.5.
Fig. 4.10. The depth to the base of the CBL ($z_{CBL}$) (y-axis) and the mean geoid height anomaly ($\delta N$) over the continental CBL (x-axis) both at 1000 My. The depth to the base of the CBL is estimated from the median depth to the $C = 0.1$ contour within the continental CBL. Triangles and squares indicate experiments with activation energies of $E^* = E_{ref}^*$ and $E_{ref}^*/3$, respectively. White, gray, and black symbols indicate experiments with buoyancy ratios of $B = 0, 1, 1.5$, respectively. Small and large symbols indicate experiments with background viscosity profiles based on HGPA and NLO, respectively. The ordinates of the horizontal dashed lines indicate a (somewhat arbitrary) upper bound of $z_{CBL}$ based on the inferred thickness of the continental tectosphere (Chapter 2). The abscissas associated with the vertical dashed lines correspond to $N_{ref} \pm 2\sigma$ as determined from TECT-1 (Chapter 3). (a) Newtonian rheology. (b) Non-Newtonian rheology. Circles represent experiments with a background viscosity profile determined by NLO and an activation energy of $E^* = E_{ref}^*/1.25$. White and gray symbols indicate experiments with buoyancy ratios of $B = 0$ and $1$, respectively.

Fig. 4.11. Parameters as in Figure 4.7. Viscosity field ($\eta$) (solid lines) with velocity arrows and the change in the composition field (see Figure 4.9) (dashed lines) superposed; $t = 1000$ My. Velocity scaling is clipped at $v_{cut} = 1$ cm/yr to exhibit lower values more clearly. Contour intervals as in Figure 4.5.
Figure 4.1
Figure 4.2
Figure 4.3
Composition

\[ t = 250 \text{ My}; \ V_{\text{max}} = 11 \text{ cm/yr} \]

\[ t = 500 \text{ My}; \ V_{\text{max}} = 9 \text{ cm/yr} \]

\[ t = 750 \text{ My}; \ V_{\text{max}} = 13 \text{ cm/yr} \]

\[ t = 1000 \text{ My}; \ V_{\text{max}} = 10 \text{ cm/yr} \]

Temperature

\[ Q_{\text{sect}} = 14; \ Q_{\text{base}} = 7 \text{ (mW/m}^2\text{)} \]

\[ Q_{\text{sect}} = 18; \ Q_{\text{base}} = 5 \text{ (mW/m}^2\text{)} \]

\[ Q_{\text{sect}} = 20; \ Q_{\text{base}} = 43 \text{ (mW/m}^2\text{)} \]

\[ Q_{\text{sect}} = 20; \ Q_{\text{base}} = 9 \text{ (mW/m}^2\text{)} \]

Figure 4.4
Figure 4.5
Figure 4.6
**Figure 4.7**

Composition and Temperature diagrams for different times and maximum velocities.

- **t = 0 My; Vmax = 5 cm/yr**
- **t = 50 My; Vmax = 7 cm/yr**
- **t = 100 My; Vmax = 3 cm/yr**
- **t = 150 My; Vmax = 4 cm/yr**
Figure 4.8
Figure 4.9
Figure 4.10
Figure 4.11

$t = 1000$ My; $V_{clip} = 1$ cm/yr
CHAPTER 5

CONCLUSIONS

The conclusions drawn from this thesis can be divided into two groups: statistical and geophysical. These two categories go hand in hand. In particular, robust error estimation is essential for many fields of geophysical inference. Underestimating errors, e.g., by assuming an inappropriate error spectrum, may lead to interpretations of differences that are not real. But overestimating errors can lead to the mistake of concluding that real differences are insignificant. We developed one approach to error estimation and used it to study the relationships between surface tectonics and, separately, seismic tomographic models and gravity data. To obtain realistic estimates of the significance of correlations between such data fields requires that one consider their spectra. Attaching significance to such correlations also requires that one know the number of degrees of freedom. Our technique, based on Monte Carlo simulations, allows us to both incorporate explicitly the spectral properties of the data, and when evaluating the correlations between two data spheres, estimate the relevant number of degrees of freedom.

As we discussed in Chapter 1, our goal was to try to resolve, through quantitative treatment, some of the controversies which have characterized the development of models of the structure and dynamics of the continental tectosphere. In this chapter, we first compare our statistical results and dynamical models with results from other studies. We next address the applicability of the models outlined in Chapter 1 in the context of the results that we obtained in this thesis. Finally, we present a model of the continental tectosphere constructed to conform with the results from this and other studies.
The results from our seismic analysis (Chapter 2) are generally consistent with those of Lerner-Lam and Jordan [1987], though their analysis was based on a degree-eight tomographic model. Through our statistical treatment, we evaluated our ability to distinguish tectonic provinces in the shear-wave travel-time anomalies as a function of depth. Most importantly, our estimates of parameter uncertainties allowed us to assess the significance of the differences in regional averages. When we repeated Lerner-Lam and Jordan's [1987] study using the 40-400 km depth interval and applied our statistical machinery, we found that the six regions of GTR1 are not all distinguishable. Using a 1% F-test, we concluded that platforms are not resolvable from shields, and old ocean basins are not distinguishable statistically from orogenic belts.

Our conclusion that the seismic signature associated with platforms and shields extends to at least 400 km depth is inconsistent with Polet and Anderson's [1995] conclusion. We used (1) different statistical techniques, (2) different tomographic models with different resolutions and coverage, and (3) different regionalizations from the ones that they used. Although the information given by Polet and Anderson [1995] is insufficient to allow us to pinpoint the contributions of each of these differences to the overall disparity in results, we can make some general comments. First, the error bars accompanying their average velocity anomalies, and hence their conclusions, appear to be the standard deviation of the individual velocity anomalies associated with each tectonic region and depth interval, rather than the (appropriate) standard deviation of the mean. Second, as discussed in Chapter 2, Zhang and Tanimoto's [1992, 1993] model which Polet and Anderson [1995] used is inconsistent with observations of SS and SS-S travel-time residuals [Su et al., 1992]. Further, Grand's [1994] model, which Polet and Anderson [1995] also used, has a limited geographic coverage, primarily centered on the Atlantic hemisphere, and averages are biased due to this incomplete coverage.
Because Grand’s [1994] model does not include some major cratonic regions (e.g., Australia and Antarctica), the regional averages are weighted more heavily by those locations that are included; moreover, regions associated with the periphery of the model’s coverage, and also the model’s sensitivity to these regions (e.g., Western Africa), could introduce further distortions. Within the 240-340 km depth interval, oceans are seismically "slow" and continents are "fast" (Table 2.1); but, relative to these global regional averages, the Atlantic is anomalously fast and Africa is anomalously slow (Figure 2.6c). Hence the dichotomy between oceanic and continental shear-wave velocities is diluted in Grand’s [1994] model by these anomalous signals and extrapolations to global regional averages are misleading.

Our statistical analysis of the relationship between surface tectonics and the long-wavelength geoid suggests that the association between these two quantities is, at best, tenuous. Specifically, we find no regional average with a significance greater than about 1.6σ. This conclusion differs substantially from the highly significant correlation between an ocean-continent function and the long-wavelength geoid (l = 2-8) reported by Forte et al. [1995]. However, we find this stated significance to be suspect for several reasons. For example, using ocean-continent functions derived from GTR1 [Jordan, 1981a] and from Mauk’s [1977] regionalization, we obtain correlation coefficients which differ from each other by more than 40%, thus demonstrating the sensitivity of the result to the regionalization used. Although the significance levels (see Chapter 3) of the correlations associated with these two correlations are, respectively, about 85% and 95%, they are dominated by the degree-two, order-zero term in the geoid. For example, if one considers only degrees l = 3-8, the significance levels of these correlations reduce to about 55% and 60%, respectively, and are hence insignificant.

Our conclusion also differs substantially from that of Souriau and Souriau [1983] who, also using a Monte Carlo scheme, found that the non-hydrostatic
geoid ($l = 3-12$) correlates significantly (at the 95% confidence level) with the surface tectonics defined by Okal [1977]. The close geoid-tectonic association presented by Souriau and Souriau [1983] is partially related to the fact that Okal's [1977] regionalization includes subduction zones; the association between the geoid and this regionalization is a result of the strong geoid-slab correlation [e.g., Hager, 1984]. When we substitute a slab-residual model geoid (Hager and Clayton, 1989) for the non-hydrostatic geoid, we find that the significance reduces to about 50%, demonstrating that the signal observed by Souriau and Souriau [1983] is largely due to the effect of slabs on the geoid.

In Chapter 4, we illustrated the mechanisms relevant for stability in our dynamical flow models, which make predictions about temperature, heat flow, buoyancy, viscosity, stress, dynamic topography, and geoid height anomalies. We discuss some of these predictions here. The thermal structure of our model continental tectosphere matches the oceanic adiabat and surface temperature but is somewhat colder than an earth-like thermal boundary layer (TBL) which contains internal heating [e.g., Pollack and Chapman, 1977]. Our models produce heat flows of about 20 mW/m² through the surface of the continental tectosphere, consistent with reduced heat flow estimates for continents of 15-20 mW/m² [e.g., Sclater et al., 1980, 1981].

Our viscosity structure is compatible with inferences from postglacial rebound studies which suggest that the average upper mantle viscosity associated with continents is of order $10^{21}$ Pa s [e.g., Peltier and Tushingham, 1989]. From an estimate of the Maxwell time, we conclude that material with a viscosity greater than about $10^{22}$ Pa s is effectively elastic and that material with a lower viscosity can be treated as a viscous fluid. This demarcation implies that our elastic lid has a thickness of approximately 300 km (see, for example, Figure 4.8d). Postglacial rebound analyses are sensitive to the effect of lid thickness at the edges of ice loads.
at about spherical-harmonic degree 30 [e.g., Hager, 1991]. The dominant signal from the center of the ice sheet at Hudson Bay, for example, is at spherical-harmonic degree 10 [Simons, 1995]. Since for flexural problems, the wavenumber at which elastic support becomes important scales as the plate thickness to the $-3/4$ power, an elastic plate thickness of 300 km would be associated with a spherical-harmonic degree of about 14, which is greater than 10. Thus, one can neither confirm nor refute the existence of an elastic lid of this thickness from the analysis of postglacial rebound data from regions near the center of the shield. On longer relaxation time scales, however, where the elastic lid thickness should coincide with the 300-600°C temperature contour, our results suggest elastic plate thicknesses of between about 100-200 km, consistent with estimates from other studies of continental elastic plate thicknesses [e.g., Bechtel et al., 1990].

Modeling numerically a domain the size of the earth with a realistic Rayleigh number, temperature, stress, and depth-dependent viscosity is computationally intensive even in two-dimensions because of the high spatial and temporal resolution required to accurately model the flow globally. However, by immersing our tectosphere in a hot (and hence reduced viscosity) isothermal and isochemical environment, the flow that our model generates produces stresses which are consistent with those which have been estimated for the base of the continental tectosphere [e.g., Hager and O'Connell, 1981]. But even our experiments which generated shear tractions an order of magnitude larger, yielded a very similar conclusion. Specifically, in both the realistic (low stress) and high stress regimes, a viscosity contrast between the convecting mantle and the tectosphere of order 10 is sufficient to stabilize the tectosphere from convective disruption.

