Mantle Circulation Models: Constraining Mantle Dynamics, Testing Plate Motion History and Calculating Dynamic Topography

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Abstract

Mantle circulation models are a modified class of mantle convection simulations assimilating recent plate motions as the surface velocity boundary condition. In this thesis, I present a suite of mantle circulation models assimilating the past 300 million years of tectonic history. By comparing model predictions of present day mantle temperature anomalies to mantle structure imaged by seismic tomography one can better understand the physical properties of Earth’s mantle. Given a mantle model with realistic physical properties, plate reconstructions can also be tested.

Mantle viscosity is the most significant property affecting mantle circulation models. For subducted slabs to sink to depths predicted by tomography studies a lower mantle viscosity increase of around thirty times is required. For models with a factor of ten increase slabs do not remain at mid-mantle depths for long enough, while a factor of one hundred increase causes slab sinking rates too slow to match imaged tomographic anomalies. An endothermic phase changes could potentially layer mantle convection into two independent layers. In models assimilating plate motions, no model containing an endothermic phase change reaches a fully layered state, even with unrealistically large, negative Clapeyron slopes. The onset of plate tectonics could potentially break down a two-layered mantle into a partially layered state, similar to the present day mantle.

Predictions of mantle heterogeneity from high-resolution, global mantle circulation
Abstract

models match well with complex mantle structure imaged by seismic tomography in the Tethys region. These models indicate that a more complicated history of subduction during the closure of the Neotethys Ocean is required to match the imaged mantle structure. Subduction is required in two locations, one at the Eurasian margin and a second behind a back-arc ocean opening in the Neotethys Ocean. Simultaneous subduction at both plate boundaries appears not to be necessary.

Global mantle circulation models estimate long-wavelength dynamic topography with amplitudes of up to five kilometres. The largest amplitude signal of dynamic topography is at plate boundaries, suggesting that near surface density variations in the mantle contribute significantly to the dynamic topography signal. The five-kilometre amplitude of topography is larger than predicted elsewhere and is explained by the inclusion of near surface density variations, commonly ignored by other global calculations of dynamic topography. If dynamic topography is defined as ‘any topography arising from flow within Earth’s mantle’ then near surface density variations are significant to the dynamic signal.

Predictions of dynamic topography from mantle circulation models reveal a dichotomy between continental and oceanic regions. Oceanic crust is a part of the mantle convection system and so predicted topography for ocean regions matches well with the expected depth versus age curve for oceanic crust. Continental regions are significantly subsided relative to oceans in the dynamic signal, suggesting that isostatic effects mask continental dynamic topography. When predictions of dynamic topography are corrected for isostatic effects and crustal thickness, an accurate estimate of Earth’s observed topography is generated. This work contributes to an on going debate on the nature of dynamic topography on global and regional scales.
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Chapter 1

Dynamic Earth

1.1 Earth

Earth is a terrestrial or rocky planet, the third from the sun in our solar system. Early in the history of the solar system planetesimals formed from a disc of dust and gas around the sun. Eventually these planetesimals collided to form larger bodies of molten metals and silicates. One of these bodies, Earth, formed 4.5 billion years ago and since then has slowly evolved to the Earth we know today. Soon after the initial formation of Earth dense metals such as liquid iron and nickel began to percolate through the silicate and form a dense metal core, a part of which, the outer core, remains liquid today. As Earth cooled the molten silicate solidified to form the mantle, which makes up around 80% of Earth by volume. The final layer is a thin crust of rock at the surface. Evidence suggests that the earliest surviving crust was formed around 3.9 billion years ago. Today, new crust is almost continuously produced at mid-ocean ridges. The crust and the uppermost mantle together make up a distinct mechanical layer called the lithosphere. It is cool and therefore strong and rigid. The lithosphere is divided into a number of individual plates which can move across the surface of Earth as the mantle below them is considerably less viscous. The
motion of these plates is considered in the theory of plate tectonics. The structure of Earth is shown in figure 1.1.

Figure 1.1: A cutaway of Earth’s interior showing the major seismic and mechanical layers and boundaries. Not to scale.

Little of Earth’s structure can be studied by direct observation, so a variety of techniques have been developed to increase human understanding of our planet. In this thesis I study Earth’s mantle by combining a number of those techniques. Numerical models are used to simulate convection, geological observations in the form of tectonic reconstructions are used as a boundary condition for the simulations and seismic tomography is a probe to understand the present day mantle structure. This thesis combines these geophysical techniques to investigate the interaction between surface and deep Earth processes. I start with a look at some of the established theory of Earth structure.
1.1.1 Earth’s Core

Seismological evidence shows that the predominately iron core is split into two individual layers. A liquid outer core (discovered by Oldham 1906), characterised in seismology by having a shear wave velocity ($V_s$) of zero, and a solid inner core (discovered by Lehmann 1936). The two layers combined account for 16% of Earth’s total volume and around 30% of its mass. The core is mostly iron with some nickel. The estimated density of the core is around 10% less than laboratory experiments suggest, indicating the presence of lighter elements such as sulphur, silicon or oxygen (Boehler 2000). Being liquid the outer core convects vigourously and is the source of the magnetic field observed at Earth’s surface. The inner core is solid and is notoriously difficult to image seismically, its age of formation is uncertain and widely debated. However, it is thought to form from the freezing of outer core material depleting it in light elements. This freezing is thought to be a driver for convection in the outer core (Fearn et al. 1981). The core-mantle boundary (CMB) at 3480km from the centre of Earth is one of the most dramatic changes on Earth owing to huge contrasts in viscosity and density.

1.1.2 Earth’s Mantle

The mantle is by far the largest part of Earth. Extending from just a few kilometres below the surface to the core mantle boundary 2900km below. It makes up two thirds of Earth’s mass and over 80% of its volume. Like the core it is impossible to observe directly. Present day knowledge of mantle structure and composition is primarily derived from methods including seismology, laboratory mineral physics experiments, geochemistry, geodesy and numerical models. The mantle is composed of magnesium silicates, which behave differently depending on the time scale. At the scale of human observation the mantle is solid, but over millions of years the silicates deform by a process known as creep
(e.g. Karato & Wu 1993). High temperature and pressure in the mantle allow point defects in the crystal structure to migrate across the lattice. Creep results in motions of a few tens of millimeters per year within the mantle. Over the 4.5 billion year history of Earth this mechanism provides an efficient way of transporting heat from the hot interior to the surface and allows numerical modellers to treat the mantle as a highly viscous, weakly convecting fluid. The convection within Earth’s mantle is estimated to have a Rayleigh number of around $10^9$ (Davies 1999). The slow convection of the mantle establishes itself into a number of cells, the observable part of which is plate tectonic cycles at the surface.

Earth’s mantle is not as simple as a single uniform shell. Analysis of seismic waves travelling through the mantle confirm variations in the density of mantle minerals both laterally and radially. Lateral variations exist due to recycling of crustal material back into the mantle as part of a convection cell, while radial variations arise from the alterations to the structure of minerals due to increasing temperature and pressure. Changes in mineral structure are also shown by high pressure and temperature laboratory experiments and may be responsible for variations in some physical properties such as seismic velocity. Changes in mineral structure also cause regions of the mantle to be anisotropic, this is particularly evident in the travel times of differently polarised seismic shear waves. The mantle is also mechanically layered. The uppermost mantle is strong and rigid and together with the crust forms the lithosphere. Below the lithosphere is a weaker, more mobile layer named the asthenosphere. Between 410 and 660km depth is where a number of changes in the mineral structure occurs. The region, called the transition zone is the boundary between the upper and lower mantle. The lower mantle is traditionally considered to be relatively homogeneous, however further studies are revealing more about the region. Chemically distinct reservoirs (e.g. Kellogg et al. 1999) and a further phase transition near the core mantle boundary (e.g. Hirose et al. 2007) may add to the complexity of the lower mantle.
Seismic waves travelling through Earth’s mantle are sensitive to the elastic properties of the medium, therefore waves propagate at different velocity through material of different density and/or temperature. A process known as seismic tomography can invert a large set of seismic data to produce a three dimensional image of seismic velocity for the entire mantle. From the tomographic image vast amounts can be learned about the heterogeneous mantle. Seismic tomography images a variety of convective behaviours in the mantle from cylindrical, plume like up-wellings to subducting slabs of lithosphere, many of which penetrate into the lower mantle. Numerical models of convection often attempt to model the features observed in seismic tomography. These models provide an excellent test of hypothesised convective behaviour and can be adapted to simulate a variety of phenomena including chemical heterogeneity, plate tectonics or transitions in mineral structure. Combining many geophysical techniques is important to understand mantle convection and answer key scientific questions.

1.1.3 Earth’s Crust

Despite being considerably less than 1% of Earth’s total mass, the crust is easily the most studied and best understood part of Earth. This is mostly due to the fact that it can be studied directly. Samples are easily accessible, unlike the core or mantle. From direct studies of the crust a wealth of information about the mantle can be obtained as the crust is the product of mantle differentiation. The crust contains a considerable amount of incompatible elements from the mantle as they, by definition, tend to partition into the melt component. The melt is naturally buoyant, so rises to the surface and eventually solidifies to form crust. Crustal material comes in two very different forms. The first is continental crust which has an average thickness of around 30 km, but is as thick as 70km in places. The oldest crustal material is found in cratons at the heart of continents and has had many
hundreds of millions of years exposed to a host of geological processes. A combination of sedimentary, metamorphic and igneous processes has left the continental crust an extremely heterogeneous part of Earth. The second type is oceanic crust, which is continuously created at mid ocean ridges. Oceanic crust is mostly formed from the partial melt of primitive mantle at relatively shallow depths, producing a reasonably uniform composition of basalt. Unlike continental crust, oceanic crust is all young, the oldest on Earth at present is around 180 million years old. This is due to the fact that oceanic crust and lithosphere becomes colder and therefore denser as it ages. Eventually oceanic lithosphere becomes dense enough to be recycled into the underlying mantle in a process known as subduction, a fundamental part of plate tectonic theory.

1.1.4 Plate Tectonics

Some of the most dramatic and destructive natural phenomena on Earth are earthquakes and volcanoes. Since such a large number of people live in areas affected by these phenomena man has sought to understand them for a long time. Today we know that these features predominantly occur at the boundaries between rigid lithospheric plates and are caused by the interactions of the plates as they move around the surface of Earth. It was the late 1960’s before plate tectonic theory was accepted by the scientific community, however the theory itself has roots in work undertaken in the early 20th century. Early geologists noticed correlations in stratigraphy and palaeontological records across oceans (the most obvious being Western Africa and Eastern South America, where even the geometry of the continents seem to fit together). This led to Alfred Wegener proposing that continents were able to ‘drift’ across Earth’s surface (Wegener 1912). The continental drift hypothesis was largely ignored by the scientific community at the time as no one could provide an explanation as to why continents would be able to plough through something as solid and dense as
oceanic crust. The preferred explanation for matching stratigraphy was land bridges linking various continents that had subsequently sunk beneath the ocean. Further studies of crustal rocks showed huge contrasts in composition, thickness and density between oceanic and continental crust. The bimodal distribution led to a rethink of the ideas surrounding how continents could move and in the 1960’s sea floor spreading (Hess 1962, Wilson 1963) was proposed as a mechanism for producing new crust and moving continents across the surface.

Rather than just continents moving across the surface it is now known that the surface is divided into lithospheric plates made up of both continental and oceanic crust as well as the uppermost mantle. At present day Earth is made up of seven large plates (Africa, Antarctic, Eurasia, India, North America, Pacific and South America) and numerous smaller plates (such as Cocos, Arabian and Nazca) separated by narrow weak boundaries. In Figure 1.2 the present day configuration of the plates is shown along with the boundaries between them. There are three types of plate boundary. They are:

1. **Convergent**: Plates are moving towards each other. Commonly, this occurs between continental crust and oceanic crust; if this is the case, the cool, dense oceanic lithosphere is forced beneath the buoyant continental lithosphere in a process known as subduction. This is the primary mechanism for recycling lithosphere into the mantle, and is very important for mantle circulation studies. The dense slabs of dense oceanic lithosphere begin to sink into the mantle under gravity, exerting a pull on the rest of the oceanic plate, this ‘slab pull’ force is the major component of the plate tectonic driving force. Lithosphere recycled into the mantle at an ocean-continent convergent margin forms the down going part of a mantle convection cell. Once the majority of oceanic lithosphere is subducted, then a collision may occur between two continents resulting in the creation of vast mountain ranges, such as the Himalayas or Alps. A
Figure 1.2: Earth’s tectonic plates showing trenches (convergent boundaries), signified by triangles in the direction of the trench, mid-ocean ridges (divergent boundaries), shown by double lines, and transform faults (conservative boundaries), shown by single lines. From Fowler (2005)

A present day example of ocean-continent collision would be the Pacific Ocean subducting beneath South America, resulting in a chain of volcanic mountains and severe earthquake activity.

2. **Divergent:** Plates are moving away from each other. This is most common at mid-ocean ridges (MORs), where hot and partially molten mantle rises to the surface creating new oceanic crust as the plates move apart. This young, hot ocean crust cools and slowly subsides as it moves away from the ridge, generating the ridge push...
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force, another contributor to the plate tectonic driving force. The cooling material also records the orientation of Earth’s magnetic field at the time of its creation, thus recording reversals in the magnetic field and giving us a mechanism to calculate the age of the crust and rate of spreading. The upwelling of mantle at mid ocean ridges is the upward component of the mantle convection cell, however upwelling at MORs tends to be passive and restricted to relatively shallow mantle. Occasionally extensional regimes occur on land in rift valleys such as the East African Rift, the location of a future ocean

3. **Conservative:** Plates are moving past one another with no creation or destruction of lithosphere. These boundaries are a common location of earthquakes, there is little or no volcanism. The most well known is the San Andreas fault in California, where several large earthquakes have damaged major centres of population. These boundaries have a less significant effect on mantle dynamics than the previous two plate boundary types.

Volcanic activity also occurs in intraplate locations known as hotspots. Hotspots are generally considered to be the result of mantle plumes and are the more active component of upward flow. There are many theories on the sources of mantle plumes and the topic is still heavily debated. Theories on the initiation of plumes range from sources at the CMB, the mantle transition zone or anywhere in between. They may have thermal or chemical origins or a combination of both. Modelling work completed with the code used in this study has been shown to produce a variety of plume like behaviours similar to those thought to exist on Earth (Davies & Davies 2009).

It is clear that the process of constant production and destruction of oceanic crust (figure 1.3) has huge impacts on the surface of Earth, not only for natural hazards but also
Figure 1.3: A sketch showing the life cycle of oceanic lithosphere, from being created at a divergent boundary to being recycled back into the mantle at convergent boundary. This process is the surface expression of a single convection cell within Earth’s mantle and is the fundamental link between the two distinct realms of Earth. The figure also demonstrates the forces driving plate tectonics. From Mussett & Khan (2000). Forces labelled on the sketch include: RP - Ridge Push, BD - Basal Drag, SP - Slab Pull and SR - Slab Rollback.

mineral deposits and oil reserves. The cycle also affects the interior of Earth, the recycling of the crust at the surface is now thought of as the surface expression of mantle convection and each process is fundamentally linked to the other. Understanding the link between the surface and the deep Earth is a key goal of geophysics. Combining numerical models of convection within the Earth with observations and predictions of surface motion from plate tectonics is one valuable method for investigating the coupling between the two realms of our planet.
1.2 Motivation and Overview

As the only part of Earth that scientists can directly study is the crust, geophysicists must employ a variety of indirect methods to gain understanding of Earth’s interior. Seismology is a powerful tool used to understand Earth’s overall radial structure, but can also be used to unravel smaller, lateral variations within the mantle, a process known as seismic tomography. Geochemical analysis of igneous rocks is used to understand the composition of the mantle from which they are sourced. High pressure and temperature laboratory experiments provide information on the properties of mantle minerals such as their predicted seismic velocities at given mantle conditions. The primary method used in the work presented in this thesis will be numerical modelling. As on vast time scales Earth’s solid mantle deforms through creep and diffusion the laws of fluid dynamics can be applied to understand its motion. The easiest way to do this is to use computer based numerical models to solve the equations of fluid dynamics. Computational fluid dynamics has changed considerably since the early work in the field (McKenzie et al. 1973, 1974, Turcotte & Oxburgh 1967). Since early models a number of computer codes have arrived advancing the field into two (Conman King et al. (1990)) then three dimensions (Citcom Moresi & Solomatov (1995), Zhong et al. (2000)). This progression also saw advances from Cartesian to spherical geometry. The code used to perform calculations presented in this work, TERRA (Baumgardner 1985, Bunge et al. 1996, 1997, Davies 2008, Yang & Baumgardner 2000) is one such three dimensional, spherical code. TERRA is a three dimensional, spherical mantle dynamics code, which was first developed in the early 1980's. Since its creation many users have continued to contribute, meaning that today TERRA is a well established, efficient code capable of modelling the whole mantle. With increasing computing resources it is now possible to scale TERRA such that the simulations are at a high convective vigour similar to that of Earth. More
recently a community of developers has been established with the goal of combining a num-
ber of users adaptations of the code into a trunk version. The result of this is a number of
important developments, such as a much improved method for handling lateral variations
in viscosity (incorporating temperature dependent viscosity was a traditional weakness of
TERRA), using both passive and active particle and variable resolution grids to decrease run
times. Much of the new functionality will be used in simulations presented in this thesis.
TERRA is described in more detail, including the grid and the mathematics involved in
Chapter 2.2

TERRA can also be modified to incorporate the constraints of plate tectonic recon-
structions as the surface velocity boundary condition of the simulation, effectively assim-
ilating plate tectonic data into the convection model. The simulations, known as mantle
circulation models provide an important link between surface processes and deep Earth pro-
cesses. Throughout this thesis I will use mantle circulation models to understand how the
behaviour of mantle convection is affected by assimilating surface plate motions. This study
has three key objectives:

1. To present a comprehensive suite of models using 300 million years of plate motion
history as the surface velocity boundary condition of a mantle convection model. In-
vestigating a variety of properties including viscosity and phase transitions and how
they affect the nature of mantle convection has never been undertaken at this level of
detail and incorporating such a long plate motion history. Such an investigation should
provide new information on the physical properties of Earth’s mantle.

