Chapter 1

Introduction

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1.1 Modeling Mantle Convection With Plates

Mantle convection and associated plate tectonics are principal controls on the thermal and geological evolution of the Earth. These processes are central to our understanding of the origin and evolution of tectonic deformation, the evolution of the thermal and compositional states of the mantle, and ultimately the evolution of the Earth as a whole. Plate creation and motion largely governs the loss of heat from the solid Earth (*Davies*, 1999), and the strength of plates may control the energy dissipation and hence heat loss over geological time (*Conrad and Hager*, 1999b). However, despite the central importance of plate dynamics, there are fundamental uncertainties on the forces resisting and driving plate motions.

Although there is consensus that the 1–10 cm/yr motion of plates is driven largely by the thermal buoyancy within subducted slabs (*Hager and O'Connell*, 1981) and perturbed by upper mantle solid-solid phase transitions (*Billen*, 2008) and cooling of oceanic lithosphere from ridge to trench, the importance of the aseismic extension of slabs within the lower mantle

(*Lithgow-Bertelloni and Richards*, 1998) remains unresolved. The strength of subducted slabs likely regulates the velocity of plate tectonics. This strength is usually described by viscosity, where yielding (defined by a yield stress) and other weakening processes can limit the viscosity locally. The vast majority of available negative buoyancy driving plates is within the transition zone and lower mantle, and if slabs are strong, then this force can be coupled directly into the edges of oceanic plates at trenches (*Zhong and Gurnis*, 1995). However, if the oceanic lithosphere is strong during initial subduction as it bends below the trench, then the dissipation within the narrow high-viscosity slab could limit plate velocity (*Conrad and Hager*, 1999a). Although the importance of plate margin and slab strength has been studied in two- and three-dimensional Cartesian models aimed at understanding the physics of subduction (*Zhong et al.*, 1998; *Cizkova et al.*, 2002; *Schellart*, 2004; *Faccenna et al.*, 2007; *Billen*, 2008) and in limited regional models that assimilate observed structure (*Billen et al.*, 2003; *Jadamec and Billen*, 2010), the incorporation of realistic rheologies into models with narrow slabs and plate boundaries has remained an elusive goal of global geodynamics. Whether slabs are weak or strong remains unresolved; previous numerical modeling efforts used slab and plate viscosities varying from 10^{21} to 10^{26} Pa s (*Billen*, 2008).

With the incorporation of strong slabs and realistic treatment of plate margins, the ability to observationally constrain models would increase substantially. Observations constraining the deformation of slabs will prove useful in global models: examples include the strain rate and state of stress within slabs from deep focus earthquakes (*Isacks and Molnar*, 1971) and the kinematics of slab rollback in subduction zones with present-day back-arc extension (*Dewey*, 1980). Large fractions of the Earth's surface ($\sim$ 15% globally) do not follow a rigid plate tectonic model but undergo deformation close to trenches and further from plate margins (*Gordon and Stein*, 1992). Some oceanic plates are deforming diffusively within their interiors, especially the Indo-Australian plates (*Gordon et al.*, 1998). The rich array of geodetic, topographic, gravitational, and seismic observations from local to regional scales constrains these deformations and could validate global dynamics if we are able to capture the commensurate scales in models with realistic rheologies.

Arguably, the biggest limitation on current progress is not observational, but computational: solution of models that incorporate realistic rheologies and local geological structure has been prohibitive historically, due to limitations in numerical methods and computational resources. Taken as a whole, current generation models poorly exploit the observational constraints on present-day deformation. For example, models of plate motion do not use observations of deformation at plate margins and interiors because numerical simulation of global mantle convection down to the scale of faulted plate boundaries has been intractable due to the wide range of time and length scales involved. Using adaptive mesh refinement (AMR) on highly parallel computers has allowed us to incorporate realistic rheologies and fine-scale observations in global mantle convection simulations, and in turn reach fundamental conclusions on the forces driving the plates and the energy dissipation throughout the solid earth by intimately linking these models to observations.

1.2 Parallel Adaptive Mesh Refinement

Capturing the large viscosity variations occurring at plate boundaries requires a mesh with about 1 km local resolution. A spacing of up to 1 km is needed in order to have a narrow weak shear zone of a few kilometers, as the development of weak shear zones of width Δw during plate failure depends on the mesh spacing Δd: $\Delta w \sim 2\Delta d$ to $4\Delta d$ (*Gurnis et al.*, 2004). Such a small mesh spacing also allows for a smooth geometry in the bending part