In all of our experiments, the tectosphere is continuously attacked from the sides by flow driven by the large thermal gradients associated with the ocean-continent transition, rather than intermittently by plumes from below [e.g., Zhong and
The size of the region of the tectosphere which deforms is restricted by (1) the strong temperature dependence of viscosity, and, to a lesser extent, (2) the high background viscosity values. We find that destruction is achieved in two ways: (1) through a mechanical removal of material and (2) via a thermal ablation process which acts in conjunction with the mechanical process. The (initially) weak regions of the tectosphere are washed away quickly by convective processes; the remaining material is removed more slowly. As the base warms through conduction, it becomes weaker due to viscosity's inverse dependence on temperature. Then, in its weakened state, it is swept away by the convection currents. For the experiments which yield chemical boundary layers which are stable over a billion year time scale, the viscosity corresponding to the edge of the ablation front is approximately $10^{20.5}$ Pa s, indicating that a viscosity contrast of about ten between the tectosphere and the mantle is sufficient to maintain stability — similar to results from high-stress analyses [Kincaid, 1990; Shapiro et al., 1991] which explicitly included basal plume attacks. Our estimates of heat flow through the surface of the continental tectosphere do not vary much with time. We speculate that this constancy of heat flow indicates that the temperature-dependence of viscosity acts as an insulator by slowing the thermal ablation process. Hence the temperature-dependence of viscosity is likely responsible for the heat flux through the base of the continental tectosphere probably being no more than a few mW/m$^2$ [Jordan, 1988]. Internal heating would lead to higher temperatures within the tectosphere, and the viscosity would therefore be reduced. One might expect that the tectosphere would then be less stable. But our models indicate that, with the expected activation energy of 520 KJ mole$^{-1}$, only the part of the mantle with $T \geq 1250^\circ$C participates in the flow. So long as internal heating does not raise the temperature above this amount, it should have little effect on the survival of the tectosphere. In addition, pressure-dependent viscosity would help stabilize the deeper part of the tecto-
sphere by increasing the viscosity in the region at the base of the tectosphere [Fleitout and Yuen, 1984].

Our numerical models generally show rising material under the oceans and downward flow beneath the tectosphere. Except when pieces of the TBL detach and sink through the base of our experimental domain, the heat flux through the base is about an order of magnitude less than the mean global heat flux generated by mantle convection [e.g., Hager and O'Connell, 1981]. Hence if we were to place our tectosphere into a fully convecting system, with realistic basal tractions, the flow might well be different from the somewhat regular pattern found in our numerical models.

While our results indicate that the seismic signal associated with platforms and shields extends to at least 400 km depth, Anderson [1989] argued that seismically fast continental mantle within the 200-400 km depth interval represents the remnants of cold oceanic lithosphere, not tectosphere. But, Zhong and Gurnis [1993] demonstrated that although continents do spend much of the time over cold regions of the mantle, they repeatedly move over upwellings. Given the globally strong relationship between surface tectonics and upper-mantle shear-wave heterogeneity revealed in Chapter 2, we contend that it is highly unlikely for all continents to currently be overlaying the remnants of cold oceanic lithosphere. Taking this argument one step further, we suggest that this heterogeneity represents the tectosphere. As Jordan [1988] noted, the majority of the tectosphere could not have attached itself to a pre-existing (thinner) version by an underplating mechanism. In this view, therefore, the tectosphere is as old as the shallower regions, which have been dated at more than one billion years old [e.g., Richardson, 1984; Walker et al., 1989; Pearson et al., 1995].

Forte et al. [1995] suggested that the buoyancy profile beneath continents reverses in sign at about 250 km depth. Specifically, they proposed that the nega-
tive buoyancy in the upper 250 km of the subcontinental mantle is supported by underlying lighter material. Interestingly, some of our acceptable models exhibit this same buoyancy reversal (see, for example, Figure 4.8). On the other hand, some of our acceptable models also yield very different buoyancy profiles (see, for example, Figure B.21). Hence, our study can not determine whether such a reversal actually exists. From our geoid analysis we can state that an average continental geoid height anomaly of -12 m, which forms the basis of Forte et al's [1995] model, is not required by the data, although it is, of course, permitted; the main point here is that the magnitude of this average is not large compared with its standard error.

Our results support some, but not all, of the components of Jordan's [1975, 1978, 1981b, 1988] tectosphere hypothesis. For example, our estimate of $\partial \ln \rho / \partial \ln \nu$, for the continental tectosphere from our model geoid, TECT-1, is consistent at the 1.5$\sigma$ level with the isopycnic hypothesis [Jordan, 1988], for which $\partial \ln \rho / \partial \ln \nu = 0$ and inconsistent at the 9$\sigma$ level with a purely thermal (chemically uncompensated) continental tectosphere (rejecting Chapman and Pollack's [1977] model of indefinite thermal boundary layer growth). However, in our model, there is only a rather narrow range in parameter space where compositional buoyancy would be necessary for the continental tectosphere, once formed, to survive. For stability over earth-age time scales, the temperature dependence of viscosity is the essential ingredient — without it, no amount of compositional buoyancy will prevent the continental tectosphere from being destroyed by mantle convection processes. The activation energy for dry olivine of 522 KJ mole$^{-1}$ [Ashby and Verrall, 1977] is sufficient, even with no compositional buoyancy and a background viscosity of only $2 \times 10^{19}$ Pa s, to stabilize a continental tectosphere for at least a billion years. But, with only a weak temperature- (or composition-) dependent viscosity, a continental tectosphere which formed in the Archaean, would not be present today,
regardless of the amount of compositional buoyancy. Only if the appropriate activation energy were roughly one third that estimated for dry olivine would compositional buoyancy play a major role in achieving long-term tectosphere stability. The primary effect of compositional buoyancy in our model cannot be viewed simply in terms of reducing the effective Rayleigh number to a value less than the critical Rayleigh number. Rather, its presence reduces the stress within the chemical boundary layer which, in conjunction with a stress-dependent rheology, increases the boundary layer viscosity and prevents further boundary layer destruction. But, since we conducted our analysis in a two-dimensional rather than a three-dimensional domain, we can not address the role of, for example, double-diffusive instabilities which could develop with the addition of a third dimension. However, the modes of instability and the mechanisms which participate in the stabilization process are dominated by differences in temperature between oceans and continents, implying that differences within oceans and within continents in the third dimensions might have only "second-order" effects on the dynamics.

We envision, therefore, a continental tectosphere which is (1) seismically faster than oceanic mantle to about 400 km depth, (2) depleted in basaltic constituents (i.e., Al₂O₃, FeO, CaO) such that it is approximately neutrally buoyant relative to the average composition of the upper mantle, (3) a few billion years old having been formed in the Archaean by advective thickening driven by continental collisions, (4) stabilized against convective disruption by the strong temperature-dependence of viscosity estimated for olivine, and (5) shielded from the heat transport associated with the underlying mantle convection by a slowing of the thermal ablation process inherent in the tectosphere's temperature dependence of viscosity. The approximately neutral buoyancy observed in the geoid for the continental tectosphere is not essential for the survival of the tectosphere, but may indicate a previous dynamical importance. We infer that for the tectosphere to have formed by ad-
vective thickening, it was essential that the ratio of compositional to thermal buoyancy was then about one. Formation via continental collisions requires that the material which now constitutes the tectosphere, was, at the time of tectosphere formation, ductile enough to form the tectosphere, but subsequently became viscous enough on a rapid time scale to avoid destruction. Stress-dependent viscosity may provide such a mechanism. If the material left behind were approximately isopycnic, the high stresses associated with the detachment of the bottom part of the thermal lithosphere would be rapidly reduced as the detached part sank [Houseman et al., 1981].
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APPENDIX A

SUPPLEMENT TO CHAPTERS 2 AND 3

SELECTION OF TECTONIC REGIONALIZATION

There are several global tectonic regionalizations available. They differ in both resolution and classification. GTR1 [Jordan, 1981] (Table A.1, Figure A.1) and regionalizations published by Okal [1977] (Table A.2) and Mauk [1977] (Table A.3) contain six, seven, and 20 regions, respectively. Both GTR1 and Mauk's [1977] regionalizations are defined on the earth's surface on a grid with 5° x 5° cells, whereas Okal's [1977] model is defined using 15° x 15° and 10° x 15° cells. Mauk's [1977] regionalization allows for as many as 10 regions to be represented in any given cell while the others are defined by a one-cell, one-region mapping. We do not use Okal's [1977] regionalization because, even below the 15° x 15° resolution level, many locations are clearly misclassified. For example, Okal [1977] labels the entire continent of Antarctica a shield whereas a significant fraction (≈ 1/3) is orogenic in nature. Okal [1977] also categorizes some islands (e.g., Iceland and Great Britain) as shields. Mauk's [1977] regionalization has many small components which cannot be distinguished with a degree-12 tomographic model; 11 of its 20 regions encompass a total of only 19.1% of the earth's surface — an average of less than 1.8% for each of these regions. Since we cannot even distinguish, within the 140-440 km depth interval, the seismic signature associated with shields from that of platforms, using a separate Archaean shield region from Mauk's [1977] regionalization would not provide us with any significant additional information. For example, were all of the Archaean shields contiguous, one would still not be able to adequately resolve this small area (2% of the earth's surface) in a
degree-12 expansion. Moreover, even with GTR1, the seismic signatures associated with platforms and shields are virtually indistinguishable. The spherical-harmonic expansion of platforms, for example, is very highly correlated with that of shields, with the exception of degree two (Figure A.2). Since the spherical-harmonic representations of the travel-time anomalies have distinctly red spectra, the long-wavelength correlations are the more relevant ones. To illustrate this point, we use the projection associated with the 340-440 km depth interval as a representative example. For GTR1, we use the regionalization \((ABC, Q, PS)\) which we select using a 1% F-Test criterion (see the next section). We construct a similar regionalization using Mauk's [1977] regionalization by combining regions \([1-7], [8-14, 16-17], [15,18-20]\). From Table A.4, one can see that the two regionalizations yield equivalent regional averages and both projections provide a variance reduction of about 43%. Further, when we subdivide the composite region \((15,18-20)\) into two pieces: \((15,18, [19,20])\) and \((15, 18, 19, 20)\) we also obtain variance reductions of 43%, indicating no significant improvement in the fit to the data.

With reliable higher-degree tomographic models, it may be possible to make finer distinctions. For example, the North American craton contains a large concentration of Archaean rocks and is associated with a negative travel-time anomaly (see, for example, Figure 2.5a), whereas the eastern African craton is associated with a large positive travel-time anomaly. A finer classification, at a resolution comparable, say, to Mauk's [1977] regionalization, would be able to discern such differences, if they exist, on a global basis.

\textbf{F-TEST}

In our investigation of the relationship between upper-mantle shear-wave heterogeneity and surface tectonics (Chapter 2), we project one-way shear-wave travel-time anomalies associated with 100 km depth intervals onto each candidate
regionalization from the set of 243 possible combinations of the six regions of GTR1. From the resulting set of projections which yield values for the parameters \( \tau_A, \tau_B, \ldots \), we select separately from each group of \( n \)-component \((2 \leq n \leq 6)\) regionalizations, the one which is associated with the greatest reduction in variance (see Chapter 2 for the definition of variance reduction). We apply the F-test to determine whether increasing by one the number of parameters in a model (e.g., \( PS \rightarrow P, S \)) significantly improves the model's fit to the data. In particular, we apply the F-test to these five chosen regionalizations to determine which one represents the resolution limit of our inversion. We use the standard definition [e.g., Bevington, 1969]:

\[
F = (N - m - 1) \frac{\chi^2(m - 1) - \chi^2(m)}{\chi^2(m - 1)}
\]

where \( N \) is the number of data and \( m \) is the number of independent parameters. For our problem, the number of independent parameter equals the number of regions minus one (i.e., \( m = 1 \rightarrow 5 \)). With data represented as spherical-harmonic coefficients, one can calculate the number of data from \((l_\text{max} + 1)^2 - l_\text{min}^2\) where \( l_\text{min} \) and \( l_\text{max} \) are, respectively, the minimum and maximum values of angular degree. Because we considered spherical-harmonic degrees \( l = 1 \text{–} 12 \) \((N = 168)\), values of \( F \) greater than about 6.8 indicate that the probability is less than 1% that the improvement in the model's fit to the data is due to chance (Table A.5). Table A.6 shows, for sample projections, the values of \( F \) for the regionalizations which were selected from the variance reduction criteria.