2. To use high resolution mantle circulation models to study the detailed seismic structure
observed in the mid-mantle below India and south Asia. This will include using tracer
particles to link mantle density anomalies to surface processes.
3. To investigate the temporal variation of surface model predictions such as heat flow and dynamic topography, the effect of mantle flow on vertical motions of Earth’s surface. This will focus particularly on calculating and defining dynamic topography and how it is possible to test predictions made by mantle circulation models.

This thesis aims to show how varying important physical properties of the mantle affects the nature of convection in simulations including plate tectonics as a surface velocity boundary condition. I will consider both radial and lateral variations in viscosity, endothermic and exothermic phase transitions, heating mode and the effect of compressibility. Mantle circulation model predictions are compared to a present day snapshot of mantle structure from seismic tomography to assess the viability of the model and gain vital information on which parameters best resemble mantle properties. In certain regions, high-resolution seismic tomography reveal complex mantle structure often thought to be caused by equally complicated surface tectonics. I will test how scaling up simulations to near Earth-like Rayleigh number and assimilating plate tectonic information significantly improves the match between mantle circulation models predictions and imaged structure from high-resolution seismic tomography studies.

One area of currently active research in geodynamics is the study of topography supported by convection within Earth’s mantle, known as dynamic topography. Currently there is a lot of debate on the nature of dynamic topography and how to extract the signal from Earth’s observed topography. Approaches range from localised field studies to global estimates derived from mantle density structure. There is little consensus between many of the different approaches as to what should be included in estimates of dynamic topography or the magnitude of topography which is dynamically supported. This motivates the final chapter of this thesis which looks at how dynamic topography can be estimated from mantle circulation models and what it contributes to the dynamic topography debate.
In this study the 300 million year plate motion history is taken from the work of researchers from the University of Lausanne. It is a refinement of the plate motion history published in Stampfli & Borel (2002, 2004) and more recently in the Ph.D. thesis of Cyril Hochard (Hochard 2008). The 300 million year plate motion history dates from late 2010 and is referred to throughout this thesis as UNIL (2009). Specific aspects of the model focused on the Mediterranean and Tethyan regions can be found in Ferrari et al. (2008) and Stampfli & Hochard (2009). A general overview of plate reconstructions is given in the methods chapter (chapter 2) and specific details related to the closure of the Tethys Ocean is given in chapter 4.

1.3 Thesis Structure

This thesis is comprised of 6 chapters. This chapter has provided some insight into the motivation and background for studying Earth’s mantle via numerical modelling. It is followed by a detailed account of the modelling process general to this thesis. In Chapter 2 I include a description of the modelling code, TERRA, used throughout the thesis, discussion of the surface velocity boundary condition provided by the University of Lausanne and validation of mantle circulation models using seismic tomography. Chapter 3 is an exploration of some of the key parameters affecting mantle convection and how they influence mantle circulation models. In this chapter I discuss both radial and lateral viscosity variations, the effect of exothermic and endothermic phase transitions, and the influence of heating modes and compressibility. Chapter 4 contains a detailed look at the Tethys region from a high resolution mantle circulation model. This chapter is motivated by interpretations of detailed seismic structure in the region which have suggested a two stage closure of the Tethys ocean, I look at seismic tomography in the region and use mantle circulation models
including tracer particles to test interpretations made using seismic tomography. Chapter 5 takes mantle circulation models and looks at the surface effects of mantle convection including the mantle component of heat flow and dynamic topography. I discuss variations in heat flow throughout the model time as well as calculating dynamic topography at present day and how it can be useful to understand variations in dynamic topography through time. Chapter 6 contains a summary of the findings of this thesis.
Chapter 2

Methodology

2.1 An introduction to mantle circulation modelling

Mantle circulation models are a relatively new family of mantle convection models that are conditioned by recent plate motion history (e.g. Bunge et al. 1998). By incorporating the motions of the tectonic plates throughout geological time as the surface velocity boundary condition of a mantle convection model a wealth of information on the properties of Earth’s mantle, the nature of mantle convection and plate motion history is obtained. Comparing the present day mantle circulation model predictions to seismic tomography (e.g. Li et al. 2008) produces a powerful tool for (a) testing published plate motion histories and (b) refining palaeo-geographies (Bunge & Grand 2000).

Mantle circulation modelling is a multi-step process, involving a phase of traditional mantle convection modelling to achieve a suitable starting point for the circulation phase. Developing a suitable initial condition is crucial to a mantle circulation model to ensure a stable start to the circulation phase of the model. As the model requires a number of steps prior to assimilating plate motions I will here present a brief overview of the modelling method. The technical aspects of the various stages are discussed later in the chapter,
including details on: (a) the modelling code, TERRA, (b) the work done to produce the
surface velocity boundary condition, (c) validating the model using seismic tomography. Dis-

cussion of specific methods, such as those for calculating heat flow and dynamic topography
are included in the relevant chapter.

Each model starts with an initial temperature field hardwired into the code. There
are a number of available options, the two used most commonly are shown in figure 2.1.
The perturbations in the initial temperature field are required so convection can begin and
the simulation can progress. The first initial temperature fields shown are derived from
broad spherical harmonic patterns (figure 2.1a,c), the second is from small scale random
perturbations in the temperature field (figure 2.1b,d). In each case the planform pattern
is consistent with depth though the magnitude of anomaly increases with depth. In the
spherical harmonic case the increase is considerably larger. Laterally the pattern of the
initial temperature distribution is consistent on each of the radial shells resulting in an initial
column like distribution of temperature anomalies. Soon after the initiation of convection
this pattern breaks down. It is necessary to ensure sufficient time for the model to progress
away from it’s numerical initiation.

The first phase of modelling is used to progress the model away from the mathem-
atical initiation (demonstrated in figure 2.1) to a distribution of temperature with up-welling
and down-welling flows. From the chosen initial condition mantle convection is simulated
with no prescribed surface velocity boundary condition. The boundary is free to deform
in the two horizontal directions but not in the radial direction, this boundary condition is
known as a free slip boundary condition. As the methodology is to forward model from
a time in Earth’s history towards present day, the exact state of the mantle at the end of
this phase is unknown and so there are a number of possible points from which to start
the forward modelling phase. In this thesis the majority of simulations will be run from
Figure 2.1: Planform plots of the initial temperature anomaly fields (in K) included in TERRA (a) 45 km depth with spherical harmonic initiation (b) 45 km depth with small scale random temperature perturbations (c) 1300 km depth with spherical harmonic initiation (d) 1300 km depth with small scale random temperature perturbations. Note the different scales.

A state where the heat input from the core (basal heating) and heat generation from radioactivity (internal heating) roughly balance the amount of heat exiting the model at the surface, known as thermal quasi-steady state. It is also possible to start the model from a point where the mantle volume is still cooling. It is impossible to accurately understand the thermal state of the mantle 300 million years ago, and so choosing any starting point brings in a number of uncertainties such as the amount of cooling involved.

To illustrate the change in the modelled temperature fields during the initial free slip phase of modelling I include two sets of planform maps of modelled temperature anomalies at stages throughout the simulation. Figures 2.2 and 2.3 illustrate the evolving
temperature field during the initial phase of modelling for a simple set of mantle parameters at near surface (45 km) and mid-mantle (1300 km) depths for each of the commonly used starting conditions.

The plots present temperature anomalies (the absolute modelled temperature with the mean temperature for that particular depth subtracted) to easily distinguish between hot up-wellings and cold down-wellings. During the early evolution of the model the symmetry of the spherical harmonic initiation is retained, however by the time quasi-steady state is reached the temperature fields from both initiations are quite similar. After 7500 time steps (Figures 2.2e,f) a generally non-symmetrical distribution of small hot and cold anomalies is observed. By 10000 time steps (Figures 2.2g,h) both cases have progressed far enough from their initial temperature condition to be considered a suitable starting point for the circulation phase of modelling. Note that the length of time in a single time step varies depending upon the state of the convection calculation. After 10000 time steps approximately half a billion years have passed once corrected from model time to 'Earth time'.

At the near surface (Figure 2.2) the magnitude of temperature anomalies is greater than in the mid mantle (Figure 2.3). Also at the near surface, the models have developed into a network of linear down-wellings (blue colours) with more cylindrical up-wellings (red colours) in between. At mid-mantle depths the vast majority of temperature anomalies for both cases are individual, cylindrical anomalies. In a more Earth-like mantle, one might expect more continuous linear down-wellings similar to subducted slab. These are clearly not present at this point, so therefore further conditioning is required to mimic Earth’s mantle prior to 300 million years before present. After 10000 time steps the similarity of the two cases starting from different initial temperature fields indicates that either can be used to produce sensible mantle circulation models. However, the importance of this modelling phase to remove the symmetry of the initiation is highlighted. For the remainder of this
thesis models are initiated using the small scale, random perturbations.

As well as considering how far the simulation has progressed from its mathematical initial condition, I also consider the thermal state. I look for a roughly quasai-steady thermal state, where the heat generated in the model from basal and internal heating is roughly equal to the heat leaving the model at the surface. Figure 2.4 plots surface heat flux as a function of model time for the small scale random initiation case. During the 10000 time steps illustrated in figure 2.3 the model has a roughly steady heat flux, with a small amount of background cooling. The plot suggests that the model is in a reasonably steady state and confirms this is a suitable point to begin the circulation phase of modelling.

Prior to the forward modelling stage it is important to condition the mantle with some structure related to plate tectonics. It is difficult to know the location of past subducted material other than that some must exist in the mantle, so the model is conditioned with the oldest available plate reconstruction fixed as the surface velocity boundary condition for up to 100 million years Earth time. This serves to generate a distribution of temperature and density anomalies in the shallow to mid-mantle reminiscent of those that would be achieved throughout geological history. Clearly the distribution is unlikely to match the given mantle structure 300 million years before present day, but it proves a more accurate starting point than beginning the circulation phase directly from the initial condition above. Furthermore, conditioning the model with the oldest provided plate reconstruction serves a

\textbf{Figure 2.2 (previous page):} Horizontal maps of temperature anomaly (K) at 45km depth.\par
\textit{Left column: from spherical harmonic initiation, right column from small scale random initiation} (a,b) after 2500 time steps, (c,d) after 5000 time steps, (e,f) after 7500 time steps, (g,h) after 10000 time steps. 10000 time steps is approximately 2.5 billion years model time, 0.5 billion years Earth time.
second purpose. Depending on the vigour of convection the velocity at the surface of the model is likely to be different from the measured velocity on Earth. The model is therefore scaled according to the root mean square (RMS) velocity of Earth’s tectonic plates and those calculated in the model. Scaling is achieved using equation 2.1 Where \( \alpha \) is the scaling factor, and \( V_e \) and \( V_m \) are the RMS velocities for Earth and the model respectively.

\[
\alpha = \frac{V_e}{V_m}
\]  

(2.1)

Scaling the model in this way is important to allow the convection code to achieve a stable solution. Lower Rayleigh number problems are a computationally cheaper way to examine a wide variety of parameters in the model. Due to the nature of lower Rayleigh number convection the velocity of convection is more sluggish. Therefore, applying Earth-like velocities at the surface of a model with lower internal velocities will make achieving a suitable solution to the convection problem tougher. As a result imposed velocities must be slowed to match the natural velocity of the solution prior to assimilating plate motions. To scale the model according to RMS velocity the velocities prescribed as the surface velocity boundary condition are multiplied by the reciprocal of \( \alpha \). To compensate for the reduction in imposed velocity the time each reconstruction is imposed for is increased by the same factor. This

---

**Figure 2.3 (previous page):** Horizontal maps of temperature anomaly (K) at 1300km depth. Left column: from spherical harmonic initiation, right column: from small scale random initiation (a,b) after 2500 convection iterations, (c,d) after 5000 convection iterations, (e,f) after 7500 convection iterations, (g,h) after 10000 convection iterations. 10000 convection iterations is approximately 2.5 billion years model time, 0.5 billion years Earth time
ensures consistency in the amount of material subducted. There is an important distinction between model time, the amount of time completed in the mantle circulation simulation, and Earth time, the amount of geological time this represents. The model time is generally a factor of $\alpha$ greater than actual Earth time which must be taken into account during the simulations.

Although velocities are scaled no other physical parameters are scaled to match. It is possible in cases of lower Rayleigh number convection that factors such as heat diffusion could lead to thicker thermal boundary layers potentially affecting the thickness and strength of slabs. Although the effect of scaling factors such as thermal diffusion will not be considered in this thesis I acknowledge the effect it could have on the models.

Figure 2.5 shows example temperature fields at this stage of the modelling process, again comparing the two different mathematical initiations with the spherical harmonic initiation on the left and the small-scale random initiation in the right column. By the time this point is reached, the difference between models derived from the two individual
starting points in minimal. The near surface temperature planforms are dominated by the
tectonic processes, with colder than average regions in regions of convergence and broad hot
regions near spreading centres. At mid-mantle depths the influence of passive up-wellings is
no longer present, however mantle structure related to down-welling slabs in introduced.

![Planform plots of temperature anomaly after a short phase of conditioning with the 300 Ma tectonic reconstruction.](image)

**Figure 2.5:** Planform plots of temperature anomaly after a short phase of conditioning with the 300 Ma tectonic reconstruction. (a) at 45 km depth originating from spherical harmonic degree initiation (b) at 45 km depth originating from small scale random temperature perturbations (c) at 1300 km depth originating from spherical harmonic degree initiation (d) at 1300 km depth originating from small scale random temperature perturbations Note the different scales.

The final stage of the process is to forward model to present day, scaling the
velocities, as above, if required. For a large number of the models presented in this thesis
the plate motion history is applied in discrete stages corresponding to available plate tectonic
reconstructions. The 300 million year plate motion history is split into 29 individual stages based upon important geological changes and events. Each of the 29 stages has a unique set of plates. The motion of each plate on the surface of a sphere is described by a rotation vector and its magnitude and is applied at the surface of the model for the given amount of time. Once the stage is complete the velocity boundary condition is instantly changed to the one describing the next stage. This method has the advantage of keeping rigid plates with infinitely small boundaries throughout the duration of the model, however, when stages change, plate boundaries can move large distances instantaneously. A second method for applying the boundary condition reduces the sometimes considerable jumps in plate boundary locations. In this method both the location of the rotation pole and its magnitude are linearly interpolated, according to equation 2.2.

\[ \omega_i = \omega_1 (1 - \beta) + \omega_2 \beta \]  

(2.2)

where \(\omega_i\) is the interpolated rotation pole, \(\omega_1\) is the rotation pole of the current stage, \(\omega_2\) is the rotation pole of the following stage and \(\beta\) is the non-dimensionalised time between stages. Equation 2.2 is technically the equation for interpolating the magnitude of the rotation vector, the interpolation of position is given by the same equation with the \(\omega\) replaced with a unit vector. This method smoothenes the transition between plate stages, with the location of the rotation pole moving from location 1 to location 2 along a great circle arc. In this case plate boundaries are not able to jump at the transitions between the stages, resulting in less abrupt changes between reconstructions. In this case plate boundaries do become ill-defined due to the interpolation, with broad regions between each plate rather than sharp boundaries. Each of the methods have positive and negative aspects to them, which are considered later in this thesis. Forward modelling is continued using one of the
methods for assimilating velocity data for the full 300 million years of available plate motion history, until the model reaches present day, where predictions can be compared to images of mantle structure derived from tomography. In simulations performed without smoothing the rotation pole locations during the forward modelling phase jumps between plates can be in the range of hundreds of kilometres. In most cases the jumps are quite small, but larger jumps occur in regions of particularly fast plate velocities such as the closure of the Tethys oceans. Ideally, one would interpolate the location of the plate boundaries rather than the velocities and the rotation pole. In spherical geometry, this is challenging to implement and is not included in this study.