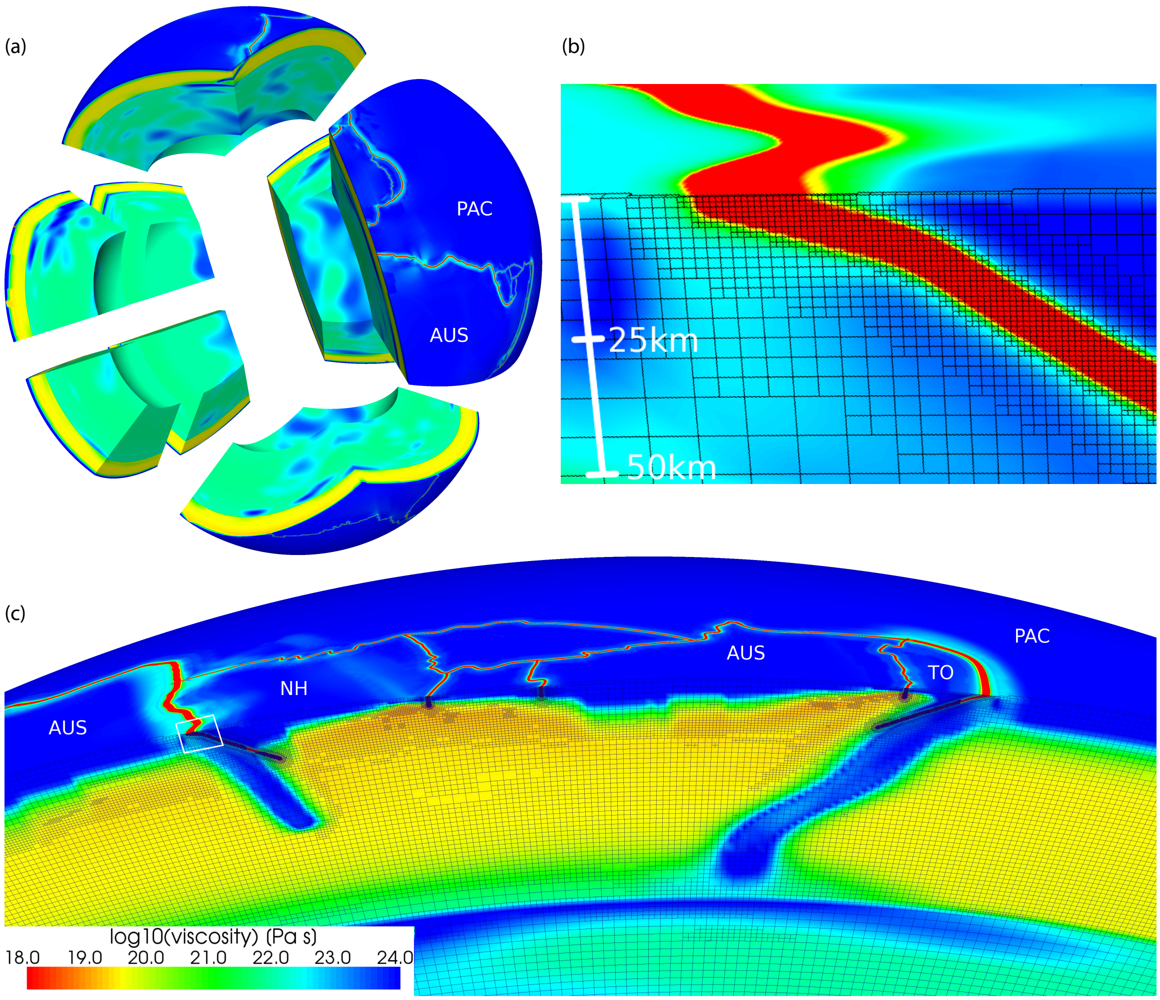

Figure 1.1. (a) Splitting of the Earth's mantle into 24 warped cubes. Each cube is identified with an adaptively subdivided octree whose octants are the mesh elements. The effective viscosity field is shown; the narrow low-viscosity zones corresponding to plate boundaries are seen as red lines on the Earth's surface. (b) Zoom into the hinge zone of the Australian plate (as indicated by the box in (c)) showing the adaptively refined mesh with a finest resolution of about 1 km. (c) Cross-section showing the refinement that occurs both around plate boundaries and dynamically in response to the nonlinear viscosity, with plastic failure in the region from the New Hebrides to Tonga in the southwest Pacific. Plates labeled: Australian (AUS), New Hebrides (NH), Tonga (TO), and Pacific (PAC).

of the subducting plate, and avoids a step pattern in the plate boundary that could affect the deformation pattern within the slab hinge. Furthermore, the gradient in viscosity permitted across an element is limited for reasons of numerical stability, and therefore refinement down to a (sub)kilometer scale is required to resolve the many orders of magnitude viscosity gradient between weak plate boundaries and the adjacent stiff plates. A globally uniform mesh with 1 km spacing would require $\sim 10^{12}$ mesh elements, beyond both the capacity of contemporary supercomputers and the reach of numerical solution methods. With AMR, we achieve this high resolution near plate boundaries, while using a coarser 5 km resolution within thermal boundary layers (including the oceanic lithosphere) and 15–50 km resolution for the rest of the mantle, thereby saving a factor of over 10^3 compared to a uniform mesh. The resulting reduction in problem size to a few hundred million elements is critical to making the simulations tractable on petascale supercomputers.