**S12_WM13: ERROR PROPAGATION**

Su et al. [1994] provide only the diagonal elements of their postfit covariance matrix; Su [personal communication, 1995] stated that the off-diagonal components of the covariance matrix would have provided little information about the model resolution. With the variances corresponding to each spherical-harmonic co-
efficient and Chebyshev polynomial pair, we could determine analytically the standard errors in the spherical-harmonic coefficients representing the travel-time anomalies. But, instead, we use a Monte Carlo approach because of its technical simplicity. We perturb each of the tomographic model's spherical-harmonic coefficients with noise obeying a Gaussian distribution whose full width at half maximum is determined by the standard error in that coefficient. From each of 10,000 sets of perturbed coefficients, we calculate spherical-harmonic coefficients representing the surface expression of the travel-time anomalies within specified depth intervals. For the 10,000 sets of coefficients representing each depth interval, we compute the standard deviation of each sample coefficient and take these values as the standard deviations associated with the travel-time anomalies. We choose 10,000 trials because we then expect errors of only about 1% due to statistical noise from the trials.

PARAMETER UNCERTAINTIES

As discussed in Chapters 2 and 3, due to the redness of the spectra of the regionalization (GTR1) and the data (respectively, shear-wave travel-time anomalies from S12_WM13 and the observed geoid, GEM-T3 [Lerch et al., 1994], referred to the earth in hydrostatic equilibrium [Nakiboglu, 1982]), we determine the standard errors in the parameter estimates from the distributions of parameter values generated from Monte Carlo simulations. For example, in Chapter 3, we select 10,000 random sets of Euler-angle triples (from a parent distribution in which all pole orientations are equally probable) and correspondingly rotate the data-residual sphere with respect to the tectonic sphere. We make the assumption here that the data-residual sphere contains all of the noise and nothing but the noise. We then add the correlated signal to the rotated data-residual sphere and project the composite onto the tectonic regionalization. The resulting histograms of parameter values
(e.g., $\gamma_A, \gamma_\beta, \ldots$) will approximate Gaussian distributions centered on the parameter value corresponding to the actual orientation of the data-residual sphere. We take the parameter values corresponding to the actual orientation of the data-residual sphere as our parameter estimates and the standard deviations of these Gaussian distributions as the parameter uncertainties. Alternatively, we can assign random (white noise) values to each coefficient describing the data-residual sphere while constraining its power spectrum to be unchanged through a degree-by-degree scaling. Histograms resulting from this approach yield very similar distributions (see Figure A.3 for a representative example) and virtually the same values for the parameter estimates and their standard errors (Tables A.7 and A.8). If one relaxes the constraint by requiring only that the total power remains unchanged, then the resulting histogram distributions are narrower than the others (Figure A.4). These smaller values for the standard errors in the parameter estimates likely coincide in the limit of very large numbers of trials with those determined from the elements of the variance vector $v \equiv \chi^2_{post} \text{diag}\left[\left[R^TWR\right]^{-1}\right]$ (Table A.8), where $\chi^2_{post}$, $R$, and $W$ are defined explicitly in Chapter 2.

**DATA IMPORTANCE**

The contribution of each datum in our least-squares estimation of regional averages can be evaluated by calculating the data importance [Minster et al., 1974]. The data importance $D_k$ of the $k$th datum is related to the diagonal components of the symmetric operator $P = W^{-1/2}R(R^TW^{-1}R)^{-1}R^TW^{-1/2}$ via $D_k = P_{kk}$ (summation not implied). The data importance of a datum can range from a value of zero (no importance) to a value of one. A data importance of one indicates that this datum is responsible for determining the equivalent of one parameter's value. The data importance of a particular datum is independent of the value of the datum. Rather, it is dependent on the part of the model space that it covers.
To assess the relative importance of each spherical-harmonic degree that we use in our projections, we plot data importance as a function of angular degree for projections onto five regionalizations (Figure A.5). For the seismic study, degree one is clearly the most important. In fact, most of the relevant information is contained within the data below degree six.

**COMPARISON OF RESULTS: S12_WM13 AND SH.10c.17**

To compare with our results from S12_WM13, we project onto GTR1 travel-time anomalies predicted by a different global tomographic model, SH.10c.17 [Masters et al., 1992]. Using the same suite of adjacent 100-km thick intervals, we find that our results from the use of this model are qualitatively the same as those we obtain using S12_WM13 (compare Figures 2.1 and A.6). Although SH.10c.17 is defined with spherical shells rather than with polynomials, which can induce smearing across layers, the relationship between surface tectonics and upper-mantle shear-wave heterogeneity is highly significant (> 99% confidence) within the 40-440 km depth interval. Within this depth interval, the variance reductions obtained with the actual orientation of the tectonic sphere are almost as large as the maximum variance reduction achieved from 10,000 random orientations of the tectonic sphere. As with S12_WM13, there is a clear divergence between the maximum variance reduction and the actual variance reduction below 440 km depth. Although the projections using each of the two tomographic models yield the same conclusion about the depth extent of the relationship between surface geology and upper-mantle shear-wave heterogeneity, above this depth, the projection of SH.10c.17 onto GTR1 explains substantially less of the variance than the corresponding projection of S12_WM13 (Figures 2.1 and A.6). But S12_WM13 contains more surface-wave data, and hence should resolve upper-mantle structure
better than SH.10c.17. It is notable that as the upper mantle is better resolved, the geology can explain more of the variance in the data.

To further compare the results obtained from these two tomographic models, we plot the projections of the travel-time anomalies associated with the upper four 100-km-thick depth intervals (Figures A.7 - A.10 to compare, respectively, with Figures 2.4 - 2.6 and A.11; above 340 km, we use a 0.5% contour spacing for SH.10c.17, in contrast to the 1% spacing for S12_WM13, because the magnitude of the anomalies in the former model are roughly half those of the latter model near the surface). Both tomographic models predict high velocity anomalies beneath North America, Eurasia, Africa, and Australia, although in contrast with the results from S12_WM13, the travel-time anomalies predicted by SH.10c.17 beneath Australia are not uniformly negative. The travel-time anomalies associated with South America and Antarctica provide large discrepancies: S12_WM13 predicts that these continents are seismically fast, whereas SH.10c.17 predicts that these continents have roughly average properties within this depth interval. In regard to the anomaly for South America, it is perhaps surprising that for SH.10c.17, the sign of the travel-time anomaly changes progressively from positive to negative within the 40-440 km depth interval (Figures A.7a, A.8a, A.9a, A.10a), in contrast to S12_WM13, for which the anomaly remains negative for that entire interval (Figures 2.4a, 2.5a, 2.6a, and A.11a). The anomaly in the vicinity of the Afar triple junction in S12_WM13 is not visible in SH.10c.17. It will be interesting to see if this feature is robust in future models and requires explanation, or whether it could be a result of poorly resolving the Red Sea and smearing its signal into Africa. As with S12_WM13, the SH.10c.17 projections indicate that the Atlantic ocean is seismically faster than the Pacific ocean (Figures A.7c, A.8c, A.9c). From the figures corresponding to these four depth intervals, it is clear that although there are distinct differences between the travel-times predicted by these tomographic
models, both models predict that continents are generally seismically faster than oceans, even at depths around 400 km.

ADDITIONAL ILLUSTRATIONS

Figures A.11 through A.13 continue the progression in depth interval that ended in Chapter 2 with the 240-340 km depth interval (Figure 2.6). Respectively, these figures display, for the 340-440, 440-540, and 540-640 km depth intervals, (a) corresponding one-way S-wave travel-time anomalies, (b) projections of these anomalies onto an ocean-continent function \((ABC, QPS)\) derived from GTR1, and (c) residual maps. With the tectonic sphere represented by \((A, B, C, Q, PS)\), Figures A.14 through A.18 show, respectively, analogous plots of the radial component of the gravity field and contributions to the geoid of the upper 120 km \([Hager, 1983]\), lower mantle \([Hager and Clayton, 1989]\), slabs \([Hager and Clayton, 1989]\), and remnant glacial isostatic disequilibrium \([Hager et al., 1984]\). Figure A.19 displays the degree-by-degree correlations between the non-hydrostatic geoid and, separately, ocean-continent functions derived from GTR1 and Mauk \([1977]\). From this figure it is clear that there is a relatively weak correlation between the geoid and oceans (or continents), save possibly for \(l=2\) and \(l=10\). Note that in Chapter 3 we make the observation that correlations \((l=2-8)\) between the geoid and the ocean-continent function is -0.28 with Mauk's \([1977]\) ocean-continent function and -0.18 with GTR1's. But this difference cannot be geophysically significant since there are only minor differences between the two ocean-continent functions, as we show in this figure via a degree-by-degree correlation between these two functions.

As we have discussed, Forte et al. \([1995]\) found that the long-wavelength non-hydrostatic geoid is negatively correlated with continents. The degree-two term in the geoid dominates this correlation because of its large-magnitude and be-
cause the corresponding term in the ocean-continent function has a high data importance relative to the higher-order terms (Figure A.5). On the other hand, if one band-passed the geoid by considering only spherical-harmonic degrees $l = 6 - 12$, one would find a positive correlation between the long-wavelength geoid and continents, as can be seen in Figure A.19. Its geophysical significance, however, is questionable. For example, Richards and Hager [1988] demonstrated that platforms and shields (they used the term "shields" to indicate essentially the same provinces as are included in $P$ and $S$) are (weakly) anti-correlated with the long-wavelength geoid (see Figure A.19). So, it is clear that one can obtain positive or negative correlations between the long-wavelength geoid and continents (or some subset) depending on the degree-band that one considers and how one weights the individual coefficients. These differences further substantiate the need for appropriately estimating uncertainties to understand the significance to attach to these types of results.

Acknowledgments. Figures A.1 and A.7 - A.18 were created using the Generic Mapping Tools system [Wessel and Smith, 1991].
REFERENCES


Masters, T.G., H. Bolton, and P. Shearer, Large-scale 3-dimensional structure of the mantle (abs), *EOS Trans. AGU (Spring Suppl.)*, 73, 201, 1992.


TABLES

Table A.1. GTR1

<table>
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<tr>
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<td>C</td>
<td>Old oceans (&gt; 100 My)</td>
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<td></td>
<td></td>
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<tr>
<td>Continents</td>
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<td>Q</td>
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</tr>
<tr>
<td>S</td>
<td>Precambrian shields and platforms</td>
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Table A.2. Okal [1977]

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<tr>
<td>S</td>
<td>Shields</td>
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Table A.3. *Mauk* [1977]

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<td>2</td>
<td>Anomaly 5-6 (10-20 My)</td>
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<td>Anomaly 6-13 (20-38 My)</td>
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<td>Sea floor older than 140 My</td>
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<td>Island arcs</td>
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<td>Mesozoic volcanics</td>
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</tr>
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<td>Cenozoic volcanics</td>
<td>1.4</td>
</tr>
<tr>
<td>13</td>
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<td>Early Paleozoic orogeny</td>
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<td>Archaean shield</td>
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Table A.4. S12_WM13: Regional averages and variance reductions associated with the 340-440 km depth interval obtained using, separately, GTR1 and the tectonic regionalization from *Mauk* [1977].