The accuracy of the simulation predictions at present day can be analysed by comparing the predicted temperature field to seismic tomography, a good analogue to present day mantle structure. Comparing predicted temperature fields with seismic velocities is the most common way to validate mantle circulation models (Bunge et al. 1998, 2002), it gives a good first order comparison, which generally illustrates the main similarities and differences between model predictions and seismic tomography. Comparing modelled temperature predictions to seismic velocity however, is not a like for like comparison. Recent developments to TERRA include code to convert the predicted temperature distribution to seismic velocity using mineral physics and thermodynamic equations using the method outlined in Stixrude & Lithgow-Bertelloni (2005, 2011). Converting from temperature to seismic velocity perturbations serves to make a more direct comparison between model predictions and present day mantle structure imaged by seismic tomography. The elastic parameters and density used to calculate seismic velocity are derived from the modelled temperature and pressure fields using the code PerPleX (Connolly 2005). The conversion to seismic velocity is then performed as a post-processing step assuming a pyrolite composition for the mantle. Other end-member compositions are available but since all the mantle circulation models presen-
Chapter 2: Methodology

ted in this thesis are entirely isochemical, pyrolite is a sensible choice. The thermodynamic approach accounts for the non-linear sensitivity of seismic velocity to temperature and pressure. Figure 2.6 demonstrates the sensitivity of density, P-wave velocity and S-wave velocity to pressure in the upper mantle (Cobden et al. 2008) and figure 2.7 shows the sensitivity of the same parameters to temperature variations (Cobden et al. 2009).

In this thesis I use the technology to convert model temperature predictions to seismic velocity perturbations in chapter four. It is important to note that this method for calculating seismic velocity will often overestimate the magnitude of seismic velocity perturbations when compared to seismic tomography (Schuberth et al. 2009a). This highlights some of the issues arising from seismic tomography, particularly with regards to resolution. Processes involved in the tomographic inversion may result in damping of the signal or a weak solution in areas of poor coverage. This perhaps accounts for some of the differences

Figure 2.6: The variation of density, P-wave velocity and S-wave velocity with depth in the mantle for pyrolite, harzburgite and MORB compositions. From (Cobden et al. 2008)
observed between tomographic amplitudes and mantle circulation model predictions. It is possible to use the resolution operator from a specific seismic tomography model to filter predictions from mantle circulation models to the same resolution as seismology. This method attempts to reconcile some of the differences in amplitude observed between the two data sets. I will comment further on this technique in chapter four.
Figure 2.7: The sensitivity of seismic velocity varies with depth (pressure). From (Cobden et al. 2009)
2.2 TERRA, a three-dimensional, spherical mantle convection code

In this thesis the three-dimensional, spherical mantle convection code TERRA (Baumgardner 1985, Bunge et al. 1996, 1997, Yang & Baumgardner 2000) is used to model the mantle at Earth-like vigour. TERRA is a well established and benchmarked code with many applications. TERRA has the capability to include temperature dependent viscosity, track chemical heterogeneity using passive or active particles and to simulate plate tectonics using plate motions as a surface velocity boundary condition. TERRA is now adapted to use 300 million years of plate motion reconstructions provided by researchers at the University of Lausanne, Switzerland (UNIL) as the surface boundary condition as well as the 119 million year history Lithgow-Bertelloni & Richards (1998) previously used in mantle circulation studies (e.g. Bunge et al. 1998, 2002).

TERRA simulates mantle convection by solving the equations for conservation of mass, momentum and energy on a grid of elements created by projecting a regular icosahedron onto a spherical shell. The mesh is continually refined by drawing great circles through the mid point of the triangles formed on the spherical shell by the projection of the icosahedron, this creates four sub-triangles for each of the original triangles as shown in Figure 2.8. Refining the mesh in this manner keeps the grid fairly uniform and avoids a cluster of points at the poles. For simplicity TERRA pairs each of the twenty original triangles to form ten diamonds prior to subdividing. Using this grid of nearly uniformly spaced points allows for larger time steps to be used (Davies 2008).

As TERRA scales well with increasing grid size, the limit to grid refinement is the available computing resources. Each refinement increases the number of grid points by eight times, thus increasing the number of processors required and the amount of computing time
required. Given current resources it is possible to refine the grid so mid-mantle grid spacing is around 50 km in each of the three dimensions. These simulations run on 32 processors over approximately 3 days for the majority of runs presented. Some cases presented in this thesis have a further grid refinement, giving a mid-mantle node spacing of 25 km. These cases require up to 512 processors and run for as long as two weeks.

Figure 2.8: Producing the grid used in TERRA from projecting a regular icosahedron onto a spherical surface. (b) to (d) show successive refinements by drawing great circles through the mid-points of regular triangles. From Oldham & Davies (2004)

Once the shell has been discretised in this manner a series of shells are stacked radially to generate a three dimensional mesh of prisms on which to perform the calculation. TERRA employs a finite element method to solve the governing equations. In simple terms the finite element method reduces a complex continuum problem composed of partial differential equations into a computationally simpler series of simultaneous algebraic equations.
When using a finite element method the three dimensional prisms are the elements, and points where edges of elements meet are termed nodes. Typically in TERRA the elements are triangular prisms and nodes occur where six element edges meet. The finite element method reduces the mathematical problem so that there are a given number of unknowns at each of the nodes. This method considerably reduces the scale of the problem by allowing vector equations to be solved simultaneously at each node. When the domain has been discretised into elements and nodes a shape function is selected to represent a given variable within an element. In TERRA the shape functions are linear. It is then possible to formulate a suite of matrix equations across all elements by combining the value of the field at adjacent nodes with the shape functions. Finally the matrix equations are combined to form a series of simultaneous equations representing the entire domain, at this point boundary conditions are incorporated and the algebraic equations are solved to produce a revised estimate of the field.

The three governing equations of fluid dynamics are the equations for the conservation of linear momentum (2.3), mass (2.4), and energy (2.5). TERRA assumes infinite Prandtl number and models the mantle as a Newtonian fluid using the finite element method. In the most general case the mathematical problem is formulated as three vector equations:

\[
\nabla \cdot \tau - \nabla P + \rho g \hat{r} = 0 \quad (2.3)
\]

\[
\frac{\partial \rho}{\partial t} = -\nabla \cdot (\rho \mathbf{u}) \quad (2.4)
\]

\[
\frac{\partial T}{\partial t} = -\nabla \cdot (T \mathbf{u}) - (\gamma - 1) T \nabla \cdot \mathbf{u} + \frac{\tau : \nabla u + \nabla (k \nabla T) + H}{\rho c_p} \quad (2.5)
\]
In equations (2.3) to (2.5) $\tau$ denotes deviatoric stress (the viscous term in the equation), $P$ is pressure as a function of density and temperature, $\rho$ is density, $g$ is gravitational acceleration, $\hat{r}$ is a unit vector in the radial direction, $u$ is fluid velocity, $T$ is temperature, $\gamma$ is the Grüneisen parameter, $k$ is thermal conductivity, $H$ is radiogenic heat production and $c_p$ is the specific heat at constant pressure. These vector equations describe the local behaviour of the fluid at each node on the grid, which has as many as 80 million points at high resolution, spaced at 25 km laterally in the middle of the model domain. In equation 2.5 the first two terms of the right hand side of the equation represent compressibility in the convection system, occurring when changes in pressure and temperature result in changes in density. Generally I will consider the incompressible approximation of convection, described later in this chapter.

For Earth it is possible to make a number of simplifications to the general method, these simplifications are used to improve the performance of the code but are still justified in the context of real Earth. The first simplification is to subtract a reference density and pressure gradient (2.6) from the conservation of momentum equation (2.3).

$$\nabla P_0 + \rho_0 g \hat{r} = 0$$

(2.6)

to give:

$$\nabla \cdot \tau - \nabla p + \Delta \rho g \hat{r} = 0$$

(2.7)

Subtracting a reference density and pressure gradient removes the lithostatic pressure from the momentum equation and removes the need to work with absolute density, instead using density differences in the calculation. A second simplification is needed as one possible
solution to conservation of mass equation (2.4) gives seismic waves. In mantle circulation modelling this is undesirable as it makes the model time steps exceptionally small, and thus large amounts of computing time are required to generate solutions. On the time scale of mantle convection, seismic waves are not a solution to the numerical problem, so a valuable approximation is to set the time gradient of density to zero, shown in equation 2.8

\[ \frac{\partial \rho}{\partial t} = 0 \]  

(2.8)

substituting this into the conservation of mass equation (2.4) gives:

\[ -\nabla \cdot (\rho \mathbf{u}) = 0 \]  

(2.9)

Equation 2.9 is known as the anelastic approximation. Making the model anelastic prevents seismic waves being returned as a solution to the conservation of mass equation resulting in a more efficient calculation of mantle convection. It is sometimes useful to model the mantle as an incompressible fluid, this eliminates change in volume with changing pressure and temperature. In the incompressible case the divergence of velocity is set to zero and thus a new series of conservation equations (2.10) to (2.12) are derived for the incompressible case.

\[ -\mu \nabla^2 \mathbf{u} - \nabla p + \rho g \hat{r} = 0 \]  

(2.10)

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

(2.11)

\[ \rho c_p \frac{\partial T}{\partial t} = \nabla \cdot (k \nabla T) + H \]  

(2.12)
Where in equation (2.10) $\tau$ is simplified to just the viscous term shown, which requires constant viscosity. Equation (2.11) shows the divergence of velocity is zero and equation (2.12) no longer contains terms for viscous dissipation and adiabatic compression. Finally, it may be useful to choose the Boussinesq approximation (2.13) to the equation of state. This takes $\rho$ as constant in all terms excepting the buoyancy term. This approximation is similar in nature to the incompressible equations, but is not truly incompressible due to the buoyancy term. This saves computing time as the equation of state is vastly simplified, to:

$$\Delta \rho = \rho_0 \alpha \Delta T$$

(2.13)

Where $\alpha$ is the thermal expansivity.

In cases where compressible convection is preferred, TERRA includes a number of versions of the equation of state. For models in this thesis where compressible convection is simulated an anelastic liquid approximation (Jarvis & McKenzie 1980) is used with radial reference values from a Murnaghan equation of state (Murnaghan 1951). These radial reference values are hard-coded into TERRA, and are published in Bunge et al. (1997). Plots of parameter variations with depth are reproduced from that paper in figure 2.9.

As previously mentioned the convection problem is initiated by setting up an initial temperature field within the domain, commonly small scale random perturbations are used at this stage. TERRA uses a Uzawa pressure correction algorithm to solve the initial field for pressure and velocity, from which the rate of change of temperature, $\frac{\partial T}{\partial t}$, is calculated. The rate of change of temperature then provides a new temperature field, so it is now possible to advance in time using a fourth order Runge-Kutta method and repeat the process iteratively.

Further details on the technical aspects of TERRA can be found in these key refer-
2.3 The surface boundary condition

The surface velocity boundary condition is another important part of this study. As I aim to understand the relationship between surface tectonic processes and the deeper Earth, ensuring a detailed and accurate plate motion history to use as the boundary condition is important. This sub-chapter looks at the plate reconstructions used within this study and some of the methods used to develop them.

Present day plate boundaries and motions are reasonably easy to understand. Plate boundaries are identified as areas of high seismicity and volcanic activity and the motions of individual plates can be accurately tracked by satellite based GPS systems. Understanding how plates evolved throughout geological time is nothing like as simple and requires a variety
of geological techniques. In the Cenozoic and the Late Mesozoic (appendix B contains an illustration of the geological time scale) it is possible to use palaeomagnetic evidence recorded in the ocean floor to track plates motions relative to one another, however beyond around 120 million years before present this method is not so useful, due to the paucity of old enough oceanic crust. In order to obtain accurate plate motion histories further back into the Mesozoic and Palaeozoic, models must include other forms of data. The model of Stampfli and colleagues at the University of Lausanne (UNIL) uses not only palaeomagnetism, but also geological relationships and climatology to piece together the motions of plates (Hochard 2008, Stampfli & Borel 2002). Initially plate motions were reconstructed relative to the stationary hotspot reference frame, more recently reconstructions are made in the hybrid reference frame of Torsvik et al. (2008). All of the mantle circulation models presented in this thesis use recent editions of the plate motion history in the global hybrid reference frame. The model, as described in Hochard (2008), pieces together plate motion history in discrete snapshots of time by locating plate boundaries precisely. Many earlier models tend to locate only continental areas, and so are not as useful for driving mantle convection models, as the surface expression of mantle convection is mostly at plate boundaries. Accurately located plate boundaries are more useful to mantle circulation modellers as down-wellings at subduction zones drive the model, and the remnants of ancient subduction zones are recovered in models.

Although the work reconstructing plate motion history was done by others, it is important to present an overview of the methods involved. Commonly plates are reconstructed by tracking the motions of continents, essentially a continental drift approach to the problem. The team at UNIL developed a methodology placing plates and their kinematics at the centre of the issue (Hochard 2008) therefore attempting to describe the motion of entire plates not just the continents.
Initially continents must be defined, continental lithosphere is divided into distinct geological terranes. Each terrane is geologically unique in the region of interest and is described by Howell (1995) as ‘a fault bounded package of rocks of regional extent characterised by a geologic history which differs from that of neighbouring terranes.’ A present day terrane map of the world from Hochard (2008) is presented in figure 2.10. Later updates to the reconstructions used in this study define more terranes and include greater coverage. Terranes, oceanic segments and initial estimates of motions are reconstructed using palaeomagnetism including palaeo-poles and magnetic anomalies, palaeogeography including palaeoclimatology and biology and the large scale geology of the region.

![Figure 2.10: A global map showing the distribution of geological terranes (GDUs) used in the UNIL (2009) plate tectonic reconstruction. This figure was produced by C. Hochard](image)

Using terranes this way still only creates a series of continents. To create a history that captures plate tectonics entirely, continents (formed from a set of terranes) must be combined with oceanic components limited by dynamic boundaries. Some plates contain only oceanic crust, others contain portions of both oceanic and continental crust. Plate
motion history is constructed in discrete time slices based on important geological events. Currently the history used with TERRA has twenty nine individual reconstructions over 300 million years. The reconstructions are not evenly spaced throughout time and occur at 300Ma, 290Ma, 270Ma, 250Ma, 240Ma, 230Ma, 220Ma, 210Ma, 200Ma, 180Ma, 165Ma, 155Ma, 142Ma, 131Ma, 121Ma, 112Ma, 103Ma, 95Ma, 84Ma, 70Ma, 57Ma, 48Ma, 40Ma, 33Ma, 20Ma, 14Ma, 10Ma, 6Ma and 0Ma (present day). The process of creating a history of plate tectonics is iterative and so as more data are added in the past and alterations are made to a reconstruction this affects the surrounding reconstructions too. Plate boundaries are defined by their bounding faults; either mid ocean ridges, subduction zones or transform faults and the plates are always defined as rigid except during continent-continent collision. Since two plates will always have velocities that are converging or diverging and plate boundaries are always conserved as time moves forwards gaps and overlap will occur between plates. In these cases new plate boundaries are formed. Convergent boundaries are formed where plates overlap and divergent boundaries formed in gaps between plates. The reconstructions are refined iteratively from an initial estimate based upon the various types of geological input. A comprehensive account of the methods used in reconstructing plate motion history is presented in Chapter 2 of Cyril Hochard’s Ph.D. thesis (Hochard 2008). Illustrations of the palaeogeography and tectonics of the UNIL (2009) reconstructions are provided in appendix C.

The most recent revision of the UNIL plate motion history incorporated into TERRA uses the updated hybrid reference frame of Torsvik et al. (2008). From 320 million years before present, a hybrid of the palaeomagnetic, African fixed hotspot, African moving hotspot and global moving hotspot reference frames is used. In the last 100 million years the hybrid reference frame uses a moving hotspot reference frame. This takes into account motion of hotspots arising from mantle convection, eliminating uncertainty arising from a
fixed hotspot reference frame derived from hotspot tracks (Torsvik et al. 2008).