However, scaling AMR to thousands of processors is a challenge (*Diachin et al.*, 2006). Adaptively refined meshes entail irregular and dynamically changing topological mesh relations, while solution on parallel computers makes it necessary to store just a small part of the mesh on each processor. These mesh partitions must be changed after each refinement and coarsening step to ensure that the computational load on individual processors is balanced. Scalable algorithms for these mesh operations as well as numerical solution of the mantle flow equations have been developed in the AMR finite element framework `p4est` (*Burstedde et al.*, 2008a, 2009a,b). These parallel AMR algorithms are based on a forest of adaptive octrees, in which multiple warped cubes are joined to represent general geometries. The spherical shell used to represent the mantle is composed of 24 cubes (Figure 1.1). Each cube is adaptively subdivided using an octree data structure, which allows for fast algorithms to manage the mesh adaptivity and to construct the mesh connectivity information required in numerical simulations. These algorithms, which have scaled to over 200,000 processor cores, are applicable to a broad spectrum of multiscale scientific and engineering problems that require high resolution in localized (possibly dynamically evolving) regions, such as near fronts, discontinuities, material interfaces, reentrant corners, and boundary and interior layers.

1.3 Thesis Overview

In this thesis, the described developments in parallel AMR techniques are used in dynamic global mantle convection computations with the application code `Rhea` to address the following questions:

1. **What is the rheology of plates, slabs, and the surrounding mantle?** How strong are plates and slabs? How weak are plate boundaries? How much weakening occurs in plates that are bending into subduction zones?

2. **What is the coupling between plates, slabs, and the surrounding mantle?** Are slabs able to act as stress guides, i.e., are stresses fully transmitted between plates and slabs? How much stress is transmitted to the surrounding mantle? Where does dissipation of energy take place?

3. **What is the importance of lower mantle structure for plate tectonics?** Does the presence of lateral viscosity variations in the lower mantle affect plate motions? Do cold, high-viscosity structures attached to slabs speed up or slow down subduction?

4. **Can we predict regional dynamics in global convection models?** Can we model the various observed behaviors of trenches, ranging between retreating (rolling back), stationary, and advancing? Can we correctly predict microplate motions? What are the

characteristics of flow around subducting slabs, what is the relative importance of trench-perpendicular and trench-parallel flow?

5. **What is the effect of regional geometry versus rheology on modeled quantities?** Are velocities, viscosity, strain rate, and state of stress in slabs governed primarily by the implemented rheology law, or by the morphology of plate boundaries and slabs, or both?

6. **What caused rapid change in plate motions in the geologic past (i.e., plate reorganizations)?** Could a change in mantle flow pattern or a change in plate boundary force be responsible; i.e., is this a mantle-driven or a plate-driven process? Can we predict such changes in plate motions with dynamic flow models?

In Chapter 2, we investigate how stress drop determined for large-magnitude deep earthquakes and strain rates in subducting slabs can constrain the strength of these slabs (question 1). We focus on two regions, containing the M_W 8.3 Bolivia and M_W 7.6 Tonga events in 1994, and determine for which yield stresses the minimum stress from stress drop estimates and minimum strain rate estimates from seismicity are satisfied.

Chapter 3 further explores this concept by testing the global models with an ensemble of model constraints, both globally and regionally, while examining a significantly larger model parameter space for the yield stress, stress exponent, and the scaling factor used to convert velocity anomalies in seismic tomography to temperature. The constraints include fit to plate motions, plateness, net rotation, the state of stress and strain rate in plates and slabs, as well as microplate motion and trench rollback. We address questions concerning plate and slab strength, coupling between the various parts of the domain including the effect of lower mantle structure on plate motions, and study slab dynamics and trench rollback in detail (questions

1–5).

We then move from the present-day Earth to its past. In Chapter 4, we address the causes and effects of the plate reorganization occurring around 50 Ma (question 6). To this end, we compute instantaneous models just before and after the reorganization using reconstructed plate boundaries and age grids, and study changes in velocities, state of stress, and strain rates at the surface as well as in the interior of the mantle. Reconstructed plate velocities form the major constraint on these models. The rheology determined in the present-day models presented in the previous chapters are used. This work is still ongoing, and is a close collaboration with Kara Matthews and the EarthByte group at the University of Sydney. The Sydney group is primarily responsible for the plate reconstructions, whereas the setup and computation of the dynamic flow models as well as their analysis is my contribution.

In Chapter 5, the findings in this thesis are summarized and put into context of our current understanding of mantle dynamics and plate tectonics, revisiting the set of questions posed in this chapter, and possible future directions in this work are presented.

The computational methods used in `Rhea` and its library `p4est` are described in detail in the Appendix, along with a suite of code benchmarks. This chapter is the result of close collaboration with Georg Stadler at the Institute for Computational Engineering and Sciences (ICES), University of Texas in Austin, and Carsten Burstedde (previously at ICES, now at Universität Bonn), who have been the principal driving force behind the development of `p4est` and `Rhea`. My responsibility in this joint effort has been designing, setting up, running, and analyzing the majority of the benchmark experiments, with the exception of the manufactured solution and the mid-oceanic ridge example.