<table>
<thead>
<tr>
<th></th>
<th>$\tau_{ABC}$ (%)</th>
<th>$\tau_{QC}$ (%)</th>
<th>$\tau_{8-14,16,17}$ (%)</th>
<th>$\tau_{PS}$ (%)</th>
<th>$\tau_{15,18-20}$ (%)</th>
<th>Variance Reduction (%)</th>
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<tr>
<td>Mauk</td>
<td>0.4</td>
<td>-0.3</td>
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<td></td>
<td></td>
<td>43</td>
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<tr>
<td>GTR1</td>
<td>0.4</td>
<td>-0.4</td>
<td>-1.0</td>
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<td>43</td>
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Table A.5. Values of $F$ corresponding to the probability that the reduction in our values of chi-square are due to chance. We apply a 1% criterion (bold).

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<th>Criterion (%)</th>
<th>F</th>
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<tr>
<td>10.0</td>
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<tr>
<td>5.0</td>
<td>3.9</td>
</tr>
<tr>
<td>2.5</td>
<td>5.1</td>
</tr>
<tr>
<td>1.0</td>
<td><strong>6.8</strong></td>
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<tr>
<td>0.5</td>
<td>8.1</td>
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<tr>
<td>0.1</td>
<td>11.3</td>
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Table A.6. F-test results from projections of shear-wave travel-time anomalies within 100-km-thick layers onto surface regionalizations. The regionalizations selected according to a 1% F-test criterion are shown in bold.

<table>
<thead>
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<th>Interval (km)</th>
<th>Regionalization</th>
<th>Variance Reduction (%)</th>
<th>F</th>
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<tr>
<td>40-140</td>
<td>ABQ</td>
<td>53.5</td>
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<td></td>
<td>A BQ</td>
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<td>A B Q</td>
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<td>A B Q</td>
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<td></td>
<td>A B Q</td>
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<tr>
<td>140-240</td>
<td>AB C Q</td>
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<tr>
<td></td>
<td>A B C Q</td>
<td>70.5</td>
<td>45.8</td>
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<td></td>
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<td>A B C Q</td>
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<td></td>
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<tr>
<td>240-340</td>
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<td></td>
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<tr>
<td></td>
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<td>A B C Q</td>
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<td>1.1</td>
</tr>
<tr>
<td></td>
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<tr>
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<td>ABC Q</td>
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Table A.7. GEM-T3 ($l = 2$-$12$) projected onto ($A, B, C, Q, PS$): Parameter estimates determined from four different methods: (1) random rotation of the data-residual sphere with respect to the tectonic sphere; (2) random assignment of spherical-harmonic coefficients of the data-residual keeping its power spectrum unchanged; (3) random assignment of spherical-harmonic coefficients of the data-residual sphere keeping its total power unchanged; and (4) weighted least-squares inversion. All Monte Carlo simulations contain 10,000 random trials.

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<td>-0.5</td>
<td>-0.5</td>
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<td>2.8</td>
<td>2.8</td>
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</tbody>
</table>
Table A.8. GEM-T3 ($l = 2-12$) projected onto $(A, B, C, Q, PS)$: Statistical standard errors in parameter estimates determined from four different methods: (1) random rotation of the data-residual sphere with respect to the tectonic sphere; (2) random assignment of spherical-harmonic coefficients of the data-residual sphere keeping its power spectrum unchanged; (3) random assignment of spherical-harmonic coefficients of the data-residual sphere keeping its total power unchanged; and (4) square-root of the diagonal elements of the postfit covariance matrix obtained from the weighted least-squares inversion. All Monte Carlo simulations contain 10,000 random trials.

<table>
<thead>
<tr>
<th>GEM-T3 (m)</th>
<th>Random Rotations</th>
<th>Power Spectrum Constraint</th>
<th>Total Power Constraint</th>
<th>Least-Squares Inversion</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\sigma(\gamma_A)$</td>
<td>18.8</td>
<td>19.0</td>
<td>11.3</td>
<td>11.5</td>
</tr>
<tr>
<td>$\sigma(\gamma_B)$</td>
<td>9.2</td>
<td>9.5</td>
<td>6.1</td>
<td>6.2</td>
</tr>
<tr>
<td>$\sigma(\gamma_C)$</td>
<td>24.4</td>
<td>25.3</td>
<td>10.9</td>
<td>11.2</td>
</tr>
<tr>
<td>$\sigma(\gamma_Q)$</td>
<td>19.3</td>
<td>19.5</td>
<td>9.0</td>
<td>9.3</td>
</tr>
<tr>
<td>$\sigma(\gamma_{PS})$</td>
<td>10.9</td>
<td>11.1</td>
<td>8.9</td>
<td>9.2</td>
</tr>
</tbody>
</table>
FIGURE CAPTIONS

Fig. A.1. GTR1 displayed, as are all of our plots of the earth's surface, on a Hammer equal-area projection. See Table A.1 for a description of each region.

Fig. A.2. Degree-by-degree correlations between platforms (P) and shields (S). Dotted lines indicate significance levels of 80%, 90%, 95%, and 99%.

Fig. A.3. GEM-T3 ($l = 2$-$12$): Histograms of parameter values (a) $\gamma_A$, (b) $\gamma_B$, (c) $\gamma_C$, (d) $\gamma_Q$, (e) $\gamma_{PS}$ obtained from Monte Carlo simulations. Unshaded rectangles represent projections onto the tectonic sphere of 10,000 random orientations of the data-residual sphere. Shaded rectangles represent projections onto the tectonic sphere of 10,000 synthetic data-residual spheres each of which has the same power spectrum. Each of these synthetic spheres is determined by assigning random values to the spherical-harmonic coefficients. Gaussian distributions, determined by the standard deviation, mean, and area of each shaded histogram, are superposed. (f) Histogram of variance reduction obtained from projections onto the tectonic sphere of 10,000 synthetic data-residual spheres each of which has the same power spectrum. The shaded and unshaded arrows indicate the variance reductions associated with the actual orientation and the maximum variance reduction, respectively.

Fig. A.4. Same as Figure A.3 for the radial component of the gravity field.

Fig. A.5. Data importance as a function of spherical-harmonic degree ($l$) associated with inversions based on the following regionalizations: thick solid line = $(A, B, C, Q, PS)$ used in our gravity field study ($l = 2$-$12$), solid line = $(A, B, CP, Q, S)$, dashed line = $(A, B, CQ, PS)$, dash-dotted line = $(ABC, Q, PS)$, and dotted line = $(ABC, QPS)$. 
Fig. A.6. SH.10c.17 ($l = 1-10$): One-way $S$-wave travel-time anomalies associated with 100 km-thick depth intervals projected onto GTR1. Open circles indicate the variance reductions associated with the actual orientation of the tectonic sphere. Asterisks represent the maximum variance reduction achieved from 10,000 random rotations of the tectonic sphere (see text for description of random distribution). The horizontal line bisecting each open circle indicates the interval's thickness. From bottom to top, the three regions of gray scale correspond to confidence levels of $<75\%$, 75-95\%, and 95-99% as determined from the Monte Carlo simulations.

Fig. A.7. (a) One-way $S$-wave travel-time anomalies from SH.10c.17 ($l = 1-10$) associated with the 40-140 km depth interval, (b) Projection of (a) onto $(A, B, CP, Q, S)$, and (c) Residual: (a) - (b). Positive contour lines are dashed and the zero contour line is thick. The contour interval is 0.5\% of the mean travel time.

Fig. A.8. (a) One-way $S$-wave travel-time anomalies from SH.10c.17 ($l = 1-10$) associated with the 140-240 km depth interval, (b) Projection of (a) onto $(A, B, CQ, PS)$, and (c) Residual: (a) - (b). Positive contour lines are dashed and the zero contour line is thick. The contour interval is 0.5\% of the mean travel time.

Fig. A.9. (a) One-way $S$-wave travel-time anomalies from SH.10c.17 ($l = 1-10$) associated with the 240-340 km depth interval, (b) Projection of (a) onto $(ABC, Q, PS)$, and (c) Residual: (a) - (b). Positive contour lines are dashed and the zero contour line is thick. The contour interval is 0.5\% of the mean travel time.

Fig. A.10. (a) One-way $S$-wave travel-time anomalies from SH.10c.17 ($l = 1-10$) associated with the 340-440 km depth interval, (b) Projection of (a) onto $(ABC, QPS)$, and (c) Residual: (a) - (b). Positive contour lines are dashed and the zero contour line is thick. The contour interval is 0.5\% of the mean travel time.
Fig. A.11. (a) One-way S-wave travel-time anomalies from S12_WM13 \((l = 1-12)\) associated with the 340-440 km depth interval, (b) Projection of (a) onto \((ABC, QPS)\), and (c) Residual: (a) - (b). Negative contour lines are dashed and the zero contour line is thick. The contour interval is 0.5% of the mean travel time.

Fig. A.12. Same as Figure A.11 except for the 440-540 km depth interval.

Fig. A.13. Same as Figure A.11 except for the 540-640 km depth interval.

Fig. A.14. (a) Radial component of the gravity field \((l = 2-12)\), (b) Projection of (a) onto \((A, B, C, Q, PS)\), and (c) Residual: (a) - (b). Negative contour lines are dashed and the zero contour line is thick. The contour interval is 20 mgal.

Fig. A.15. (a) Contribution of the upper 120 km to the geoid \((l = 2-12)\), (b) Projection of (a) onto \((A, B, C, Q, PS)\), and (c) Residual: (a) - (b). Negative contour lines are dashed and the zero contour line is thick. The contour interval is 2 m.

Fig. A.16. (a) Contribution of the lower mantle to the geoid \((l = 2-12)\), (b) Projection of (a) onto \((A, B, C, Q, PS)\), and (c) Residual: (a) - (b). Negative contour lines are dashed and the zero contour line is thick. The contour interval is 10 m.

Fig. A.17. (a) Contribution of slabs to the geoid \((l = 2-12)\), (b) Projection of (a) onto \((A, B, C, Q, PS)\), and (c) Residual: (a) - (b). Negative contour lines are dashed and the zero contour line is thick. The contour interval is 10 m.

Fig. A.18. (a) Contribution of remnant glacial isostatic disequilibrium to the geoid \((l = 2-12)\), (b) Projection of (a) onto \((A, B, C, Q, PS)\), and (c) Residual: (a) - (b). Negative contour lines are dashed and the zero contour line is thick. The contour interval is 5 m.
Fig. A.19. Degree-by-degree correlations between (1) the non-hydrostatic geoid and platforms and shields (PS) (dash-dotted line), (2) the non-hydrostatic geoid and an ocean-continent function derived from GTR1 (thin solid line), (3) the non-hydrostatic geoid and an ocean-continent function derived from Mauk [1977] (thin dashed line), and (4) the two ocean-continent functions (thick solid line). Dotted lines indicate significance levels of 80%, 90%, 95%, and 99%.
Figure A.2
Figure A.4
Figure A.5
Figure A.6
Figure A.7
Figure A.8
Figure A.11
Figure A.13
Figure A.15
Figure A.17
Figure A.18
Figure A.19
APPENDIX B

SUPPLEMENT TO CHAPTER 4

PREFACE

In Chapter 4, we discuss a series of two-dimensional numerical experiments in which we investigate the effects on continental tectosphere stability of several factors: (1) activation energy (used to define the temperature-dependence of viscosity), (2) compositional buoyancy, and (3) linear or nonlinear rheology. In this appendix, we first present an analytical analysis of continental tectosphere stability. The study [Shapiro and Jordan, 1989] is necessarily simplified and does not include some very important effects such as the temperature- and stress-dependence of viscosity. However, the results from this exercise provide us with some insight as to the probable modes of instability. In the second section of this appendix, we consider numerically the regime in which the convective tractions on the base of the continental tectosphere are much larger than the few bar range estimated for the earth [e.g., Hager and O'Connell, 1981]. These numerical experiments [Shapiro et al., 1991a] allow us to determine the maximum contrast in viscosity between the continental tectosphere and the mantle necessary to prevent destruction of the continental tectosphere. Finally, in the third section of this appendix, we discuss the sensitivity of our results from Chapter 4 to the upper-limit chosen for (dimensionless) viscosity, finite element grid resolution, domain symmetry, and velocity, temperature, and composition boundary conditions. We include at the end of this appendix a representative sample of figures corresponding to the numerical experiments outlined in Chapter 4.
LINEAR STABILITY ANALYSIS