In chapter four of this thesis I will look in more detail at one specific region, the subduction history of the Neotethys ocean. This ocean closed between roughly 150 and 50 million years before present as the Indian continent moved northwards before colliding with Asia. Previous mantle circulation models using the plate reconstructions of Lithgow-Bertelloni & Richards (1998) contained only 119 million years of plate motion history in 11 discrete stages. Each stage of these reconstructions has between ten and twelve plates. These reconstructions contain what is now thought to be a relatively simplistic view of the Tethys ocean closure. One considerable advantage of using the longer, more detailed plate motion history described here is the extra detail in the closure of the Tethys oceans from 300 million years to present day. The benefits of longer plate motion history in recovering shorter wavelength mantle features are discussed at length in chapter four.

All plate tectonic reconstructions will have weaknesses, this is due to the inherently subjective nature of generating such reconstructions. Here, I will briefly discuss some of the limitations of the UNIL reconstruction when compared to similar published reconstructions, such as the GPlates reconstructions of University of Sydney. As mentioned above, the UNIL reconstruction is particularly suited to use as the surface boundary condition of a mantle circulation model as it is generated by locating plate boundaries as accurately as possible through geological time. Other reconstructions which reconstruct continents from a collage of terranes do not have the same global coverage that is possible with the UNIL reconstruction. Weaknesses in such a model are generally larger the further back in time the model extends. For the most recent 120 million years it is possible to accurately reconstruct oceanic plates from magnetic anomalies associated with sea floor spreading. Most reconstructions are broadly similar in the period where magnetic anomalies can be reconstructed. Furthermore large palaeo-continents such as Laurentia and cratonic Africa, Australia and Asia can
be reconstructed back in time using published palaeomagnetic data, often resulting in similar interpretations. Differences generally occur in the longitudinal location of these continents.

As a tool palaeomagnetism can give the latitude of a samples formation accurately, however longitude is a weakness. In the UNIL model smaller terranes are pieced together to form larger plates based on information derived from many sources in the published geological literature. This is a similar approach to most other plate tectonic reconstructions and is subject to the human error of the scientists piecing together the model. The UNIL model has undergone nearly thirty years of continuous development and revision and should represent a reasonably accurate model. Prior to the simulations included in this thesis some models were undertaken using the plate reconstructions of Lithgow-Bertelloni and Richards (1997). Results indicated a broadly similar pattern at the largest scale, but the shorter, less detailed history proved weaker at the finer resolutions. These simulations are very similar to early mantle circulation models performed by Bunge and colleagues as referenced throughout this thesis.

The EarthByte tectonic reconstructions (GPlates) have some favourable features, such as the locations of plate boundaries at 1 million year intervals. Locating the plate boundaries in the GPlates plate motion history is based upon magnetic anomalies in the ocean crust. This limits the reconstructed plate boundaries to around 140 million years. During the time frame from 140 Ma to present many reconstructions are similar as the methods rely less on human interpretation and more on observables, like the oceanic magnetic anomalies. The advantage of using the UNIL plate motion history in this study is longer time frame for which reconstructed plate boundaries are available. Obviously this brings some inherent weaknesses as described above but these will be tested to some degree later in the thesis.
2.4 Model validation

The accuracy of the predictions of mantle circulation models can be assessed by comparisons to present day mantle structure imaged by seismic tomography. As a first order comparison matching up seismically fast regions observed in tomography with cold regions in a mantle circulation model provides a good indication of the accuracy of the mantle circulation model. As seismic velocity is sensitive to the bulk modulus of the sampled mantle, seismic waves travelling through more rigid, and by association more dense, mantle tend to travel at higher velocity. Commonly, more dense mantle is thought to be associated with cold material from subduction zones and therefore it is possible to assume that cool mantle in mantle circulation models might correlate with fast mantle in seismic tomography. Converting predicted temperature fields to seismic velocity (as described in chapter 2.1) serves to improve the validity of the comparison.

As the model is driven primarily by surface tectonics the key features observable in the results are likely to be associated with plate boundaries. Hot up-wellings will occur at divergent margins, but are mostly passive features and are constrained to the uppermost mantle, near to the surface. More active up-wellings, such as hot rising plumes, are not as easily affected by surface tectonics and so are not likely to appear in Earth-like locations. The size and quantity of up-wellings may yield some information on how Earth-like a model is. Therefore, the simplest way to understand a mantle circulation model is to look at the prominent down-wellings associated with convergent margins, as they are directly conditioned by the surface velocity boundary condition. Two down-wellings that appear robust in previous mantle circulation studies and numerous tomography models are associated with the subduction of the Tethys oceans and the Farallon ocean. Numerous tomography models (e.g. Grand 2002, Li et al. 2008, Ritsema et al. 2004, Romanowicz 2003) have a faster
than average seismic velocity at mid mantle depths under North America and South Asia interpreted to be the Farallon and Tethys ocean slabs respectively. These slabs are robust, as they appear in most interpretations of seismic data and have frequently been used to help analyse mantle circulation models (Bunge et al. 2003, Bunge & Davies 2001, Bunge & Grand 2000). Figure 2.11 shows the location of the cold anomalies interpreted as the Tethys and Farallon subductions at 1300km depth in the mantle for the body wave tomography model of Li et al. (2008).

![Figure 2.11: P-Wave tomography at 1300km depth, showing the Farallon and Tethys slabs as fast regions. From Li et al. (2008)](image)

In this study using extended plate motion history adds the potential for locating subducted material associated with older orogens in the lower mantle. It is therefore important to identify more robust features in tomography. An issue with tomography is the lack of resolution at greater depths, so it is even more important to ensure any features observed are robust across a number of models. Two potential locations for further investigation including a cold deep mantle anomaly under Siberia proposed to be the remnants of a Jurassic age subduction (van der Voo et al. 1999a) and a lower mantle slab under the
Tasman Sea, in the complex south Pacific (Schellart et al. 2009).

Furthermore, a recent analysis of the tomography model produced by Utrecht University (Amaru 2007) presents a series of approximately 30 fast seismic anomalies which can be related to surface events in the past 300 million years (van der Meer et al. 2010). The analysis includes a more detailed look at the Tethys region and identifies distinct locations based upon smaller events in tectonic history. Reproducing some or all of these ‘slabs’ will prove an excellent test of the mantle circulation model and the plate motion history.

Tomographic images are created by splitting the mantle into a series of blocks, as seismic rays pass through the block their velocity is affected by the material properties of that block. A simple example of this method is shown in figure 2.12.

The seismic waves travelling through the elastically less rigid blocks travel at lower
velocity and thus take longer to arrive at the station. Equations (2.14) and (2.15) relate P and S wave velocities respectively to the physical properties of the mantle.

\[ V_p = \sqrt{\frac{K + \frac{4}{3} \mu}{\rho}} \]  \hspace{1cm} (2.14)

\[ V_s = \sqrt{\frac{\mu}{\rho}} \]  \hspace{1cm} (2.15)

where \( K \) and \( \mu \) are the bulk and shear moduli and \( \rho \) is the density. A complex inversion is undertaken to calculate the velocity of each of the blocks, and deviations from a one dimensional reference model, such as PREM (Dziewonski & Anderson 1981) are calculated. Once deviations from a reference model are calculated, a number of slices are stacked in three dimensions to produce a whole mantle model of velocity perturbations. From the whole mantle velocity perturbations it is possible to plot vertical and horizontal cross sections showing the deviation from the one dimensional model. Most tomography models are presented as a series of cross sections, usually horizontal cross sections relating to a single depth. These images are the basis of the comparison with mantle circulation model predictions.

The resolution of the tomography model is dependent upon the size of the blocks used in the inversion (smaller blocks lead to higher resolution) and the number of waves passing through the block. Accuracy of tomographic images is improved by having a larger number of rays passing through a block. Commonly mantle circulation models are at a higher resolution than tomography, especially deeper in the mantle. Models with larger data sets and smaller blocks are becoming more common, however, in areas of low coverage resolution may still be an issue.
Chapter 3

Understanding mantle material properties using mantle circulation models

3.1 Introduction

There are a vast number of unknown physical properties of the Earth’s mantle, most of which will play an important role in the way the mantle convects. As the mantle cannot be observed by direct observation, other than in very few specific locations, it is challenging for Earth scientists to understand these unknown properties. Instead numerous indirect methods are used to build up an understanding of the convective behaviour of the mantle. A variety of geophysical methods, including seismology, geochemistry, geodesy, laboratory experiments and numerical models present a picture of the evolution of Earth’s mantle. Despite this there is still a lot of uncertainty relating to some of the fundamental physical properties of Earth’s mantle. Numerical models of mantle convection can be used to
explore a wide range of physical properties and test the behaviour of mantle convection given certain conditions. This chapter looks at using mantle circulation models to understand how varying mantle parameters affects mantle flow, particularly in the mid-mantle where mantle circulation models have been shown to generate down-wellings resembling faster than average mantle from seismic tomography (Bunge et al. 2002).

In Chapter 2 I discussed how the method for performing mantle circulation models in TERRA is based upon the original methods used by Bunge and co-authors. A number of other studies use similar methods for calculating geodynamic models and comparing the results to tomographic images. Recently Steinberger et al. (2012) showed the geodynamic models and seismic tomography often provide information on mantle heterogeneity at different scales. They use the comparison to suggest potential improvements to geodynamic models. The modelling method used by Steinberger et al. (2012) differs from the one I use in this thesis. Rather than applying velocities derive from plate motion history as a boundary condition for the model the authors introduce slabs into the upper mantle in regions of tectonic convergence. Although their method differs slightly to the one used in TERRA it indicates the value of using comparisons between numerical models and seismic tomography studies. As well as using mantle circulation models to improve geodynamic models they can also be used to investigate reference frames for plate tectonic models (Shephard et al. 2012). The authors note that the five reference frames tested in the study can produce notable differences between the mantle heterogeneity structures produced. This also indicates the value of the mantle circulation model method and ability of using comparisons between mantle circulation models and seismic tomography to test a number of important phenomena on Earth.

This chapter contains a brief summary of the methods used in mantle circulation modelling and a look at global seismic tomography. In section 3.4 I test a series of mantle
models with depth and temperature dependent viscosity. Section 3.5 considers the effect of both exothermic and endothermic phase transitions on mantle circulation models. Finally, I look at how mantle circulation models behave given a series of other properties and conditions such as heating mode and compressible convection in section 3.6.

3.2 Summary of methods

A detailed account of the methods used throughout this thesis is provided in chapter 2. Here, I will briefly summarise the methods used relevant to this chapter of the thesis. This chapter involves mantle circulation models using the code TERRA (Baumgardner 1985, Bunge et al. 1997, Davies & Davies 2009, Yang & Baumgardner 2000). TERRA solves the principal equations of fluid dynamics for the conservation of mass, momentum and energy. Equations are solved using the finite element method on a three-dimensional spherical grid. The simulations considered in this chapter are undertaken at a resolution known in TERRA as MT = 128. This notation refers to a grid with spacing of approximately 50 km in each direction in the mid-mantle. Due to the spherical grid, surface resolution is coarser while resolution at the core-mantle boundary is finer. Each of the simulations involves a surface velocity boundary condition taken from plate motion reconstructions. The reconstructions are provided by researchers at UNIL and are described in detail in chapter 2.

The modelling process involves three stages. First a standard mantle convection simulation is run using free slip for both top and bottom boundary conditions. This phase is run until the model reaches a thermal quasi-steady state where heat input is roughly equal to heat output. The second step involves conditioning the mantle to resemble mantle conditions 300 million years before present. Of course, these conditions are unknown, therefore for this stage I run the simulation with the oldest available reconstruction (300 Ma) for a short
time, generally 50 million years. This provides an initial condition with a distribution of density and temperature anomalies related to both linear down-welling structures and plume like up-welling structures. Once a satisfactory initial condition is developed the model is run forward from three hundred million years before present applying plate motions as the surface velocity boundary condition. During the 300 million years of forward modelling plate motion is applied in 29 discrete stages based upon the 29 reconstructions which make up the UNIL (2009) plate motion history.

The validity of models is considered by comparing the modelled temperature fields to seismic velocity perturbations obtained from tomography studies. Models are judged on how well the modelled down-welling temperature anomalies match to seismically faster than average mantle.

3.3 Seismic tomography

Bunge et al. (2002) showed that mantle circulation models at MT = 128 (50 km grid spacing) resolution can be used to accurately model two regional fast seismic anomalies, imaged at mid-mantle depths in horizontal slices through tomography. These are located beneath South Asia and North America and are referred to as the Tethys and Farallon anomalies respectively. These two regions of faster than average mantle are robust across the variety of tomography studies considered here. Throughout this chapter I will consider the comparison between the modelled temperature anomalies and seismic velocity perturbations from a variety of tomography models. Since the paper of Bunge et al. (2002) there have been considerable advances in seismic tomography, including making whole mantle tomographic data sets available online. I will look at six tomography studies to confirm that the two regions of faster than average mantle are robust across a number of tomography models.
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The six models considered are; P-mean and S-mean (Becker & Boschi 2002); GyPSuM-p and GyPSuM-s (Simmons et al. 2010); MITP08 (Li et al. 2008) and S40RTS (Ritsema et al. 2011). The studies are chosen to represent a variety of resolutions and seismic data types. Figures 3.1 and 3.2 show global horizontal slices through the selected seismic tomography studies. The velocity perturbations are presented on two scales, the left columns contain the P-wave models on a scale ranging from -0.6% to +0.6%. The right-hand columns contain the S-wave models on a scale from -1.2% to +1.2%. It is fair to say that the two regions of faster than average mantle observed by Bunge et al. (2002) are robust across the six tomography studies presented in figures 3.1 and 3.2. I will therefore compare the modelled temperature anomalies to these seismic tomography models, looking particularly at the two regions of long-lived convergence, Tethys and Farallon.
Figure 3.1: Horizontal cross section of wave speed perturbations from six seismic tomography models at 925 km depth. Models considered are MITP08 (Li et al. 2008), S40RTS (Ritsema et al. 2011), GyPSuM-p & GyPSuM-s (Simmons et al. 2010) and P-mean & S-mean (Becker & Boschi 2002)
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Figure 3.2: Horizontal cross section of wave speed perturbations from six seismic tomography models at 1375 km depth. Models considered are MITP08 (Li et al. 2008), S40RTS (Ritsema et al. 2011), GyPSuM-p & GyPSuM-s (Simmons et al. 2010) and P-mean & S-mean (Becker & Boschi 2002)
3.4 The effect of depth and temperature dependent viscosity upon mantle circulation models

Viscosity is arguably the most fundamental property of Earth's dynamic mantle. It is known that on a geological time scale the mantle can be well represented by a highly viscous, weakly convecting fluid. However, the actual value for viscosity and its variability throughout the mantle are poorly constrained. There are many factors affecting mantle viscosity such as the pressure (depth) and temperature. Factors such as mineral structure, grain size and chemistry can also affect the viscosity of the mantle (Tackley 2012). Furthermore, recent developments such as the observation of transitions in the spin-state of Iron potentially have further effects on mantle viscosity (Badro et al. 2004, Lin et al. 2007). The mantle's radial viscosity variations are likely to be complicated. Suggestions of low viscosity layers in the lower mantle (Matyska et al. 2011, Wentzcovitch et al. 2009), strong radial variations in the upper mantle (Marquart et al. 2005) or mid-mantle viscosity hills (Morra et al. 2010) all contribute to the complexity. Given that so many factors affect mantle viscosity it is easy to see why variations are poorly constrained and understood.

In this chapter, I will generally consider relatively simple models of global mantle convection, comparing the modelled down-wellings to fast regions imaged by seismic tomography. To understand how viscosity variations affect the global flow pattern I neglect some of the complexities such as chemistry and spin-state transitions and consider mantle viscosity structures with depth and temperature dependent viscosity variations only. Inversions of observable data sets, such as the geoid and post-glacial rebound data, can be used to obtain a radial profile of viscosity changes with depth (Forte & Mitrovica 1996, King 1993, Mitrovica & Forte 2004, Ricard & Wuming 1991). These techniques commonly result in a viscosity profile with a thin, high viscosity layer of around $10^{22}$ Pa s in the uppermost 150 km
of mantle. The upper mantle has a viscosity of around $10^{21}$ Pa s down to the transition zone where there can be a further local decrease to $10^{20}$ Pa s. In the lower mantle viscosity is commonly observed to gradually increase to a maximum value of around $10^{22}$ Pa s or higher at 1500 km to 2000 km depth. Below 2500 km the viscosity decreases towards the core-mantle boundary. Despite general agreement there is some variability between different published radial viscosity structures. There is some disagreement on the depth and magnitude of the low viscosity region of the upper mantle. In some cases a reduction of around one order of magnitude in the transition zone is chosen while in other cases a shallower low viscosity asthenosphere is preferred with viscosity decreasing to as low as $10^{19}$ Pa s (Marquart et al. 2005). Deeper in the mantle the increase in viscosity may be more gradual only reaching a maximum viscosity in the deepest mantle (Forte et al. 1993).