The thermal Rayleigh number appropriate for a Cartesian regime with fixed temperatures along the top and bottom and containing a fluid with constant properties is \(Ra_t \equiv \rho \alpha \Delta T g d^3 / \eta \kappa_t\) where \(\rho\) is the density; \(\alpha\) the coefficient of volume expansion due to temperature; \(\Delta T\) the (uniform) temperature drop across the depth interval, \(d\); \(g\) the acceleration due to gravity; \(\eta\) the dynamic viscosity; and \(\kappa_t\) the thermal diffusivity. Consider the situation envisaged by Jordan [1975], where the mass excesses that would be associated with a cold, thick, thermal boundary layer (TBL) are locally compensated by variations in composition. To characterize the compositional effect on buoyancy, we introduce a buoyancy ratio defined as

\[ \Delta B = \frac{\Delta \hat{\rho}}{\rho \alpha \Delta T} \]

where \(\Delta \hat{\rho}\) is the change in normative density \(\hat{\rho}\) and \(\Delta T\) represents a deviation from the reference temperature (see Chapter 4). If we ignore lateral variations and assume that the normative density has the same depth dependence as temperature, the following inequality determines marginal stability:

\[ B \geq 1 - \frac{Ra_4}{Ra_t} \]

[Jordan, 1988] where \(Ra_4\) is the critical Rayleigh number. When the buoyancy due to composition exactly cancels that due to temperature at every depth between the base of the mechanical boundary layer and the base of the CBL, the isopycnic equation

\[ \delta \hat{\rho}(z) = \rho \alpha \delta T(z) \]

[Jordan, 1988] and the above inequality are both satisfied and the configuration is stable. Of course, some subisopycnic states also satisfy the inequality.

However, the above stability analysis ignores the potentially destabilizing effect of horizontal gradients. To investigate this effect, we follow Stevenson's [1979] two-dimensional approach and model the transition between oceanic and continental tectospheric conditions with linear thermal and compositional gradients. We consider an incompressible ideal Newtonian fluid with constant properties confined within two horizontal free surfaces of infinite extent. The linearized two-dimensional Boussinesq equations for flows with negligible Reynolds numbers can
be written in dimensionless form in terms of the stream function, \( \Psi(x, z, t) \), with \( x \) the horizontal and \( z \) the vertical coordinate:

\[
\nabla^4 \Psi = Ra_T \frac{\partial T}{\partial x} - Ra_c \frac{\partial C}{\partial x}
\]

\[
\left( \frac{\partial}{\partial t} - \nabla^2 \right) T = -\frac{\partial \Psi}{\partial x}
\]

\[
\left( \frac{\partial}{\partial t} - \frac{1}{Le} \nabla^2 \right) C = -\frac{\partial \Psi}{\partial x}
\]

[e.g., Veronis, 1965]. Here the Lewis number is defined as \( Le \equiv \kappa_T / \kappa_c \) where \( \kappa_c \) denotes compositional diffusivity. The compositional Rayleigh number is defined as \( Ra_c \equiv \rho \beta \Delta C g d^2 / \eta \kappa_T \) where \( \Delta C \) is the compositional equivalent of \( \Delta T \) and \( \beta \) is the compositional equivalent of \( \alpha \) defined with the opposite sign.

We define \( \theta \equiv (\theta_x, \theta_z) \) and \( \phi \equiv (\phi_x, \phi_z) \) as the gradients in temperature and composition, respectively, and let the horizontal density variations due to changes in temperature exactly offset those due to composition, \( \alpha \theta_x = \beta \phi_x \). A spatial stream function, \( \psi \), defined from the form \( \Psi = \psi(x, z) e^{i(\theta_x, \theta_z)} \) satisfies the eighth-order partial differential equation:

\[
(\tau^{-1} - \nabla^2) \left( \tau^{-1} - \frac{1}{Le} \nabla^2 \right) \nabla^4 \psi = \tau^{-1} Ra_T \left( B - 1 \right) \frac{\partial^2 \psi}{\partial x^2}
\]

\[
-Ra_T \left( B \frac{\partial^2}{\partial x^2} - \gamma \frac{\partial^2}{\partial x \partial z} \right) \nabla^2 \psi + \frac{Ra_T}{Le} \left( \frac{\partial^2}{\partial x^2} - \gamma \frac{\partial^2}{\partial x \partial z} \right) \nabla^2 \psi
\]

(B.1)

Here \( \tau \) is time, \( \tau \) represents the characteristic time constant associated with growth or decay, \( \gamma \) is the ratio of horizontal to vertical temperature gradients, \( \gamma \equiv \theta_x / \theta_z \), and \( B = Ra_c / Ra_T \). There are solutions to this equation of the form: \( \psi = e^{i(k_x x + k_z z)} \).

Substituting \( \psi \) into (B.1), and letting \( \zeta = k_x / k_z \), we find the characteristic equation:
\[ \tau^{-2} + \left[ k_2^2 \left( 1 + \frac{1}{Le} \right) \left( 1 + \zeta^2 \right) + \frac{Ra_f(B - 1)}{k_2^2 + (1 + \zeta^2)^2} \right] \tau^{-1} + \frac{Ra_f \left( B \zeta - \gamma \right) - \frac{1}{Le} \left( \zeta - \gamma \right)}{1 + \zeta^2} = 0 \]

Assuming infinite Lewis number (see Chapter 4), we obtain:

\[ \tau^{-2} + \left[ k_2^2 \left( 1 + \zeta^2 \right) + \frac{Ra_f(B - 1)}{k_2^2 + (1 + \zeta^2)^2} \right] \tau^{-1} + \frac{Ra_f \left( B \zeta - \gamma \right)}{1 + \zeta^2} = 0 \]

For the isopycnic condition, the amplitudes of modes which satisfy the provision that \( \zeta - \gamma > 0 \), decay exponentially with time. Figure B.1 shows, for \( \gamma = 1 \), contours of characteristic times of decay and growth.

Using representative values of the relevant parameters, i.e., \( \kappa_f = 10^6 \text{ m}^2 \text{ s}^{-1} \), \( B = 1 \), \( d = 200 \text{ km} \), a kinematic viscosity \( \nu = \eta/\rho = 3 \times 10^{17} \text{ m}^2 \text{ s}^{-1} \), and \( \alpha \theta_z = \alpha \theta_s = 10^{-4} \text{ m}^{-1} \) (i.e., a 1% density change in \( 10^3 \) km), Stevenson [1979] found that the most unstable mode has a time constant of about 200 My with a horizontal length scale (between upwellings) of about 500 km (Figure B.2) — far shorter than the several billion years required by the longevity constraint [e.g., Jordan, 1988]. The vertical gradients used by Stevenson [1979] correspond to a 0.2% density change over the depth of the layer — a value about five times smaller than that appropriate for the continental tectosphere [Jordan, 1988]. Using \( \alpha \theta_z = 5 \alpha \theta_s = 5 \times 10^{-4} \text{ m}^{-1} \) also corresponds to rapidly growing modes but with horizontal length scales of the order of the size of continents (Figure B.2).

Figure B.3 displays a plot of the time constants for the most unstable modes as a function of horizontal gradient for \( \alpha \theta_z d = 1.8\% \), \( g = 9.8 \text{ m s}^{-2} \), \( \alpha = 3 \times 10^{-5} ^\circ\text{C}^{-1} \), \( \Delta T = 600 ^\circ\text{C} \), \( \kappa_f = 10^6 \text{ m}^2 \text{ s}^{-1} \), \( d = 300 \text{ km} \), \( \rho = 3.3 \text{ Mg m}^{-3} \), and \( \eta = 10^{21} \text{ Pa s} \). Large horizontal thermal and compositional gradients are required at the periphery of cratons to match oceanic and continental tectospheric conditions. For such gradients, the continental tectosphere is very unstable; thus with \( \alpha \theta_z d = \)
1.8%, it could not survive unless the horizontal thermal gradients were unreasonably small. However, in the middle of a craton these gradients are small enough to allow survival of the structure. As shown in the blow-up of this figure, the most unstable modes associated with small horizontal gradients have horizontal length scales much larger than the craton itself and have time constants greater than the age of the earth.

Although this analytical exercise is necessarily simplified, the conclusion that instability in the continental tectosphere is at least in part governed by horizontal thermal and compositional gradients is strong.

**HIGH CONVECTIVE STRESS**

**NUMERICAL FINITE AMPLITUDE ANALYSIS**

A continental tectosphere must also be able to withstand disruption caused by the basal stresses associated with a convecting mantle. Modeling numerically an earth with a realistic Rayleigh number is computationally intensive even in two dimensions because of the high spatial and temporal resolution required to accurately model the regions of high velocity flow. To compare with the results reported in Chapter 4, we investigate the following, less realistic, case where the basal convective stresses are approximately 60 bars — about 10 times that considered to be appropriate for the earth [e.g., *Hager and O'Connell*, 1981].

Our two-dimensional box contains a 400 km thick, 3000 km long, continental tectosphere described by a suite of depth-dependent viscosity profiles. By considering only spatially-fixed viscosities, we avoid using the computationally intensive algorithm required to model time-dependent material properties. While such a simplification is not useful for approximating temperature-dependent viscosity, it may be appropriate for modeling composition-dependent viscosity as long as the compositional field remains stationary. The continental tectosphere is bordered al-
ternately by four variable-length (800 - 1600 km), 125 km thick, high (dimensionless) viscosity ($\eta = 1000$; $\eta = 1$ corresponds to approximately $1.5 \times 10^{23}$ Pa s) TBLs and three 400 km long, 150 km thick, lower viscosity ($\eta = 0.1$) weak zones. Below these structures we include a mantle of unit (dimensionless) viscosity extending to a depth of 3000 km. In the upper 125 km, each continental tectosphere viscosity profile mimics the viscosity of the upper TBLs. Below this depth, each profile has a different starting value, $\eta_o$, and then decreases exponentially with increasing depth to match the mantle viscosity at the base of the continental tectosphere (Figure B.4). We impose free-slip velocity boundary conditions along the top and bottom and a wrap-around velocity boundary condition along the sides. We fix the (dimensionless) temperatures along the top and bottom of the box at zero and one, respectively. To obtain an initial thermal structure appropriate for a continental tectosphere immersed in a convecting mantle, we (temporarily) fix the temperature at a depth of 2700 km (creating a lower TBL) and along the bottom of the oceanic TBLs, on the sides and bases of the weak zones, and on the sides and base of the continental tectosphere, to a value ($T_m = 0.6$) approximating the mean temperature of such a convecting system. Subject to these temperature boundary conditions, we use ConMan [King et al., 1990] to solve the thermal conduction equation. From the resulting temperature field, we create a composition field such that at every node other than those in the lower TBL, $C = T_m - T$. For the nodes in the lower TBL, we let $C = 0$ since we assume that the tectosphere is the only chemically compensated region. With these fields, the basal and surficial temperature boundary conditions, and $Ra_T = 5 \times 10^5$, we conduct numerical experiments using ConMan with buoyancy ratios of zero and one and the various continental tectosphere viscosity profiles discussed above.