Alternatively it is possible to invert seismic tomography to obtain a radial profile of viscosity variations based upon the inferred density structure (King & Masters 1992). This method produces a similar viscosity profile to the above. Notable differences include a slightly thicker low viscosity region in the transition zone extending slightly deeper into the lower mantle and characterised by a viscosity drop of nearly 2 orders of magnitude between 600 and 900 km depth. There is also little in the way of viscosity reduction approaching the core mantle boundary. Similarly, Steinberger & Calderwood (2006) compute radial viscosity profiles using multiple techniques, including mineral physics data and a density structure calculated from seismic tomography. They suggest a viscosity profile with a hill in the lower mantle and calculate a highest viscosity of $10^{23}$ Pa s estimated to be 1000 times greater than the upper mantle viscosity.

A different approach to understanding mantle viscosity is to use high pressure, high temperature mineral physics experiments and theory. For the lower mantle in particular experiments are few and far between and require accurate estimates of temperature and grain
size (Karato 2008). Despite the challenges of experimenting at lower mantle conditions, experiments suggest that a lower mantle viscosity of $10^{22}$ Pas is plausible (Yamazaki & Karato 2001). Experiments of this nature produce profiles of viscosity against depth show little variation in the lower mantle but Karato (2008) does note in chapter 19 that lower mantle viscosity is at a minimum at mid-lower mantle depths.

3.4.1 Viscosity variations with depth

Most published models of mantle viscosity, when simplified, agree upon a two or three layer viscosity model with a step increase into the lower mantle and a more viscous layer near the surface. Therefore, I will first consider these simple cases of radially varying viscosity with a maximum of three layers.

Modelling by Bunge et al. (1996) used TERRA to investigate the effect of Rayleigh number and lower mantle viscosity increase in a three-dimensional spherical mantle convection simulation. The simulations show distinct difference between simulations with and without a factor of 30 increase in viscosity in the lower mantle. Simulations with an isoviscous mantle develop cylindrical down-wellings, whilst the simulations including the lower mantle viscosity increase develop long, linear down-welling sheets. Bunge et al. (1996) note that increasing the Rayleigh number of a convection simulation enhances the effect. In the higher Rayleigh number, isoviscous case cylindrical down-wellings are more closely spaced and in the stratified viscosity case the convection planform is dominated by sheet-like down-wellings.

Jarvis & Lowman (2007) demonstrate the importance of a viscous stratification on the sinking of subducted material. In two dimensions they model the effect of slabs of different geometry sinking in a mantle with stratified viscosity. This is an important consideration for this study as mantle models are attempting to match mid-mantle slabs
imaged by seismic tomography. For each case they consider the time it takes to clear the mid-mantle of slab material, termed ‘slab-survival’ time. The results of a series of simulations by Jarvis & Lowman (2007) clearly demonstrate that larger viscosity contrasts into the lower mantle slow the rate of slab descent. For large slabs, survival times range from 100 thousand years in an unstratified mantle to greater than 500 million years with a viscosity increase of 1000 times. Tomography studies (e.g. van der Voo et al. (1999a,b)) suggest slabs from Tethyan subduction which ceased $\approx 140$ Myr before present reside in the mid-mantle. It is, therefore crucial that a lower mantle viscosity is chosen to allow subducted slabs to remain in the mid-mantle for an appropriate time. A sensible range of step viscosity increases to consider would be between a factor of 10 and 100, where the slab survival times of Jarvis & Lowman (2007) are in the range of tens to hundreds of million years. The magnitude of the step increase in viscosity, and therefore the absolute lower mantle viscosity is clearly an important consideration for this study. The mantle circulation models are attempting to produce temperature fields closely resembling the seismic structure imaged by tomography and therefore the sinking speed of subducted material is crucial to matching the locations of fast anomalies.

This sub-chapter considers the effect of similar viscosity variations on three dimensional, spherical geometry mantle circulation models with assimilated plate motions. I aim to understand whether lower mantle viscosity has an effect on subducting material and what range of lower mantle viscosity increases produce temperature distributions similar to current understanding of Earth’s mantle. Since lower mantle viscosity is likely to be the primary factor in the sinking speed of subducted material (Čižková et al. 2012) the match between modelled subducted slabs and fast regions in tomography will provide a useful constraint on the lower mantle viscosity. I start with a reasonably simple three-layer structure of viscosity. The reference viscosity specified will be for the upper mantle down to 660 km.
The lower mantle will typically be a factor greater than reference viscosity and exist from 660 km to the core mantle boundary. This step increase is graded in over a few hundred kilometres to prevent numerical instability arising from large changes in viscosity over small distances. Finally the third layer is a high viscosity layer, a factor of 50 greater than the reference viscosity in the top 150 - 200 km of the model domain. This is included to simulate a lithosphere within the model. Although the top most layer is not a true lithosphere as it is not a rigid layer it serves to generate an upper thermal boundary layer.

Six mantle circulation models with identical model parameters, except the step increase of viscosity into the lower mantle are considered. Models are incompressible and Boussinesq with both internal and bottom heating. The basic parameters used are summarised in table 3.1. Five of the six cases (101 - 105) consider a three layer radial viscosity model with the upper mantle at the reference viscosity and a thin high viscosity layer to mimic a lithosphere. The lower mantle (from 660 km downwards) has an increase in viscosity of a factor given in table 3.2. The sixth simulation (case 100) is purely isoviscous containing no lower mantle viscosity increase (as case 101) and no near surface viscosity increase. The modelling method is as described in chapter 3.2 and I compare the predicted model temperature anomalies to seismic tomography from six tomography studies.

The radial viscosity variations for each of the cases considered in this subsection are illustrated in Figure 3.3. Note that in the figure some lines overlap. The red line indicates the purely isoviscous case (100), in which in there is no lower mantle viscosity increase, identical to case 101, the orange line. Cases 101 - 105 all have the same uppermost mantle viscosity structure, indicated by the purple line, differing only in the lower mantle. Also note that for numerical convenience the 'step' viscosity increase is graded in over a depth of a few hundred kilometres.
Table 3.1: Model parameters used in simulations to investigate the effects of radial viscosity variations

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reference viscosity (Pa s)</td>
<td>2x10^{22}</td>
</tr>
<tr>
<td>Lower mantle viscosity</td>
<td>See table 3.2</td>
</tr>
<tr>
<td>Temperature dependent viscosity</td>
<td>None</td>
</tr>
<tr>
<td>Mineral phase changes</td>
<td>None</td>
</tr>
<tr>
<td>Equation of state</td>
<td>Incompressible &amp; Boussinesq</td>
</tr>
<tr>
<td>Thermal conductivity (W m^{-1} K^{-1})</td>
<td>4</td>
</tr>
<tr>
<td>Internal heat production (W m^{-3})</td>
<td>4x10^{-12}</td>
</tr>
<tr>
<td>CMB Temperature (K)</td>
<td>3000</td>
</tr>
</tbody>
</table>

Table 3.2: Variable parameters for the six simulations considering varying the radial viscosity profile. Values are increases relative to the reference viscosity, see table 3.1

Figure 3.4 presents global maps of temperature anomaly at 925 km depth for each of the cases described in table 3.2. Each of the plots is on a scale saturating at ± 600 K with reds being hotter than average mantle and blues being colder. It is evident from these plots that simply varying a single parameter, the lower mantle viscosity, has a significant effect upon the nature of convection within the mid-mantle. At the broadest scale each of the simulations produces a temperature field with cold anomalies in similar locations to those imaged by tomography. This confirms that the models are being accurately conditioned by plate motion history and that it is sensible to compare regions of fast seismic velocity to colder than average mantle. Despite the general agreement between each of the model cases
and the robust fast anomalies from seismic tomography the nature of the down-wellings in each model differs. The key difference is the thickness of the cold slabs. Slab thickness increases with increasing viscosity.

At 925 km depth it is difficult to select a preferred model. Based solely upon a visual comparison a different model might be selected depending upon the tomography study considered. Due to resolution differences between tomography studies (S-wave models generally have lower spatial resolution than P-wave models) a different circulation model may be chosen as the best fit depending upon the tomography study. If comparing to the spherical harmonic degree forty S-wave tomography model, S40RTS, a mantle circulation model with a viscosity increase of 100 times (case 105, figure 3.4f) into the lower mantle may be appropriate. However, when comparing to MITP08, a high-resolution body wave
Figure 3.4: Horizontal cross sections through model temperature anomaly predictions from cases 100 - 105 at 925 km depth. Blue colours represent colder than average mantle whilst reds are hotter than average mantle. Labels indicate the case number and the lower mantle viscosity increase.
tomography study, a more modest viscosity jump of 30 times (case 103, figure 3.4d) fits better. Models containing no viscosity increase into the lower mantle result in a poor fit to tomography. Down going slabs are thin in these cases and often more cylindrical than slab like (case 100 & 101, figures 3.4a & b). Although cold material is located beneath convergence zones there is not enough colder than average mantle to produce the imaged velocity perturbations.

There are significant differences between the two simulations with no viscosity increase (cases 100 & 101) and those with a viscosity increase into the lower mantle. In the cases with no increase down-wellings are considerably thinner and mostly more cylindrical in geometry. The temperature anomalies in cases 100 & 101 are also smaller in magnitude when compared to the other cases. These two simulations are quite similar to each other suggesting that the lower mantle viscosity increase is more important than the inclusion of a more viscous lithosphere layer. In cases not presented in this thesis with a viscosity increase in the lower mantle, altering the near surface radial viscosity profile has a limited effect upon the nature of convection. Including a more viscous near surface layer tends to thicken the thermal boundary layer and therefore thicken the slabs sinking into the lower mantle, however the effect is not hugely significant. The magnitude of the lower mantle viscosity is likely to be the more important factor. In later parts of this thesis the near surface increase in viscosity is not included in the radial viscosity profile, but it accounted for by a temperature dependent viscosity law.

Figure 3.5 considers the same model cases at the depth of 1375 km depth in the shell. At this depth tomography indicates broader fast anomalies, possibly as result of slab thickening or accumulation. However at this depth the spherical shell is smaller than the spherical shell at the surface, meaning that anomalies at deeper mantle depths appear larger with present day coastlines super-imposed. This should not present a major problem when
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examining these figures as each of the plots is at the same depth.

The modelled temperature field at this slightly deeper depth indicates that models with a higher viscosity jump in the lower mantle (case 105) do not provide a good match to tomography. A viscosity increase of a factor of 100 slows the down-welling flow enough that the match between it and tomography is not so good, particularly in the Tethys region. There is some similarity between the temperature field at 1375 km depth in case 103 and the temperature field at 925 km depth in case 105, suggesting that in the higher viscosity case down-welling flow is restricted enough that slabs do not reach the depths indicated by tomography. For these deeper slices I suggest that case 103 provides that better match between modelled temperature field and imaged fast anomalies in tomography. The fit is particularly good to S-wave models such as S40RTS, possibly an effect of the model resolution.

In figure 3.6 I present a direct comparison with a subset of the radial viscosity cases (100, 102, 103, 105) and compare them directly to two tomography models (S40RTS (Ritsema et al. 2011) and MITP08 (Li et al. 2008)). I select a 'threshold' value for each data set and plot a contour of that value on a global Hammer projection. The contours essentially define slab boundaries. For S40RTS I select S-wave velocity greater than 0.6% faster than average as the threshold and for MITP08 I select P-wave velocity greater than 0.3% faster than average. Regions with seismic velocity faster than the threshold value are plotted in the figures as a black contour. MITP08 is plotted in the left column and S40RTS in the right column. The red contours on the figures indicate regions of modelled mantle 300 K colder than the layer average temperature, areas of overlap indicate good match between seismic velocity perturbations from tomography and modelled colder than average mantle.

Each of the plots in figure 3.6 is at 1375 km depth with present day coastlines plotted to provide geographic reference points. This figure further confirms that an isoviscous mantle
Figure 3.5: Horizontal cross sections through model temperature anomaly predictions from cases 100 - 105 at 1375 km depth. Blue colours represent colder than average mantle whilst reds are hotter than average mantle. Labels indicate the case number and the lower mantle viscosity increase.
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Comparison to MITP08

(a) Case 100
  isoviscous

(b) Case 100
  isoviscous

(c) Case 102
  x10

(d) Case 102
  x10

(e) Case 103
  x30

(f) Case 103
  x30

(g) Case 105
  x100

(h) Case 105
  x100
(case 100, figure 3.6a,b) is unlikely to be realistic. In this case material 300 K colder than average is not particularly prominent in either the Tethys or Farallon regions. For case 100 the volume of material present is very unlikely to produce the faster than average regions present in tomography. Likewise a viscosity jump of a factor of 100 (case 105, figure 3.6g,h) or more seems implausible too. Although a large volume of cold material is present under North America there is nothing present in the Tethys region at all despite mantle circulation models traditionally being stronger in the Tethys region.

The remaining two simulations present some interesting results. Selecting a preferred model from these two is very much dependent upon which tomography model is used as a reference point. When comparing to the P-wave model MITP08 case 102 (figure 3.6c) provides a better fit as anomalies are a similar size laterally to the anomalies imaged by the P-wave tomography, locations also match up well particularly in the Tethys region. However, in comparison to S40RTS case 103 (figure 3.6f) looks to be a better fit, again primarily in the Tethys region. Case 103 matches the broad anomaly beneath India well. Both cases are noticeably weaker in the Farallon region at this depth, case 102 has a longer more linear anomaly crossing the entire continent from north to south where as in case 103 the features are more broken up. In both cases the model predicts a slab to the west of the location imaged in tomography. This is a known issue with mantle circulation models, observed by

**Figure 3.6 (previous page):** A comparison between fast seismic velocity anomalies from MITP08 & S40RTS with modelled cold temperature anomalies at 1375 km depth. The black contours represent regions of greater than 0.3% velocity perturbation in MITP08 (left column) and greater than 0.6% velocity perturbation in S40RTS (right column) Regions 300 K colder than average from the model cases are contoured in red and overlain. Cases 100 (isoviscous), 102 (factor 10 increase), 103 (factor 30 increase) and 105 (factor 100 increase) are presented.
Bunge et al. (2002) and thought to be a result of TERRA not capturing the shallow dip of the observed Farallon slab during the Laramide orogeny. It is worth noting that increasing the lower mantle viscosity shifts the slab eastward, closer to the imaged location. There is further discussion of the effect of radially varying viscosity in chapter 3.4.3

3.4.2 Temperature dependent viscosity

Up until this point, I have only considered a simple mantle viscosity based upon variations with depth. As suggested in the introduction to this section, the radial profile of viscosity is likely to be more complicated that considered in the previous sub-set of simulations. Here, I present a series of mantle circulation models including temperature dependent viscosity. Including temperature dependent viscosity should have two important effects on the simulations. Firstly, the radial average viscosity should be affected by the ‘background’ mantle temperature, resulting in a radial viscosity profile with a mid-mantle viscosity hill similar to described in Mitrovica & Forte (2004) or Morra et al. (2010). Secondly, the temperature dependence equation will introduce some lateral variability to the viscosity with mantle colder than its surroundings having higher viscosity. In theory this should increase slab strength particularly near the surface resulting in a better match to tomography.

Zhong et al. (2000) performed mantle convection simulations with layered and temperature dependent viscosity alongside surface plates. They found, that similar to Bunge et al. (1996, 1997), long-wavelength mantle structures could be generated by layered viscosity without surface plates. They also observe that temperature dependent viscosity alone has the ability to generate long wavelength structures, however a combination of both layered and temperature dependent viscosity shortens the wavelength of the modelled features. Zhong et al. (2000) go on to say this could be the result of a significant reduction in Rayleigh number due to the higher viscosity in deeper portions of the model. The observations in Zhong et al.

To examine the effect of temperature dependent viscosity upon mantle circulation models I have undertaken a number of simulations based upon case 103 from section 3.4.1. Each of the simulations contains the same basic parameters set described in table 3.3. These parameters are similar to those used to investigate radially varying viscosity, however, a factor of 30 step viscosity increase into the lower mantle is included in each case. For each of these simulations the uppermost viscosity layer (intended to represent the lithosphere) has been reduced from a factor of 50 increase to a factor of 10 as the temperature dependent viscosity causes an increase in viscosity due to colder temperatures near the surface.