Figure B.5 shows a snapshot of the temperature and composition fields for the continental tectosphere's viscosity described by the $\eta_o = 1$ profile shown in
Figure B.4. The initial upwelling shown directly below the continental tectosphere causes the composition to be washed out of the tectosphere and into the mantle. Figure B.6 shows a temperature and composition field snapshot for the tectosphere's viscosity described by the $\eta_0 = 631$ profile shown in Figure B.4. In this case, most of the tectosphere has remained undisturbed by the underlying convection even after 14 billion years. Stability profiles based on which nodes within the continental tectosphere contain at least 75% of their initial composition values represents our (arbitrary) operational definition of stability. Identical stability profiles result for buoyancy ratios of zero and one (Figure B.4). From these profiles one can see that viscosity contrasts of 20-30 are sufficient to maintain stability. Our stability criterion is clearly ad-hoc but the fact that the results are sensitive to viscosity but insensitive to compositional buoyancy implies that for this high-stress case, viscosity can stabilize a tectosphere while the effect of compositional buoyancy on stability is negligible.

ADDITIONAL ILLUSTRATIONS

To complement the figures contained within Chapter 4, we include here additional figures illustrating, for a few representative experiments, the time-evolution of the composition and temperature fields. In addition, for each of these sample experiments, we include a multi-plot figure displaying geoid height anomalies ($t = 0, 10, 100, \text{ and } 1000 \text{ My}$), dynamic topography ($t = 0, 10, 100, \text{ and } 1000 \text{ My}$), viscosity ($t = 0, 1000 \text{ My}$), stress ($t = 0, 1000 \text{ My}$), composition and temperature difference fields ($C(1000) - C(0); T(1000) - T(0)$), and the time progression of $A$ and $\Delta z_{\text{CML}}$ (both defined in Chapter 4). For most of the remaining experiments outlined in Chapter 4, including all of those represented in Figure 4.9, we provide, in the interest of saving paper, only a suite of multi-plot figures like those described above. However, since we do not include figures showing the time evolution of
the composition and temperature fields, we replace the initial viscosity and stress fields with the final \((t = 1000 \text{ My})\) composition and temperature fields.

All of our Newtonian rheology experiments based on an activation energy of \(E^* = 522 \text{ KJ mole}^{-1}\) (Table 4.1) yield very similar flow patterns. For our representative example, we display the results from the experiment defined by \(B = 0\), and a background viscosity profile \(\eta_b(z) = \text{HGPA} [\text{Hager and Richards, 1989}]\) (Figures B.7 - B.9). The results corresponding to the other experiments within this category are shown in Figures B.10 through B.14. The results from the non-Newtonian analogs of these experiments are also insensitive to \(B\) and to \(\eta_b(z)\).

The non-Newtonian equivalents of Figures B.7 through B.14 are shown in Figures B.15 through B.22.

For many of the subsequent sensitivity tests, we use the Newtonian rheology experiment defined by the parameters: \(E^* = E^\ast_{\text{ref}}/3\), \(B = 0\), and \(\eta_b(z) = \text{HGPA}\) as a test case (Figures B.23 - B.25). Figures B.26 though B.36 contain the multi-figure plots (see above) associated with the results from all of our other experiments which use an activation energy of \(E^* = E^\ast_{\text{ref}}/3\).

Because, in our experiments, activation energies of \(E^* = E^\ast_{\text{ref}}/9\) lead to rapid destruction of the continental CBL, we include here the results from only one example experiment \((B = 1, n = 3, \eta_b(z) = \text{NLO} [\text{Nakada and Lambeck, 1989}]\) (Figures B.37 - B.39). By comparing these figures with the (otherwise) analogous ones with \(E^* = E^\ast_{\text{ref}}/3\) (Figures 4.6, 4.7, and 4.8), we see that reducing the activation energy from about 180 to 60 KJ mole\(^{-1}\), greatly affects the long-term thickness of the CBL. Without a sufficiently high activation energy to suppress the instabilities caused by the large thermal gradients, the continental tectosphere will not survive for long.
SENSITIVITY TESTS

1. MAXIMUM VISCOSITY

With standard numerical convection experiments, one can adequately model rigid sections (e.g., plates) by fixing the viscosity of these regions to a value three orders of magnitude larger than the background viscosity of the convecting domain [e.g., King, 1990]. Typically, in experiments using either spatially-fixed or temperature-dependent viscosity, one limits the range of viscosity values by choosing some maximum value. However, in Chapter 4 we investigate the dynamics of thermal and compositional boundary layers, rather than that of the underlying fluid system. For this class of study, results depend strongly on the maximum viscosity value selected. As an example, consider the disparity in the conclusions drawn from two experiments which differ only in the upper bound of allowable values of viscosity. Following the experimental design outlined in Chapter 4 (with background viscosity $\eta_b(z) = 1$), we clip the viscosity at two values: $\eta_{\text{clip}} = 10^3$ and $10^4$. Note, with $E^* = E_{\text{ref}}^*/9$, $B = 0$, $n = 1$, and $\eta_b(z) = \text{HGPA}$, clipping the viscosity at values ranging from $10^5$ to $10^{10}$ yield essentially identical flow fields.

Clipping the viscosity at $\eta_{\text{clip}} = 10^3$, Shapiro et al. [1991b] found two modes of tectosphere instability: (1) a short-wavelength mode initiated by the large horizontal thermal gradients at the periphery of the continental tectosphere required to match oceanic and continental conditions and (2) a long-wavelength mode associated with the whole-scale sinking of the continental tectosphere. The first instability is characterized by a "drip" which propagates from the edge towards the center of the structure. Strong temperature-dependence of viscosity helps to stabilize this mode. But, once this instability has been suppressed, the second instability associated with the sinking of the continental tectosphere becomes the dominant mode of destruction. Adding compositional buoyancy to the continental tectosphere sig-
nificantly reduces its rate of sinking. In short, Shapiro et al. [1991b] concluded that by employing the combination of temperature-dependent viscosity and compositional buoyancy, one could contain these modes of instability and preserve a continental tectosphere for billions of years. This sinking mode could be relevant to earth processes if faulting, initiated by stresses of approximately 100 bars, caused deformation of the lithosphere.

In contrast, no such "sinking mode" develops when we clip the viscosity at \( \eta_{\text{clip}} = 10^9 \). Instead, once the short-wavelength instability is stabilized — either solely by temperature-dependent viscosity or with the combination of compositional buoyancy and stress-dependent rheology — the continental tectosphere is stabilized.

2. finite element grid resolution

To ascertain whether our 76 x 38 grid of finite elements provides us with sufficient resolution to accurately model boundary layer dynamics, we repeat one of the Newtonian rheology experiments outlined in Chapter 4 (\( E^* = E_{\text{rd}}^* / 3, \ B = 0, \ n = 1, \ \eta_b(z) = \text{HGPA} \); see Figures B24 - B.25) using a 152 x 76 element grid. Because it is much less computationally intensive to model Newtonian than stress-dependent rheology (e.g., for a stress exponent of three, it requires roughly three times the number of CPU cycles for every time step), we choose to compare results from Newtonian calculations. Figures B.40 - B.42 show the results, after 500 My, obtained from the high resolution grids. The results shown in Figures B.40 - B.42 are virtually, indistinguishable from those shown in Figures B.23 - B.25 and demonstrate that increasing the resolution of our grid will not likely alter any of our stability assessments. Given the large number of elements within this grid, it is not cost-effective to run this experiment to our 1000 My characteristic time.
3. EXPERIMENTAL DOMAIN SYMMETRY

Our application of reflective boundary conditions produces (by definition) a symmetric experimental domain. To test the effect on our results of this enforced symmetry, we construct a two-dimensional domain with a full (uncentered) continental tectosphere of matching resolution (Figure B.43). We repeat the same experiment discussed in the previous section and obtain equivalent results (Figures B.43 to B.45 to be compared with Figures B.23 - B.25).

4. BOUNDARY CONDITIONS

Here we discuss our selection of boundary conditions. We choose a no-slip, vertical flow-through boundary condition along the base \( (u_z = \partial u_z/\partial z = 0) \), where \( u \) is velocity. By requiring the flow to be vertical at the bottom and allowing material to leave the domain, we better model the upper section of a deeper dynamic system. Applying the more traditional boundary conditions \( (u_z = \partial u_z/\partial z = 0) \) along the base results in flow fields like those shown in Figures B.46 and B.47 where parts of the tectosphere end up along the base of the domain. Fixing the temperature at \( T = 1 \) at the base of the domain does not significantly affect the flow (Figures B.48 and B.49), although the interior does not cool to as low a value. One assumption inherent in our analysis is that the oceanic TBL remains unaffected by any continental CBL decay. By fixing the temperature along the base of the oceanic TBL we can maintain its integrity. The effect of neglecting NLO's 100 fold increase in viscosity at 670 km is insignificant (Figures B.50 - B.52).

5. TECTOSPHERE WIDTH

Each of our models is characterized by a single "drip" which propagates from the edge of the CBL towards the center. To test whether a wider tectosphere would simultaneously produce more than one instability and, if so, what effect these additional instabilities would have on the dynamics, we create a tectosphere
that is twice as wide as those that we use in our standard experiments, keeping the
width of the ocean and the depth of the domain constant. Three "drips" develop
along the base of the tectosphere creating a much more complicated flow pattern be-
neath the tectosphere (Figures B.53 and B.54). Even with these additional insta-
bilities, the results from this experiment are very similar to those from the corre-
spending one with the smaller tectosphere (compare Figure 4.9 with B.55, noting
that there is a vertical exaggeration of a factor of 1.5 in Figure B.55). Both experi-
mements yield a stable tectosphere with associated geoid height anomalies which are
consistent with platforms and shields. The only significant differences in these
summary plots is in the contour map of the second invariant of the stress tensor.
The stress field corresponding with the wider tectosphere has higher values in the
non-tectosphere region. These higher stress values are a result of the additional
"drips", of which some detach from the tectosphere away from its center.

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porate compositional variations into the code.
REFERENCES


FIGURE CAPTIONS

Fig. B.1. Contours of characteristic times of decay and growth in millions of years as a function of length scale for $\gamma = 1$. Stable and unstable modes are separated by the line corresponding to $k_x = k_y$.

Fig. B.2. Time constant, $\tau$, plotted as a function of horizontal wavelength for both the situation analyzed by Stevenson [1979] and the one in which $\theta_s$ was increased by a factor of five from Stevenson's value.

Fig. B.3. Minimum time constant, $\tau_{\text{min}}$, as a function of the ratio of horizontal to vertical gradients ($\tan^{-1} \gamma$). Time constants associated with ratios in temperature gradients greater than about five degrees ($\approx 0.1$ radian) are much less than the proposed age of the continental tectosphere. The most unstable modes corresponding to such ratios when less than five degrees have time constants exceeding the proposed age of the continental tectosphere and horizontal length scales larger than the continental tectosphere itself.

Fig. B.4. Depth-dependent viscosity laws applied to the continental tectosphere. Dots correspond to the viscosity at each node. All profiles mimic the viscosity of the upper TBLs and all exponentially decrease with depth from a suite of different starting values, $\eta_o$, to match the mantle viscosity at the base of the continental tectosphere. The profiles corresponding to $\eta_o = 1$ (short-dashed) and $\eta_o = 631$ (long-dashed) are the two cases discussed in the text. Shaded dots represent those nodes which, after 14 billion years, contain at least 75% of their original composition and unshaded dots represent those nodes which contain less than this amount after this time interval.
Fig. B.5. Snapshot after 125 My of the temperature (top) and composition (bottom) fields for the viscosity profile corresponding to $\eta_0 = 1$. The initial upwelling reaches the base of the continental tectosphere and causes the material which defines the CBL to flow into the mantle. Contours of temperature and composition each have intervals of 0.1.

Fig. B.6. Snapshot after about 14 billion years of the temperature (top) and composition (bottom) fields for the viscosity profile corresponding to $\eta_0 = 631$. Much of the CBL is still present. Contours of temperature and composition each have intervals of 0.1.

Fig. B.7. Parameters: $E^* = E^*_0$, $B = 0$, $n = 1$, $\eta_h(z) = \text{HGPA}$. Four equitemporal frames: $t = (a) 0$, $(b) 50$, $(c) 100$, and $(d) 150$ My. Each frame contains (left) contours of composition (purely a tracer field having no effect on the dynamics of the fluid) with superposed velocity arrows, (center) the difference, as a function of depth, between the average mid-continental ($x \leq 200$ km) density and the average lateral density (see equation 4.2), and (mirrored - right) contours of temperature with superposed velocity arrows. Composition is contoured in increments of 0.1 with $C = 1$ at the top and $C = 0.1$ at the base of the CBL. Temperature is contoured in increments of 0.1 with $T = 0$ at the top and $T = 1$ at the base of the TBL. The $C = T = 0.1$ contours are thick.

Fig. B.8. Four equitemporal frames: $t = (a) 250$, $(b) 500$, $(c) 750$, and $(d) 1000$ My. Parameters and description of frames as in Figure B.7.

Fig. B.9. Parameters as in Figure B.7. (a) Geoid height anomalies ($\delta N$) at $t = 0$ (dotted line), 10 (dashed line), 100 (thin solid line), and 1000 (thick solid line) My. There is little difference between $\delta N$ at $t = 0$ and 10 My so the dotted and dashed lines appear superposed. (b) Dynamic topography ($h$) at $t = 0$ (dotted line), 10
(dashed line), 100 (thin solid line), and 1000 (thick solid line) My. There is little difference between \( h \) at \( t = 0 \) and 10 My so the dotted and dashed lines appear superposed. (c) Viscosity field (\( \eta \)) with superposed velocity arrows (left) and second invariant of the stress tensor (\( \pi(II) \)) (mirrored - right) at \( t = 0 \). Viscosity contours are spaced by factors of 100 with the thick line representing the lowest contour level (\( 10^{20} \) Pa s). For the stress field, the thick line represents the lowest contour level (five bars) and each succeeding contour indicates a stress value a factor of two larger than that for the immediately preceding contour. (d) Viscosity field (\( \eta \)) with superposed velocity arrows (left) and second invariant of the stress tensor (\( \pi(II) \)) (mirrored - right) at \( t = 1000 \) My. Contour intervals as in (c). (e) Initial composition field subtracted from the composition field at \( t = 1000 \) My (left) and initial temperature field subtracted from the temperature field at \( t = 1000 \) My (mirrored - right). Dimensionless contours are spaced in increments of 0.1 with dashed lines representing a loss of composition / temperature and thick solid lines representing a gain. The zero contours are shown with thin solid lines. (f) Area (\( A \)) of the -0.1 difference contour, normalized by the area of the initial oceanic and continental CBL, representing a loss of composition (thin line, open circles), and the change (\( \Delta z_{\text{CBL}} \)) in the depth of the base of the continental CBL (\( C = 0.1 \)) at 1000 My expressed as a percentage of the initial depth (thick line, asterisks). The depth of the base of the continental CBL is estimated from the median depth of the \( C = 0.1 \) contour within the continental CBL. (g) Conductive heat flux (\( Q \)) through the surface of the continent (thin line, open circles) and advective heat flux through the base of the domain (thick line, asterisks).

Fig. B.10. Parameters: \( E^* = E^{*}_{\text{ref}}, B = 1, n = 1, \eta_k(z) = \text{HGPA} \). (a) Same as Figure B.9a. (b) Same as Figure B.9b. (c) Same as Figure B.9d. (d) Same as Figure B.9e. (e) (left) Contours of composition (purely a tracer field having no ef-
fect on the dynamics of the fluid) with superposed velocity arrows, (center) the difference, as a function of depth, between the average mid-continental ($x \leq 200$ km) density and the average lateral density (see equation 4.2), and (mirrored - right) contours of temperature with superposed velocity arrows all at $t = 1000$ My. Contour levels as in Figure B.7. (f) Same as in Figure B.9f. (g) Same as in Figure B.9g.

Fig. B.11. Parameters: $E^* = E^*_{ref}$, $B = 1.5$, $n = 1$, $\eta_b(z) = \text{HGPA}$. Frames as described in Figure B.10.

Fig. B.12. Parameters: $E^* = E^*_{ref}$, $B = 0$, $n = 1$, $\eta_b(z) = \text{NLO}$. Frames as described in Figure B.10.

Fig. B.13. Parameters: $E^* = E^*_{ref}$, $B = 1$, $n = 1$, $\eta_b(z) = \text{NLO}$. Frames as described in Figure B.10.

Fig. B.14. Parameters: $E^* = E^*_{ref}$, $B = 1.5$, $n = 1$, $\eta_b(z) = \text{NLO}$. Frames as described in Figure B.10.

Fig. B.15. Parameters: $E^* = E^*_{ref}$, $B = 0$, $n = 3$, $\eta_b(z) = \text{HGPA}$. Four equitemporal frames: $t = (a) 0$, (b) 50, (c) 100, and (d) 150 My. Frames as described in Figure B.7.

Fig. B.16. Parameters as in Figure B.15. Four equitemporal frames: $t = (a) 250$, (b) 500, (c) 750, and (d) 1000 My. Frames as described in Figure B.7.

Fig. B.17. Parameters as in Figure B.15. Frames as described in Figure B.9.

Fig. B.18. Parameters: $E^* = E^*_{ref}$, $B = 1$, $n = 3$, $\eta_b(z) = \text{HGPA}$. Frames as described in Figure B.10.
Fig. B.19. Parameters: \( E^* = E_{\text{ref}}^* \), \( B = 1.5 \), \( n = 3 \), \( \eta_b(z) = \text{HGPA} \). Frames as described in Figure B.10.

Fig. B.20. Parameters: \( E^* = E_{\text{ref}}^* \), \( B = 0 \), \( n = 3 \), \( \eta_b(z) = \text{NLO} \). Frames as described in Figure B.10.

Fig. B.21. Parameters: \( E^* = E_{\text{ref}}^* \), \( B = 1 \), \( n = 3 \), \( \eta_b(z) = \text{NLO} \). Frames as described in Figure B.10.

Fig. B.22. Parameters: \( E^* = E_{\text{ref}}^* \), \( B = 1.5 \), \( n = 3 \), \( \eta_b(z) = \text{NLO} \). Frames as described in Figure B.10.

Fig. B.23. Parameters: \( E^* = E_{\text{ref}}^*/3 \), \( B = 0 \), \( n = 1 \), \( \eta_b(z) = \text{HGPA} \). Four equitemporal frames: \( t = (a) 0, (b) 50, (c) 100, \) and \( (d) 150 \) My. Frames as described in Figure B.7.

Fig. B.24. Parameters as in Figure B.23. Four equitemporal frames: \( t = (a) 250, \) \( (b) 500, (c) 750, \) and \( (d) 1000 \) My. Frames as described in Figure B.7.

Fig. B.25. Parameters as in Figure B.23. Frames as described in Figure B.9.

Fig. B.26. Parameters: \( E^* = E_{\text{ref}}^*/3 \), \( B = 1 \), \( n = 1 \), \( \eta_b(z) = \text{HGPA} \). Frames as described in Figure B.10.

Fig. B.27. Parameters: \( E^* = E_{\text{ref}}^*/3 \), \( B = 1.5 \), \( n = 1 \), \( \eta_b(z) = \text{HGPA} \). Frames as described in Figure B.10.

Fig. B.28. Parameters: \( E^* = E_{\text{ref}}^*/3 \), \( B = 0 \), \( n = 1 \), \( \eta_b(z) = \text{NLO} \). Frames as described in Figure B.10.

Fig. B.29. Parameters: \( E^* = E_{\text{ref}}^*/3 \), \( B = 1 \), \( n = 1 \), \( \eta_b(z) = \text{NLO} \). Frames as described in Figure B.10.
Fig. B.30. Parameters: \( E^* = E_{\text{ref}}^*/3, B = 1.5, n = 1, \eta_b(z) = \text{NLO} \). Frames as described in Figure B.10.

Fig. B.31. Parameters: \( E^* = E_{\text{ref}}^*/3, B = 0, n = 3, \eta_b(z) = \text{HGPA} \). Frames as described in Figure B.10.

Fig. B.32. Parameters: \( E^* = E_{\text{ref}}^*/3, B = 1, n = 3, \eta_b(z) = \text{HGPA} \). Frames as described in Figure B.10.

Fig. B.33. Parameters: \( E^* = E_{\text{ref}}^*/3, B = 1.5, n = 3, \eta_b(z) = \text{HGPA} \). Frames as described in Figure B.10.

Fig. B.34. Parameters: \( E^* = E_{\text{ref}}^*/3, B = 0, n = 3, \eta_b(z) = \text{NLO} \). Frames as described in Figure B.10.

Fig. B.35. Parameters: \( E^* = E_{\text{ref}}^*/3, B = 1, n = 3, \eta_b(z) = \text{NLO} \). Frames as described in Figure B.10.

Fig. B.36. Parameters: \( E^* = E_{\text{ref}}^*/3, B = 1.5, n = 3, \eta_b(z) = \text{NLO} \). Frames as described in Figure B.10.

Fig. B.37. Parameters: \( E^* = E_{\text{ref}}^*/9, B = 1, n = 3, \eta_b(z) = \text{NLO} \). Four equitemporal frames: \( t = (a) 0, (b) 50, (c) 100, \) and \( (d) 150 \) My. Frames as described in Figure B.7.

Fig. B.38. Parameters as in Figure B.37. Four equitemporal frames: \( t = (a) 250, (b) 500, (c) 750, \) and \( (d) 1000 \) My. Frames as described in Figure B.7.

Fig. B.39. Parameters as in Figure B.37. Frames as described in Figure B.9.
Fig. B.40. Grid: 152 x 76 array of square elements. Parameters as in Figure B.23. Frames as described in Figure B.7.

Fig. B.41. Grid and parameters as in Figure B.40. Frames as described in Figure B.7.

Fig. B.42. Grid and parameters as in Figure B.40. Frames as described in Figure B.9.

Fig. B.43. Grid: 152 x 38 array of square elements. Parameters as in Figure B.23. Frames as described in Figure B.7. Note that there is a vertical exaggeration of a factor of two.

Fig. B.44. Grid and parameters as in Figure B.43. Frames as described in Figure B.7. Note that there is a vertical exaggeration of a factor of two.

Fig. B.45. Grid and parameters as in Figure B.43. Frames as described in Figure B.9. Note that there is a vertical exaggeration of a factor of two.

Fig. B.46. Velocity boundary conditions along the bottom of the domain: 
\[ u_z = \frac{\partial u_z}{\partial z} = 0. \]
Four equitemporal frames: \( t = (a) 0, (b) 50, (c) 100, \) and \( (d) 150 \) My. Parameters as in Figure B.23. Frames as described in Figure B.7.

Fig. B.47. Boundary conditions and parameters as in Figure B.46. Four equitemporal frames: \( t = (a) 250, (b) 500, (c) 750, \) and \( (d) 1000 \) My. Frames as described in Figure B.7.

Fig. B.48. Fixed temperature \((T = 1)\) along the base of the domain. Four equitemporal frames: \( t = (a) 0, (b) 50, (c) 100, \) and \( (d) 150 \) My. Parameters as in Figure B.23. Frames as described in Figure B.7.
Fig. B.49. Boundary conditions and parameters as in Figure B.48 Four equitemporal frames: \( t = (a) 250, (b) 500, (c) 750, \) and \( (d) 1000 \) My. Frames as described in Figure B.7.

Fig. B.50. Parameters: \( E^* = E^*_{nf}/3, B = 1, n = 3, \eta_b(z) = \text{NLO}. \) Background viscosity model: NLO, including the 100 fold increase in viscosity at 670 km depth.
Four equitemporal frames: \( t = (a) 0, (b) 50, (c) 100, \) and \( (d) 150 \) My. Frames as described in Figure B.7. Compare with Figure 4.6 where the background viscosity model does not include the 100 fold increase.