A notable weakness in TERRA is an inability to handle large lateral viscosity variations over small distances. Recent improvements to the code by the TERRA community have improved the code in this respect so here I include progressively larger lateral viscosity variations until TERRA becomes numerically unstable and cannot solve the convection calculation. Table 3.4 is a summary of the simulations undertaken. Temperature dependence is implemented according to equation 3.1, the variables for each simulation are listed in table 3.4.

\[
\eta(T) = \eta_0 \exp[-E_a T] \tag{3.1}
\]

where temperature is non-dimensionalised by \(\Delta T\) and \(E_a\) is the fundamental variable controlling the degree of temperature dependence in the model.

It should be noted that the minimum viscosity at which TERRA can accurately perform the convection calculation is highly dependent upon resolution. In order to obtain
**Table 3.3:** Model parameters used in simulations to investigate the effects of radial viscosity variations. All reference viscosities are given prior to the addition of temperature dependence.

Simulations with Rayleigh number as close as possible to Earth-like the reference viscosity included in the calculations is usually close to the lower limit of viscosity that TERRA can compute at this resolution. For this reason the temperature dependent viscosity in hotter than average regions is capped. The minimum viscosity allowed is one half of the reference viscosity. This is acceptable as I generally consider only down-welling flows and their comparison to fast anomalies in seismic tomography. The increase in viscosity due to cold mantle is only capped by the temperature dependent viscosity equation, no artificial maximum is set so that the effect of temperature dependence on down going slabs is still simulated.

Figure 3.7 shows the variations of viscosity with temperature for four of the above cases. The dark blue line is the average radial viscosity for the simulation after the forward modelling phase. Dashed blue lines indicate the minimum and maximum viscosity variations with depth. In case 103 only the radial viscosity variations are shown as no temperature dependent viscosity is included. As the temperature dependence of viscosity increases, as expected, there is a larger range of viscosity at a given depth. There is also a trend
Table 3.4: Variable parameters for the seven simulations used for examining the effect of temperature dependent viscosity on mantle circulation models. $\eta(T)$ factor is the approximate variation from the radial viscosity profile (table 3.3) viscosity arising from the given $E_a$ when used in equation 3.1

<table>
<thead>
<tr>
<th>Case</th>
<th>$E_a$</th>
<th>$\eta(T)$ factor</th>
</tr>
</thead>
<tbody>
<tr>
<td>Case 103</td>
<td>0</td>
<td>1</td>
</tr>
<tr>
<td>Case 125</td>
<td>2.30</td>
<td>10</td>
</tr>
<tr>
<td>Case 126</td>
<td>3.91</td>
<td>50</td>
</tr>
<tr>
<td>Case 127</td>
<td>4.61</td>
<td>100</td>
</tr>
<tr>
<td>Case 127a</td>
<td>5.70</td>
<td>300</td>
</tr>
<tr>
<td>Case 128</td>
<td>6.91</td>
<td>1000</td>
</tr>
<tr>
<td>Case 129</td>
<td>9.21</td>
<td>10000</td>
</tr>
</tbody>
</table>

of increasing mean viscosity as the temperature dependence factor increases. Including temperature dependent viscosity in the models produces a radial average viscosity structure with a sharp decrease in viscosity close to the core mantle boundary, similar to that observed by Mitrovica & Forte (2004). The models do not develop a very low viscosity layer in the upper mantle as suggested in certain studies, largely a consequence of the modelling code being unable to handle very low viscosity at this resolution and the imposed minimum viscosity described above. Including a highly mobile layer in the upper mantle would be possible in the radial viscosity structure if TERRA was able to perform the calculation, but I will not consider it here since current computational resources do not allow sufficient resolution.

Of the above cases three could not be solved by TERRA. Cases 128 & 129 both became numerically unstable during the generation of an initial condition. Limitations in TERRA’s solution algorithm is suspected to be the cause. Varying the solution tolerances slightly improved the situation, however an acceptable solution was not reached. Simulation 127a was included to assess which magnitude of temperature dependent viscosity the
simulations fail. It includes a temperature dependent viscosity factor of $\pm 300$. In this case the code solved the standard convection problem accurately to generate an initial condition. However with the application of plate motions as the surface boundary condition TERRA began to develop instabilities and could not solve the problem. From herein, I will consider only simulation cases 125 - 127. These cases are enough to understand the effect of moderate amounts of temperature dependent viscosity on mantle circulation models. However, I note the limitation of the study as large degrees of temperature dependent viscosity cannot be simulated. Once again I consider the match between modelled down-wellings and seismically fast mantle observed in tomography. Figures 3.8 and 3.9 show the mantle temperature

**Figure 3.7:** Graphs demonstrating the radial viscosity variation for the reference case, 103, the average radial viscosity (solid line), maximum and minimum viscosities (dashed line) for temperature dependent viscosity cases 125 - 127.
anomalies at 925 km and 1400 km depth respectively for these three cases with temperature dependent viscosity and the reference case with no temperature dependent viscosity. Again all plots are on the same temperature scale saturating at ±600 K from the radial average.

Model temperature predictions at 925 km depth for temperature dependent viscosity cases

Figure 3.8: Horizontal cross sections through model temperature predictions from cases 103 & 125 - 127 at 925 km depth. Blue colours represent colder than average mantle whilst reds are hotter than average mantle.

Figure 3.8 shows that in the shallow mid-mantle the degree of temperature dependence does not have a huge effect on the planform of convection. Patterns between the cases are very similar and show thicker slabs with the inclusion of temperature dependent viscosity, particularly beneath the Americas. Somewhat counter-intuitively, simulations with higher degrees of temperature dependence generate less coherent down-welling structures.
Case 125, with a temperature dependent viscosity factor of 10 (figure 3.8b) generates a more coherent, slab-like down-welling in the mantle beneath North America than either of the other simulations. Each of the cases including a degree of temperature dependence generates more voluminous down-wellings beneath North America, resulting in slightly better match to tomography. In the Tethys region models with temperature dependent viscosity seem to generate more cylindrical down-wellings particularly with larger magnitudes of temperature dependence. At this global scale there is little to choose between each of the cases in terms of comparisons to seismic tomography. Each of the three cases shows improvement.
in the match in the Farallon region, but a less good fit in the Tethys region.

Slightly deeper at 1400 km below the surface (figure 3.9) the effect of temperature dependence is slightly more obvious. Both cases 125 & 126 (figures 3.9b,c) do not show much difference from each other and match tomography well in regions of long-lived convergence. However, including a temperature dependent viscosity of factor 100 (figure 3.9d) or more makes the comparisons weaker. In these cases down-wellings are cylindrical in nature and more broken up. Geographically the location of colder than average remains as expected, however the more slab-like behaviour that may be expected is not obvious.

As in the sub-chapter on radially varying viscosity, I compare more directly between tomography models MITP08 and S40RTS. Using the same thresholds for slab boundaries as in that sub-chapter, i.e., $+0.3\% V_p$ perturbation for MITP08, $+0.6\% V_s$ perturbation for S40RTS and 300 K colder than average for the model. Figure 3.10 shows the direct comparison between the two data sets at 1400 km depth. This figure shows that the two cases with lower degrees of temperature dependence in their viscosity (cases 125 & 126, figures 3.10a-d) produce reasonably good matches to both of the plotted tomography data sets. Both cases fit better to S40RTS than MITP08 in terms of lateral extent of anomalies. With case 127 (figures 3.10e,f) the lateral extent of the anomalies is not significantly different from the other two simulations. However the location of subducted material at 1400 km depth does not match as well with locations imaged by seismic tomography. The large volume anomalies beneath the Tethys ocean are not matched particularly well and there is significantly less material under North America than might be expected. It may be that up-wellings are significant in this simulation, as there is a hotter than average region under the Indian continent which may be responsible for some of the poor match.

Overall including temperature dependent viscosity in the mantle circulation models affects slab thickness. Even a modest amount of temperature dependence at this resolution
Figure 3.10: A comparison between fast seismic velocity anomalies from MITP08 & S40RTS with modelled cold temperature anomalies at 1400 km depth. The black contours represent regions of greater than 0.3% velocity perturbation in MITP08 (left column) and greater than 0.6% velocity perturbation in S40RTS (right column). Regions 300 K colder than average from the model cases are contoured in red and overlain. Cases 125 (temperature dependence factor 10), 126 (temperature dependence factor 50) and 127 (temperature dependence factor 100) are presented.
results in thicker slabs more akin to those imaged by S40RTS. The general affect on the flow pattern is less pronounced, and the radial viscosity structure plays a larger role in the general comparison between model predictions and seismic velocity perturbations. Including temperature dependent viscosity in this way also affects the radial viscosity profile, an important consideration at this lower resolution where the reference viscosity is higher.

3.4.3 Discussion of the effects of variable viscosity

Varying viscosity clearly has a key role on how TERRA simulates mantle circulation. Even reasonably small variations can produce large changes in the results, as evidenced in chapter 3.4.1. These models are calculated on a grid with spacing of approximately 50 km in each direction at mid-mantle depths. For TERRA to be numerically resolved at this resolution it requires input parameters resulting in a Rayleigh (Ra) number slightly lower than might be expected for Earth. At lower Ra the scales of convection are broader resulting in the patterns observed here. It is evident that the lower mantle viscosity plays a key role in mantle convection. For models to accurately produce subducted slabs in locations imaged by tomography a lower mantle viscosity increase relative to the upper mantle of a factor of approximately thirty times is required. Equally an increase of one hundred times globally is probably too large. Such observations fit well with radial viscosity variations hypothesised from other geophysical methods (e.g. Jarvis & Lowman 2007, Mitrovica & Forte 2004).

In a 360° vertical cross section at 90° east to 90° west, both of the previously discussed, long-lived subduction zones are observed (figure 3.11). Alongside the Tethys and Farallon anomalies a third robust feature in the lowermost mantle is also observed. van der Voo et al. (1999a) suggested this anomaly, ranging from 1500 km down to the core mantle boundary is the signature of Jurassic age Mongol-Okhotsk subduction. Figure 3.11 shows an example cross section from the MITP08 Li et al. (2008) seismic tomography model. In the
figure blue regions are seismically faster than average mantle and the three distinct regions are visible. In the west, down to mid-mantle depths fast mantle associated with Farallon subduction is observed. Note in this region the cross section is almost parallel to the strike of the subduction zone. In the east, at mid-mantle depths, a feature representing Tethyan subduction is present. To the north of this is a feature extending from the surface to the core mantle boundary, interpreted as Mongol-Okhotsk subduction. In these cases the cross section will be perpendicular to subduction.

\[\text{Figure 3.11: An example of three dimensional visualisation of seismic tomography (MITP08). Three regions of faster than average mantle are imaged in blue colours}\]

Figure 3.12 demonstrates the effect of radially increasing viscosity on the comparison between modelled temperature anomalies and seismic tomography. The figure shows cross sections at $90^\circ$ east through four data sets. Parts (a) & (b) are cross sections of seismic velocity through MITP08 and S40RTS respectively. Parts (c) & (d) are cross sections of
present day modelled temperature anomaly for cases with a factor of 10 increase in viscosity (c) and a factor of 100 increase in viscosity (d) relative to the upper mantle reference viscosity. Figure 3.12 illustrates the difference that a lower mantle increase of one order of magnitude can have. Case 102 generally shows colder than average mantle lying at the core mantle boundary under regions of convergence. The mid-mantle temperatures are closer to ambient. The lack of colder than average mantle at mid-mantle depths suggests that sinking velocity of subducted material is too high in this case, thus the viscosity is too low. Case 102 (figure 3.12c) does produce a good fit to the long, thin fast anomaly imaged in the west at equatorial latitudes. Case 105 (figure 3.12d) shows the opposite. In this case material does not appear to sink to deep enough depths. Broad features are at mid mantle depths and generally do not extend to the depths suggested by tomography. This simulation does produce a cold region which is very similar to the imaged Mongol-Okhotsk slab, particularly in MITP08. Both simulations fail to capture the shallower structure imaged in the Farallon region, this is probably due to a combination of the shallow dip of the Farallon slab, and these cross sections being along strike in the region. Furthermore the effect of higher viscosity on Rayleigh number and therefore the energy of convection is clear in the size of the plume in the lower right quadrant of figure 3.12d. The thickness of the plume is indicative of the sluggish convection arising from lower Rayleigh number convection. These differences demonstrate that a factor of 10 increase in viscosity is too little to accurately match mantle structure, while a factor of 100 increase is likely too high. Although each of the models does capture some of the imaged features well. The overall suggestion is that a simple three layer viscosity structure is likely to be an over simplification of the viscosity of Earth’s mantle.

In cases involving temperature dependent viscosity I observe a slightly unexpected result. As the factor of temperature dependence increases the more slab-like behaviour expected is not observed. Instead the models develop more cylindrical down-wellings, which
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Figure 3.12: An example of three dimensional visualisation of seismic tomography (MITP08). Three regions of faster than average mantle are imaged in blue colours.
by 1400 km depth do not match well with imaged seismically fast mantle. Figure 3.13 shows plots of lateral viscosity variations at 1400 km depth for the model cases 125, 126, & 127. Each plot is presented on the same scale for simplicity. The scale increases logarithmically from $10^{23}$ Pas to $10^{25}$ Pas. It is evident from these plots that the increasing magnitude of temperature dependence results in increasing the mean radial viscosity. In cases 127 much of the mantle at 1400 km depth has a viscosity of greater than $10^{24}$ Pas, equivalent to a factor of 100 increase in a simulation with radially varying viscosity only. As shown in figure 3.5 such a model prevents subducting material from reaching the depths at which it is imaged by tomography. Since the lateral viscosity variations in case 127 are large enough to give large volumes of the mantle such a high viscosity it is perhaps less surprising that these simulations do not match well to tomography. The effect is likely to be compounded in the lower resolution studies. As previously mentioned, TERRA requires a slightly higher reference viscosity to be numerically stable at the 50 km grid spacing. At higher resolution when a more Earth-like upper mantle viscosity is chosen, a larger degree of temperature dependence may be more suitable. In these cases the maximum viscosity in the lower mantle will be lower and so will probably be less restrictive to the sinking material. This work highlights the importance of resolution in these simulations. In the simulations the temperature dependent viscosity generates lateral viscosity variations that in places become very high, thus convection is particularly sluggish.

These findings are consistent with Zhong et al. (2000) who suggest that the combination of temperature and radially varying viscosity results in mantle models with shorter-wavelength structure. Whilst temperature dependent viscosity adds some more complexity to both the radial variations and lateral variations of viscosity it seems that the more simple three-layer radial viscosity profile captures seismic structure to the first order. Radial viscosity variations have a larger effect upon the convective pattern of mantle circulation models
Figure 3.13: Horizontal cross sections demonstrating lateral viscosity variations at 1400 km depth in the model for the temperature dependent viscosity cases 125-127a. Viscosity is calculated at present day and presented on a logarithmic scale.
with an order of magnitude variations producing large changes in the convective system. Temperature dependent viscosity although important does not play such a large role and at this resolution can only be included with small to moderate dependence.
3.5 The effect of mineral phase changes at 410 and 660 km depth upon mantle circulation models

One dimensional profiles of variations of seismic velocity and density with depth, for example PREM (Dziewonski & Anderson 1981), reveal two sharp increases in seismic velocity and density between 400 and 700 km depth in Earth’s mantle. These increases are thought to be a result of phase changes in the olivine mineral system (Ringwood 1970). Between these depths, minerals undergo transitions from olivine to wadsleyite to spinel-structure (ringwoodite) to ferropericlase and perovskite-structure minerals (Deuss & Woodhouse 2001, Ita & Stixrude 1992, Ringwood 1975). The changes in mineral structure result in increased density and it should be expected an increase in seismic velocity. The boundary separating the phases is frequently approximated by a straight line in pressure-temperature space. The gradient of the line, \( \frac{dP}{dT} \), is known as the Clapeyron slope. The transition from olivine to wadsleyite is an exothermic process (the reaction gives out energy to its surroundings) and therefore has a positive Clapeyron slope. In contrast the phase change from ringwoodite to ferropericlase and Mg-perovskite is endothermic (absorbs heat from the surroundings) and has a negative Clapeyron slope (Christensen 1995). Both have implications for the dynamics of convection. For example, the density increase occurs at shallower depth in colder than average mantle during an exothermic phase change, resulting in more dense, negatively buoyant material at shallow depth, enhancing convection. An endothermic phase transition (as in Figure 3.14) in restrictive to flow as in colder than average material the transition occurs slightly deeper resulting in lower density minerals (ringwoodite) surrounded by the more dense phase (perovskite). The same restrictive behaviour applies to hotter than average up-welling material. In an endothermic phase change the transition from perovskite to ringwoodite occurs shallower (figure 3.14B) generating a restoring buoyancy.
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Given large enough density difference and Clapeyron slope the colder material is locally more buoyant and so the transition acts as a barrier to flow. This is dynamically important as the ringwoodite to ferropericlase and Mg-perovskite transition is endothermic and most likely occurs at 660 km depth, the boundary between the upper and lower mantle (Ito & Takahashi 1989). Phase changes with a sufficient negative Clapeyron slope are capable of layering a convecting system (Christensen 1995, Olson & Yuen 1982). In this section I will investigate the effect of two phase transformations, an exothermic transformation at 410 km depth and an endothermic transformation at 660 km depth upon the nature of convection in mantle circulation models.