Fig. B.51. Background viscosity model and parameters as in Figure B.50. Four equitemporal frames: \( t = (a) 250, (b) 500, (c) 750, \) and \( (d) 1000 \) My. Frames as described in Figure B.7. Compare with Figure 4.7 where the background viscosity model does not include the 100 fold increase.

Fig. B.52. Background viscosity model and parameters as in Figure B.50. Frames as described in Figure B.9. Compare with Figure 4.8 where the background viscosity model does not include the 100 fold increase.

Fig. B.53. Parameters: \( E^* = E^*_{nf}/3, B = 1, n = 3, \eta_b(z) = \text{NLO}. \) Wider tectosphere — increased from 800 to 1600 km. Four equitemporal frames: \( t = (a) 0, (b) 50, (c) 100, \) and \( (d) 150 \) My. Frames as described in Figure B.7. Note that there is a vertical exaggeration of a factor of 1.5.

Fig. B.54. Wider tectosphere — increased from 800 to 1600 km. Four equitemporal frames: \( t = (a) 250, (b) 500, (c) 750, \) and \( (d) 1000 \) My. Parameters as in Figure B.53. Frames as described in Figure B.7. Note that there is a vertical exaggeration of a factor of 1.5.
Fig. B.55. Wider tectosphere — increased from 800 to 1600 km. Parameters as in Figure B.53. Frames as described in Figure B.9. Note that there is a vertical exaggeration of a factor of 1.5.
Figure B.1: \( \tau_{\max} \) as a Function of Length Scales for \( \tan(\theta) \), \( \theta_y = 15^\circ \)

Stable Modes

Unstable Modes

\[ \pi L_z \text{ (km)} \]

\[ \pi/k_y \text{ (km)} \]

200 400 600 800 1000
Figure B.2
\[ \chi \theta d = 1.8\%, \ d = 300 \ km \]

Figure B.3
Figure B.4
Figure B.7
Figure B.9
Figure B.10
Figure B.11
$200 - 200 = 0$

$\delta N$

$z (\text{km})$

$C(1000 \text{ My}) - C(0)$

$T(1000 \text{ My}) - T(0)$

$A; \Delta z_{\text{CBL}} / z_{\text{CBL}}(0)$

Figure B.12
Figure B.13
Figure B.14
Composition

$\delta p / \rho$

Temperature

$Q_{\text{sect}} = 11; \ Q_{\text{base}} = 3 \ \text{(mW/m}^2\text{)}$ (d)

$Q_{\text{sect}} = 11; \ Q_{\text{base}} = 10 \ \text{(mW/m}^2\text{)}$ (b)

$Q_{\text{sect}} = 11; \ Q_{\text{base}} = 2 \ \text{(mW/m}^2\text{)}$ (a)

$t = 0 \ \text{My}; \ V_{\max} = 5 \ \text{cm/yr}$

$t = 50 \ \text{My}; \ V_{\max} = 3 \ \text{cm/yr}$

$t = 100 \ \text{My}; \ V_{\max} = 2 \ \text{cm/yr}$

$t = 150 \ \text{My}; \ V_{\max} = 2 \ \text{cm/yr}$

$\times (\text{km})$

$0-2 \ %$

$0 \ 2 \ 0 \ 500 \ 1000 \ 1500 \ \times (\text{km})$

Figure B.15
Figure B.16
Figure B.17
Figure B.20
Figure B.21
Figure B.23
Figure B.24
Figure B.25
Figure B.26
Figure B.27
Figure B.29
Figure B.30
Figure B.31
Figure B.32
Figure B.33
Figure B.34
Figure B.35
Figure B.36
Composition

\[ t = 0 \text{ My}; \ V_{\text{max}} = 8 \text{ cm/yr} \]

\[ t = 50 \text{ My}; \ V_{\text{max}} = 12 \text{ cm/yr} \]

\[ t = 100 \text{ My}; \ V_{\text{max}} = 12 \text{ cm/yr} \]

\[ t = 150 \text{ My}; \ V_{\text{max}} = 10 \text{ cm/yr} \]

Temperature

\[ Q_{\text{act}} = 11; \ Q_{\text{base}} = 1 \text{ (mW/m}^2\text{)} \]

\[ Q_{\text{act}} = 11; \ Q_{\text{base}} = 27 \text{ (mW/m}^2\text{)} \]

\[ Q_{\text{act}} = 12; \ Q_{\text{base}} = 12 \text{ (mW/m}^2\text{)} \]

\[ Q_{\text{act}} = 12; \ Q_{\text{base}} = 9 \text{ (mW/m}^2\text{)} \]

Figure B.37
Figure B.39
Composition

\[ t = 0 \text{ My}; \ V_{\text{max}} = 7 \text{ cm/yr} \]

\[ t = 50 \text{ My}; \ V_{\text{max}} = 5 \text{ cm/yr} \]

\[ t = 100 \text{ My}; \ V_{\text{max}} = 5 \text{ cm/yr} \]

\[ t = 150 \text{ My}; \ V_{\text{max}} = 6 \text{ cm/yr} \]

Temperature

\[ Q_{\text{sect}} = 11; \ Q_{\text{base}} = 2 \ (\text{mW/m}^2) \ (a) \]

\[ Q_{\text{sect}} = 11; \ Q_{\text{base}} = 14 \ (\text{mW/m}^2) \ (b) \]

\[ Q_{\text{sect}} = 12; \ Q_{\text{base}} = 15 \ (\text{mW/m}^2) \ (c) \]

\[ Q_{\text{sect}} = 12; \ Q_{\text{base}} = 14 \ (\text{mW/m}^2) \ (d) \]

Figure B.40
Composition

$t = 0$ My; $V_{max} = 8$ cm/yr

$z$ (km)

$t = 0$ My; $V_{max} = 8$ cm/yr

$z$ (km)

$\delta \rho / \rho$

$O_{sect} = 11; O_{base} = 2$ (mW/m$^2$) (a)

$Q_{sect} = 11; Q_{base} = 14$ (mW/m$^2$) (b)

$t = 100$ My; $V_{max} = 5$ cm/yr

$z$ (km)

$Q_{sect} = 12; Q_{base} = 15$ (mW/m$^2$) (c)

$t = 150$ My; $V_{max} = 5$ cm/yr

$z$ (km)

$Q_{sect} = 12; Q_{base} = 14$ (mW/m$^2$) (d)

Figure B.43
Composition

Temperature

\[ t = 250 \text{ My;} \quad V_{\text{max}} = 6 \text{ cm/yr} \]

\[ \delta \rho/\rho \]

\[ Q_{\text{test}} = 13; \quad Q_{\text{base}} = 14 \ (\text{mW/m}^2) \]

\[ x \quad (\text{km}) \]

\[ z \quad (\text{km}) \]

\[ t = 500 \text{ My;} \quad V_{\text{max}} = 6 \text{ cm/yr} \]

\[ \delta \rho/\rho \]

\[ Q_{\text{test}} = 14; \quad Q_{\text{base}} = 13 \ (\text{mW/m}^2) \]

\[ x \quad (\text{km}) \]

\[ z \quad (\text{km}) \]

\[ t = 750 \text{ My;} \quad V_{\text{max}} = 6 \text{ cm/yr} \]

\[ \delta \rho/\rho \]

\[ Q_{\text{test}} = 15; \quad Q_{\text{base}} = 13 \ (\text{mW/m}^2) \]

\[ x \quad (\text{km}) \]

\[ z \quad (\text{km}) \]

\[ t = 1000 \text{ My;} \quad V_{\text{max}} = 6 \text{ cm/yr} \]

\[ \delta \rho/\rho \]

\[ Q_{\text{test}} = 16; \quad Q_{\text{base}} = 13 \ (\text{mW/m}^2) \]

\[ x \quad (\text{km}) \]

\[ z \quad (\text{km}) \]

Figure B.44
Figure B.45
Composition

Temperature

Figure B.46
Figure B.47
Composition Temperature

$\delta p / \rho$

$a$

$t = 0$ My; $V_{\text{max}} = 7$ cm/yr

$b$

$t = 50$ My; $V_{\text{max}} = 4$ cm/yr

$c$

$t = 100$ My; $V_{\text{max}} = 5$ cm/yr

$d$

$t = 150$ My; $V_{\text{max}} = 5$ cm/yr

$Q_{\text{ext}} = 11$; $Q_{\text{base}} = 2$ (mW/m²)

$Q_{\text{ext}} = 11$; $Q_{\text{base}} = 6$ (mW/m²)

$Q_{\text{ext}} = 11$; $Q_{\text{base}} = 6$ (mW/m²)

$Q_{\text{ext}} = 11$; $Q_{\text{base}} = 6$ (mW/m²)

Figure B.48
Figure B.49
Composition

\[ t = 0 \text{ My}; \ V_{\text{max}} = 4 \text{ cm/yr} \]

\[ t = 50 \text{ My}; \ V_{\text{max}} = 2 \text{ cm/yr} \]

\[ t = 100 \text{ My}; \ V_{\text{max}} = 3 \text{ cm/yr} \]

\[ t = 150 \text{ My}; \ V_{\text{max}} = 3 \text{ cm/yr} \]

Temperature

\[ Q_{\text{ext}} = 11; \ Q_{\text{base}} = 1 \text{ (mW/m}^2) \]

\[ Q_{\text{ext}} = 11; \ Q_{\text{base}} = 6 \text{ (mW/m}^2) \]

\[ Q_{\text{ext}} = 11; \ Q_{\text{base}} = 6 \text{ (mW/m}^2) \]

\[ Q_{\text{ext}} = 11; \ Q_{\text{base}} = 6 \text{ (mW/m}^2) \]

Figure B.50
Composition

\[ t = 250 \text{ My}; \quad V_{\text{max}} = 3 \text{ cm/yr} \]

\[ t = 500 \text{ My}; \quad V_{\text{max}} = 3 \text{ cm/yr} \]

\[ t = 750 \text{ My}; \quad V_{\text{max}} = 3 \text{ cm/yr} \]

\[ t = 1000 \text{ My}; \quad V_{\text{max}} = 2 \text{ cm/yr} \]

Temperature

\[ Q_{\text{sec}} = 11; \quad Q_{\text{base}} = 4 \text{ (mW/m}^2\text{)} \]

\[ Q_{\text{sec}} = 12; \quad Q_{\text{base}} = 4 \text{ (mW/m}^2\text{)} \]

\[ Q_{\text{sec}} = 12; \quad Q_{\text{base}} = 8 \text{ (mW/m}^2\text{)} \]

Figure B.51
Figure B.52
Figure B.53
Composition

\( t = 250 \text{ My; } V_{\text{max}} = 7 \text{ cm/yr} \)

\( t = 500 \text{ My; } V_{\text{max}} = 6 \text{ cm/yr} \)

\( t = 750 \text{ My; } V_{\text{max}} = 4 \text{ cm/yr} \)

\( t = 1000 \text{ My; } V_{\text{max}} = 13 \text{ cm/yr} \)

Temperature

\( \delta \rho / \rho \)

\( Q_{\text{ext}} = 10; \ Q_{\text{base}} = 9 \ (\text{mW/m}^2) \)

\( Q_{\text{ext}} = 10; \ Q_{\text{base}} = 8 \ (\text{mW/m}^2) \)

\( Q_{\text{ext}} = 10; \ Q_{\text{base}} = 8 \ (\text{mW/m}^2) \)

\( Q_{\text{ext}} = 11; \ Q_{\text{base}} = 7 \ (\text{mW/m}^2) \)

Figure B.54
Figure B.55
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