**Figure 3.14:** Illustration of layering process caused by an endothermic phase change at 660 km depth. After Wolstencroft & Davies (2011).

3.5.1 The endothermic phase change at 660 km depth

It has long been speculated that the endothermic phase change at 660 km depth may be responsible for some degree of layering in mantle convection. This may be two inde-

There have been numerous numerical models investigating the effect of phase transitions on mantle convection (e.g. Bunge et al. (1997), Liu et al. (1991), Machetel et al. (1995), Machetel & Weber (1991), Peltier & Solheim (1992), Tackley et al. (1994), Zhao et al. (1992)). As might be expected these models produce a variety of mantle behaviours including; causing up-wellings and down-wellings to pause in the transition zone (Bunge et al. 1997), intermittently layered convection (Machetel et al. 1995, Machetel & Weber 1991), strong layering with 'brief excursions' into whole mantle convection (Peltier & Solheim 1992) and varying behaviour depending upon the physical properties of flow features (Liu et al. 1991, Tackley et al. 1994, Zhao et al. 1992). As these early simulations often considered a restricted range of Rayleigh number and Clapeyron slope it is perhaps not unexpected that a variety of convective behaviours are found across a range of studies. Simultaneous investigations by Yanagisawa et al. (2010) and Wolstencroft & Davies (2011) attempt to classify convection simulations using TERRA in the parameter space between Rayleigh number and phase buoyancy parameter (closely related to the Clapeyron slope). Both investigations reveal three distinct domains of convection; one where the whole mantle convects as a single layer, one where the mantle forms two independently convecting layers
either side of the 660 km discontinuity and a third partially-layered or transitional system, characterised by either periods of two layer convection with periodic break down of layering and a flush of material from upper to lower mantle or where significant parts of the mantle show layered behaviour whilst others evidence unimpeded flow. To illustrate the distinction between the three described domains of mantle convection the results of the Wolstencroft & Davies (2011) study are reproduced in figure 3.15.

Figure 3.15: The data of Wolstencroft & Davies (2011) plotted on a graph of Rayleigh number against phase buoyancy parameter

In the present day Earth estimates of Clapeyron slope at 660km vary from \(-0.5\) MPaK\(^{-1}\) to \(-3.5\) MPaK\(^{-1}\). Different values are predicted from different methods. Mineral physics estimates for dry mantle tend to favour the less negative values (Ishii et al.
2012, Katsura et al. 2003, Litasov et al. 2005). However, in a wet mantle more negative values may be expected (Ohtani & Litasov 2006). Seismological constraints generally favour more negative Clapeyron slopes, between -2 \( \text{MPaK}^{-1} \) and -3.5 \( \text{MPaK}^{-1} \) (Lebedev et al. 2002). I choose a series of Clapeyron slopes for the model based upon the values chosen by Wolstencroft & Davies (2011). Although some of these values are considerably larger than might be expected on Earth it is important to consider extreme cases, particularly as these simulations are at Rayleigh number lower than the best estimates for Earth.

In this section I investigate the effect of the 660 km phase transition on mantle circulation models. I present a comparison of 8 cases, based upon case 103 from section 3.4 and examine the effect of the phase transition on the match between model predictions and imaged slabs from seismic tomography. This section also looks at how the combination of assimilated plate motions and the endothermic phase change affects the nature of convection and which of the three convective regimes the simulations fall into. The eight cases presented here have the same basic parameters as the other models throughout this chapter (listed in table 3.5). For this sub-set of simulations the chosen radial viscosity profile is as in case 103, a factor of 30 increase into the lower mantle. No temperature dependent viscosity is included. Only the magnitude of the Clapeyron slope at 660 km depth is varied between simulations. The cases considered are summarised in table 3.6. This is in line with Wolstencroft & Davies (2011) as it keeps the ratio of Rayleigh number to internally heated Rayleigh number constant ensuring models can be compared.

To understand the effect the phase transition has on convection I consider radial temperature maps at depths just above and just below the transition, in this section I choose 600 km and 800 km. In strongly layered cases it is expected that down-welling material subducting from surface convergent boundaries would pond or stall above the transition,
### Table 3.5: Model parameters used in simulations to investigate the effects of 660 km phase transition.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reference viscosity (Pa s)</td>
<td>$2 \times 10^{22}$</td>
</tr>
<tr>
<td>Lower mantle viscosity</td>
<td>Factor of 30 increase</td>
</tr>
<tr>
<td>Temperature dependent viscosity</td>
<td>None</td>
</tr>
<tr>
<td>Clapeyron slope at 660 km depth (MPaK$^{-1}$)</td>
<td>see table 3.6</td>
</tr>
<tr>
<td>Density jump across 660 km phase transition</td>
<td>9.1 %</td>
</tr>
<tr>
<td>Clapeyron slope at 410 km depth (MPaK$^{-1}$)</td>
<td>none</td>
</tr>
<tr>
<td>Density jump across 410 km phase transition</td>
<td>6.4 %</td>
</tr>
<tr>
<td>Equation of state</td>
<td>Incompressible &amp; Bousinesq</td>
</tr>
<tr>
<td>Thermal conductivity (Wm$^{-1}$K$^{-1}$)</td>
<td>4</td>
</tr>
<tr>
<td>Internal heat production (Wm$^{-3}$)</td>
<td>$4 \times 10^{-12}$</td>
</tr>
<tr>
<td>CMB Temperature (K)</td>
<td>3000</td>
</tr>
<tr>
<td>Ratio of Ra to bottom heated Ra</td>
<td>0.0718</td>
</tr>
</tbody>
</table>

resulting in broad cold regions. Likewise beneath the transition up-welling plumes unable to break through the layering would result in regions of hotter than average mantle. In cases with two independently convecting layers there may be very little correlation between temperature fields above and below the phase transition. In cases where ponding of material is followed by a breakdown in the layering mechanism patterns of temperature either side of the transition will appear different with the upper mantle flow dominated by slab like down-wellings and the lower mantle with more cylindrical shaped features reminiscent of slab avalanches. Figures 3.16 and 3.17 show these horizontal temperature slices through each of the simulations at 600 km and 800 km depth respectively. The plots present the temperature anomaly, i.e. the calculated mantle temperature with the layer mean temperature subtracted.

In each of the cases the planform of down-going material is very similar to the reference case. Cold anomalies below the transition are consistent with cold anomalies above the transition in each individual case. Evidently the velocity boundary condition is dominating
Case Clapeyron slope at 660 km depth
Case 103 0
Case 106 -0.5 MPaK$^{-1}$
Case 107 -1.0 MPaK$^{-1}$
Case 108 -2.0 MPaK$^{-1}$
Case 109 -4.0 MPaK$^{-1}$
Case 110 -8.0 MPaK$^{-1}$
Case 111 -16 MPaK$^{-1}$
Case 112 -32 MPaK$^{-1}$

Table 3.6: Clapeyron slope variations from reference case 103 for seven simulations considering a single endothermic phase change at 660 km depth.

the convection at upper/uppermost lower mantle depths. The phase change clearly affects the convection, particularly at more negative Clapeyron slopes. Above the transition zone (Figure 3.16) depth cold anomalies are still located beneath major convergence zones. In cases with larger magnitude phase changes these anomalies are considerably broader laterally as well as appearing larger in magnitude. The general pattern of cold material beneath regions of long lived convergence is preserved regardless of magnitude of Clapeyron slope. Comparing the cases to the reference case (103, figure 3.16a) the most noticeable change with increasingly negative Clapeyron slope is a thickening of 'slabs' beneath convergence zones. The thickening is most noticeable between cases 108 (cl660 = -2 MPaK$^{-1}$, figure 3.16d) and 110 (cl660 = -8 MPaK$^{-1}$, figure 3.16f).

Immediately below the transition at 800 km (Figure 3.17) the effect of the phase transition upon the nature of convection becomes more obvious. In cases with a Clapeyron slope of less than -2 MPaK$^{-1}$, convection does not appear to be heavily affected by the phase transition. The pattern of temperature anomaly below the transition is very similar to the pattern at 600 km. Both up-welling plumes and down going slabs pass through the upper mantle-lower mantle boundary. In case 109 (Cl660 = -4 MPaK$^{-1}$, figure 3.17e) the temperature fields reveal a more transitional nature through the phase change. Material
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Temperature field at 600 km depth for cases with a single phase transition at 660 km depth

(a) Case 103  
cl660 = 0

(b) Case 106  
cl660 = -0.5

(c) Case 107  
cl660 = -1.0

(d) Case 108  
cl660 = -2.0

(e) Case 109  
cl660 = -4.0

(f) Case 110  
cl660 = -8.0

(g) Case 111  
cl660 = -16

(h) Case 112  
cl660 = -32

Temperature Anomaly (K)
appears to stall, as evidenced by thickened down-wellings at 600 km. Despite the pattern of
temperature anomalies being quite similar at 600 km and 800 km there are some differences
including a more cylindrical pattern to the cold down-wellings below the transition. At more
negative Clapeyron slopes the transition becomes very restrictive to flow. Up-wellings in
Cases 110 - 112 (cl660 = -8 to -32 MPaK$^{-1}$, figure 3.17f-g) are broad cylindrical features
resembling plume heads. They are much larger laterally than similar up-wellings in cases 106
- 108 (figures figure 3.17b-d. Down-wellings in cases 110 to112 are still located in beneath
areas of long-lived convergence but are vastly broader than might be expected.

At mid-mantle depths the comparison of the model temperature predictions to
seismic velocity from tomography studies provides a further constraint upon the effect of
phase changes on mantle convection. At 1400 km depth I continue to consider the two
regional scale fast anomalies imaged across a variety of seismic tomography studies. The
first striking north to south under the North American continent. The second is a north-
west to south-east trending fast anomaly stretching from the present day Mediterranean to
the Indian continent. These fast anomalies are thought to exist from the subduction of the
Farallon and Tethys oceans respectively and are called by those names herein. To re-cap
from the previous sub-chapter on viscosity variations figure 3.18 shows global slices of seismic
velocity variation for two of the considered tomography studies. The horizontal cross sections
presented are at depths close to 1400 km. Figure 3.18a is S-wave velocity perturbation from
spherical harmonic degree 40 model, S40RTS at 1390 km depth (Ritsema et al. 2011) and
Figure 3.18b is P-wave velocity perturbation from body wave model, MITP08 (Li et al.

Figure 3.16 (previous page): Horizontal temperature anomaly slices from cases 103 -
112 at 600 km depth in the model. Cases are labelled with the Clapeyron slope at 660km in
MPaK$^{-1}$
Temperature field at 800 km depth for cases with a single phase transition at 660 km depth

(a) Case 103
c_l660 = 0

(b) Case 106
c_l660 = -0.5

(c) Case 107
c_l660 = -1.0

(d) Case 108
c_l660 = -2.0

(e) Case 109
c_l660 = -4.0

(f) Case 110
c_l660 = -8.0

(g) Case 111
c_l660 = -16

Case 112
c_l660 = -32
2008) at 1400 km depth. These figures highlight the faster than average (blue) mantle with which model predictions are compared which are robust across all the tomography models considered throughout this chapter.

Figure 3.19 shows further horizontal cross sections of the modelled temperature anomaly for the cases described above at 1400 km depth.

Comparing cold anomalies representing subducted slabs from the model predictions provides information on how the phase transition affects convective flow. If the phase transition is highly restrictive to flow the modelled temperature anomalies are highly unlikely to match mid-mantle seismic fast anomalies from tomography. In particular the good match between colder than average mantle and fast seismic anomalies observed in section 3.4 will not be evident as subducted material is prevented from reaching the lower mantle by the phase transition. The temperature anomaly plots in figure 3.19 tell a very similar story to the shallower plots. In cases with a Clapeyron slope between -0.5 and -2 \( MPaK^{-1} \) (figures 3.19b-d) there is very little difference between the cold anomalies and those in the reference case. The Tethys and Farallon anomalies are present and there is little change in lateral extent of the anomalies. Once again the difference begins to become apparent with cases 109 (\( cl660 = -4\ MPaK^{-1} \), figure 3.19e) and 110 (\( cl660 = -8\ MPaK^{-1} \), figures 3.19f). In these two cases and cases with more negative Clapeyron slope the temperature anomalies at 1400 km depth begin to diverge from the reference case and do not match as well with tomography. Intuitively, the more negative Clapeyron slopes produce temperature fields differing more from tomography. Even at the most negative Clapeyron slope there is no

**Figure 3.17 (previous page):** Horizontal temperature anomaly slices from cases 103 - 112 at 800 km depth in the model. Cases are labelled with the Clapeyron slope at 660km in \( MPaK^{-1} \).
Figure 3.18: Horizontal slices of velocity perturbation from two recent tomography studies showing p and s wave variations at approximately 1400km depth.
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Temperature field at 1400 km depth for cases with a single phase transition at 660 km depth

(a) Case 103
cl660 = 0

(b) Case 106
cl660 = -0.5

(c) Case 107
cl660 = -1.0

(d) Case 108
cl660 = -2.0

(e) Case 109
cl660 = -4.0

(f) Case 110
cl660 = -8.0

(g) Case 111
cl660 = -16

(h) Case 112
cl660 = -32

Temperature Anomaly (K)
evidence to suggest that the mantle enters an entirely two layer convection regime. The temperature fields at 1400 km depth still show some cold material in the mid-mantle beneath areas of long-lived subduction, although it does not match well with tomography the colder than average anomalies are not as broadly distributed as one might expect in an independently convecting lower mantle.

To fully understand the fit between tomography and model predictions I plot contours of a threshold value for MITP08 and the modelled temperature anomaly in Figure 3.20. The figure shows horizontal slices at 1400 km depth with black contours representing the outline of regions with greater than $+0.3\% V_p$ perturbation in the MITP08 tomography study. Overlaid onto this are red contours representing regions of 300 K colder than average mantle for model cases 107 to 112. These plots confirm the conclusions made above. Although no case is a perfect match to the tomography of Li et al. (2008), cases with Clapeyron slopes of between -1 and -2 $MPaK^{-1}$ (cases 107 & 108, figures 3.20c,d) produce cold down-wellings in linear patterns very similar to those imaged by tomography. There is some lateral mismatch in the Farallon region, as noted before this probably exists as the model does not accurately capture the shallowly dipping nature of the Farallon slab during the Cretaceous and early Tertiary. The temperature field from case 109 (figure 3.20e) captures some aspects of the tomography well, it is probably the best match to Farallon tomography, although there is still some westward mis-location. There are some larger volume, slab-like features in the Tethys region, however, there is some mismatch in location. In the cases with higher magnitude phase changes the fit between tomography and temperature is quite poor. The linear

**Figure 3.19 (previous page):** Horizontal temperature anomaly slices from cases 103 - 112 at 1400 km depth in the model. Cases are labelled with the Clapeyron slope at 660km in $MPaK^{-1}$
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(a) Case 103
c_{l660} = 0

(b) Case 106
c_{l660} = -0.5

(c) Case 107
c_{l660} = -1.0

(d) Case 108
c_{l660} = -2.0

(e) Case 109
c_{l660} = -4.0

(f) Case 110
c_{l660} = -8.0

(g) Case 111
c_{l660} = -16

(h) Case 112
c_{l660} = -32
nature of the fast anomalies is lost and the location of cold regions relative to seismically fast regions is poor. This suggests that some layering is occurring at more negative Clapeyron slope phase changes as the subducted slabs are not maintaining their geometry or locations, but the fact that some cold/down-welling material does reach this depth suggest that even at Clapeyron slopes of $-32 \text{ MPaK}^{-1}$ (figure 3.20h) the layering is only partial. One further case with a 660 km phase change with a Clapeyron slop of $-64 \text{ MPaK}^{-1}$ was simulated, in this case the code appeared to be numerically unstable but results hinted at only partial layering with some down-welling material still passing through the transition. Thus far, most evidence suggests that given plate motions as the surface velocity boundary condition it becomes very difficult to achieve a fully layered convective regime.

A further criterion for defining which regime a simulation falls into (as defined by Wolstencroft & Davies (2011)) is based upon the absolute radial mass flux across each radial layer of the model. In all cases radial mass flux is zero at upper and lower boundaries as there is no mass transfer into or out of the model. In a fully layered state mass flux across the 660 km transition would also be equal to zero as no material passes across the boundary. For a simulation in the whole mantle regime a plot of radial mass flux against depth is a smooth curve with a mid-mantle maximum. Any simulation with a radial mass flux profile of

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**Figure 3.20 (previous page):** Demonstrating the comparison between fast seismic velocity anomalies from MITP08 with modelled cold temperature anomalies. Regions with a velocity perturbation of greater than 0.3% in MITP08 are outline in thick black lines, regions 300 K colder than average from the model cases 103 & 106-112 are outlined in red. Areas of overlap indicate a good match between modelled temperature field and p-wave tomography. Present day coastlines are included for reference. Figures are labelled with Clapeyron slope at 660 km depth in MPaK$^{-1}$.
between 10% and 90% of the maximum at 660 km depth is deemed transitional or partially layered. Figure 3.21 shows a comparison of two hypothetical mantle convection cases (a: whole mantle convection, b: fully layered mantle convection). These figures demonstrate the idealised profiles for the end-member convection scenarios for simulations without an imposed velocity boundary condition.

**Figure 3.21:** Schematic of mass flux (solid line) and layer average temperature (dashed line) profiles for (a) whole mantle convection and (b) two-layer convection. After Wolstencroft & Davies (2011)

In Figure 3.22 I present graphs of the radial mass flux against depth for the six cases from 107 to 112. The mass flux profiles for the mantle circulation models are not identical to those for the standard convection models presented by Wolstencroft & Davies (2011). Two differences are noted; first, in the assimilated plate motion cases, the radial mass flux at the surface appears to be non-zero. This is a graphical illusion due to the effect of imposing velocities as the surface boundary condition. Imposed velocities are strictly horizontal with no radial component, however, there is likely to be high near surface mass
flux due to the imposed velocity boundary condition. Secondly, the peak of mass flux for mantle circulation models is shallower than in the standard cases. This hints that during mantle circulation simulations more mass is transferred between radial layers in the upper mantle, probably as a result of relatively quick sinking of subducted material through the upper mantle. Despite the differences in radial mass flux profiles between the two different types of model a significant decrease in radial mass flux is still expected when introducing a negative Clapeyron slope phase change at 660 km depth in the model.

**Figure 3.22:** Graphs of radial mass flux against depth for six cases with Clapeyron slopes ranging from -1 to -32 MPaK$^{-1}$. Mass flux is calculated at present day after 300 million years of mantle circulation modelling. Clapeyron slopes at 660 km depth in MPaK$^{-1}$ are labelled on each plot.
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The radial mass flux profiles support the evidence from the temperature anomaly maps in figures 3.16 to 3.19. For cases with a Clapeyron slope of less than \(-4\, MPaK^{-1}\) (figures 3.22a-c), the reduction in mass flux across 660 km is very small, suggesting that the phase transitions have a very minor layering effect with the vast majority of convection occurring on whole mantle scales. The findings are not dissimilar from the results of Wolstencroft & Davies (2011) at this Rayleigh number and low Clapeyron slope. At more negative Clapeyron slopes, convective behaviour differs slightly from cases with no imposed plate motions. The decreases in radial mass flux across the 660 km boundary are more evident in cases with no assimilated plate motions with transition to a fully layered system occurring at a Clapeyron slope of approximately \(-10\, MPaK^{-1}\). A key observation in these plots is that the mantle never appears to develop a fully layered convection regime. Even at an unrealistically high Clapeyron slope of \(-32\, MPaK^{-1}\) (figure 3.22f), the decrease in mass flux is only around one half of the maximum.

The difference between the two classes of simulations are evident. In mantle convection simulations with no assimilated data, it is easy to classify models into three separate convective regimes. When conducting similar simulations whilst assimilating plate tectonic information, the models fit into only two regimes, whole mantle convection and a transitional partially layered regime. There is further discussion of these results in section 3.5.3.

3.5.2 The exothermic phase change at 410 km depth

The exothermic phase transition from olivine to wadsleyite at 410 km depth in the mantle is thought to be enhancing to convection. The positive Clapeyron slope means that the change occurs shallower in colder than average, sinking material. The increase in density resulting from the transformation therefore has negative buoyancy and promotes sinking. In hotter than average mantle, the phase change occurs so that less dense material
is present deeper, thus its buoyancy promotes convection. Experimental mineral physics
suggests values of between 2.5 MPaK\(^{-1}\) and 4.0 MPaK\(^{-1}\) (Katsura et al. 2004) for the
Clapeyron slope of the 410 km deep phase transition, values consistent with seismic evidence
(Bina & Helffrich 1994, Lebedev et al. 2002).

In this section I present five cases considering the effect of the exothermic phase
change from olivine to wadsleyite at 410 km depth in the mantle. In these simulations the
parameters are kept identical to case 109, each simulation has mostly similar parameters
to those in section 3.4.1. The model properties are repeated in table 3.7. A factor of 30
viscosity increase into the lower mantle and the -4 MPaK\(^{-1}\) endothermic phase change at
660 km depth are included in the basic parameter set. The only parameter varied between
these cases is the value for the Clapeyron slope at 410 km. The simulations considered are
summarised in table 3.8.

In previous studies of mantle convection with phase transitions, a Clapeyron slope of -4

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reference viscosity (Pa s)</td>
<td>2x10(^{22})</td>
</tr>
<tr>
<td>Lower mantle viscosity</td>
<td>Factor of 30 increase</td>
</tr>
<tr>
<td>Temperature dependent viscosity</td>
<td>None</td>
</tr>
<tr>
<td>Clapeyron slope at 660 km depth (MPaK(^{-1}))</td>
<td>-4.0</td>
</tr>
<tr>
<td>Density jump across 660 km phase transition</td>
<td>9.1 %</td>
</tr>
<tr>
<td>Clapeyron slope at 410 km depth (MPaK(^{-1}))</td>
<td>see table 3.8</td>
</tr>
<tr>
<td>Density jump across 410 km phase transition</td>
<td>6.4 %</td>
</tr>
<tr>
<td>Equation of state</td>
<td>Incompressible &amp; Bousinesq</td>
</tr>
<tr>
<td>Thermal conductivity (Wm(^{-1})K(^{-1}))</td>
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</tr>
<tr>
<td>Internal heat production (Wm(^{-3}))</td>
<td>4x10(^{-12})</td>
</tr>
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<td>CMB Temperature (K)</td>
<td>3000</td>
</tr>
<tr>
<td>Ratio of Ra to bottom heated Ra</td>
<td>0.0718</td>
</tr>
</tbody>
</table>

Table 3.7: Model parameters used in simulations to investigate the effects of varying the
Clapeyron slope of the 410 km deep phase transition given a Clapeyron slope of -4.0 MPaK\(^{-1}\)
at 660 km depth.
Table 3.8: Clapeyron slope variations from reference case 109 at the 410 km phase transitions for five simulations considering both transition zone phase changes.

\[
\begin{array}{ll}
\text{Case} & \text{Clapeyron slope at 410 km depth} \\
\hline
\text{Case 109} & 0 \, MPaK^{-1} \\
\text{Case 113} & 0.5 \, MPaK^{-1} \\
\text{Case 114} & 1.0 \, MPaK^{-1} \\
\text{Case 115} & 2.0 \, MPaK^{-1} \\
\text{Case 116} & 4.0 \, MPaK^{-1} \\
\text{Case 117} & 8.0 \, MPaK^{-1} \\
\end{array}
\]

\[
MPaK^{-1} \text{ is often considered for the 660 km deep phase transition. For this reason I choose case 109 as the reference case for this sub-set of simulations. In section 3.5.1 case 109 resulted in convection which showed few characteristics of layering. The mass flux suggested that subducting material passed unhindered through the upper mantle lower mantle boundary. This is a useful reference case as it should be useful to demonstrate the effect of a phase transition at 410 km depth on the convective system. As with the simulations in section 3.5.1 I will consider the horizontal cross sections through the mantle model at depths slightly shallower and slightly deeper than the 660 km deep phase transition. As in cases with only an endothermic phase change examining these temperature maps should reveal the effect of the exothermic phase transition on the flow. Figures 3.23 and 3.24 show the two sets of plots for this sub-set of simulations at 600 km and 800 km depth.}

It is difficult to discern from these plots the effect of the olivine to wadsleyite phase transition at 410 km depth. At low magnitude Clapeyron slopes for the 410 km transition (figures 3.23 & 3.24a-c) the simulation is visually inseparable from the reference case with no phase change at 410 km. Even with more positive values for the Clapeyron slope (figures 3.23 & 3.24d-f) there is very little noticeable difference between the simulations other than a slight decrease in the magnitude of the anomalies. The horizontal temperature anomaly maps are strikingly similar between each of the cases. As observed in chapter 3.5.1,
Figure 3.23: Horizontal cross sections of temperature anomaly from cases 109 & 113 - 117 at 600 km depth in the model. Each case is labelled with the Clapeyron slope of the phase change at 410 km depth. All quoted Clapeyron slopes are in MPa K$^{-1}$. 
Figure 3.24: Horizontal cross sections of temperature anomaly from cases 109 & 113 - 117 at 800 km depth in the model. Each case is labelled with the Clapeyron slope of the phase change at 410 km depth. All quoted Clapeyron slopes are in MPaK$^{-1}$. 
figures 3.23 & 3.24 also show that there is very little disruption to the flow pattern across the 660 km phase transition for phase changes with Clapeyron slopes of this magnitude. The effects observed across the upper mantle to lower mantle boundary are very similar and therefore I suggest that including a second phase transition at 410 km depth has little effect on the overall mantle dynamics. For this sub-set of simulations this is very much in agreement with Wolstencroft & Davies (2011) in saying that including a 410 km phase change has little effect on the nature of mantle convection.

As these cases show that including an exothermic phase transition at 410 km depth has very little effect on the computed temperature field in the upper mantle it is expected that there will be few differences at mid-mantle depths. Each case should differ very little from the reference and provide a similar match to seismic tomography. For completeness, I include temperature anomaly maps at 1400 km depth for the reference case 109 and the cases 113 -117 in figure 3.25. As expected, the map of cold anomalies at 1400 km depth reveals very little difference between the reference case and any of the models including a phase change at 410 km depth. There is perhaps a small reduction in anomaly amplitude in cases with the largest Clapeyron slope at 410 km other than this the mid-mantle temperature field confirms the observations made in the transition zone. It is clear that including the exothermic olivine to wadsleyite phase transition at 410 km depth in a mantle circulation model has very little effect on the nature of convection produced in the simulation. In these simulations the negative Clapeyron slope of the ringwoodite to perovskite mineral phase change is playing the dominant role in convection.

I have also considered a further sub-set of models with varying Clapeyron slope at 410 km depth based upon case 108. This is the case with a factor of 30 increase in viscosity and a $-2.0 \text{ MPaK}^{-1}$ Clapeyron slope at 660 km depth. This sub-set of simulations is considered to attempt to understand the effect of a 410 km deep exothermic phase change
Figure 3.25: Horizontal cross sections of temperature anomaly from cases 109 & 113 - 117 at 1400 km depth in the model. Each case is labelled with the Clapeyron slope of the phase change at 410 km depth. All quoted Clapeyron slopes are in MPaK$^{-1}$. 
on a more weakly layered system. For this sub-set of simulations the reference case (108) shows no signs of layering at all. Down-welling slabs modelled for this case remain similar in thickness and location to cases with no phase changes. This sub-set of models should be useful to demonstrate if a positive Clapeyron slope at 410 km depth has any effect on convection in these simulations. Table 3.10 considers five models with varying Clapeyron slope at 410 km depth. The values chosen are identical to the previous sub-set of models based upon case 109. The basic parameters for the simulation are also as in the above sub-set. They are reproduced in table 3.9.

<table>
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<th>Value</th>
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</thead>
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<tr>
<td>Reference viscosity ($Pa\cdot s$)</td>
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<tr>
<td>Lower mantle viscosity</td>
<td>Factor of 30 increase</td>
</tr>
<tr>
<td>Temperature dependent viscosity</td>
<td>None</td>
</tr>
<tr>
<td>Clapeyron slope at 660 km depth ($MPaK^{-1}$)</td>
<td>-2.0</td>
</tr>
<tr>
<td>Density jump across 660 km phase transition</td>
<td>9.1 %</td>
</tr>
<tr>
<td>Clapeyron slope at 410 km depth ($MPaK^{-1}$)</td>
<td>see table 3.10</td>
</tr>
<tr>
<td>Density jump across 410 km phase transition</td>
<td>6.4 %</td>
</tr>
<tr>
<td>Equation of state</td>
<td>Incompressible &amp; Bousinesq</td>
</tr>
<tr>
<td>Thermal conductivity ($W m^{-1} K^{-1}$)</td>
<td>4</td>
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<tr>
<td>Internal heat production ($W m^{-3}$)</td>
<td>$4 \times 10^{-12}$</td>
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<td>CMB Temperature ($K$)</td>
<td>3000</td>
</tr>
<tr>
<td>Ratio of Ra to bottom heated Ra</td>
<td>0.0718</td>
</tr>
</tbody>
</table>

Table 3.9: Model parameters used in simulations to investigate the effects of varying the Clapeyron slope of the 410 km deep phase transition given a Clapeyron slope of -2.0 $MPaK^{-1}$ at 660 km depth.

I follow the same approach as earlier in this chapter. Horizontal temperature maps in, and just below the transition zone are used to examine the effect of the two phase changes in the upper mantle. Comparisons of modelled temperature anomalies to seismic tomography at 1400 km depth should reveal the more general effect on convection down into
Case Description

<table>
<thead>
<tr>
<th>Case</th>
<th>Description</th>
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<tbody>
<tr>
<td>Case 108</td>
<td>$0 \text{ MPaK}^{-1}$</td>
</tr>
<tr>
<td>Case 118</td>
<td>$0.5 \text{ MPaK}^{-1}$</td>
</tr>
<tr>
<td>Case 119</td>
<td>$1.0 \text{ MPaK}^{-1}$</td>
</tr>
<tr>
<td>Case 120</td>
<td>$2.0 \text{ MPaK}^{-1}$</td>
</tr>
<tr>
<td>Case 121</td>
<td>$4.0 \text{ MPaK}^{-1}$</td>
</tr>
<tr>
<td>Case 122</td>
<td>$8.0 \text{ MPaK}^{-1}$</td>
</tr>
</tbody>
</table>

Table 3.10: *Clapeyron slope variations from reference case 108 at the 410 km phase transitions for five simulations considering both transition zone phase changes.*

As in the simulations with a larger negative phase change at 660 km this sub-set of simulations shows very little difference between cases with and without a 410 km phase change. At 600 km depth (figure 3.26) there is some thinning of cold anomalies associated with subduction, particularly at larger Clapeyron slope. Cold anomalies are still located in the same regions as in the reference case. The thinning of slabs in the transition zone suggests that at higher Clapeyron slope the 410 km phase change is promoting downward flow in the subducting material more than the 660 km transition is restricting it in this case. Since thinning slabs were less obvious in cases with more negative Clapeyron slope at 660 km this is a plausible conclusion. Beneath the transition zone at 800 km depth close inspection reveals a similar story. To the first order there appears to be very little difference between each of the cases, however there does appear to be some thinning of colder anomalies at the largest modelled values of Clapeyron slope for exothermic phase transition.
Figure 3.26: Horizontal cross sections of temperature anomaly from cases 108 & 118 - 122 at 600 km depth in the model. Each case is labelled with the Clapeyron slope of the phase change at 410 km depth. All quoted Clapeyron slopes are in MPaK$^{-1}$.
Figure 3.27: Horizontal cross sections of temperature anomaly from cases 108 & 118 - 122 at 800 km depth in the model. Each case is labelled with the Clapeyron slope of the phase change at 410 km depth. All quoted Clapeyron slopes are in MPaK$^{-1}$.
Since there are suggestions that with low magnitude negative Clapeyron slope at 660 km and a large positive Clapeyron slope at 410 km that there may be some convection enhancing effects it is worth considering the mid-mantle temperature anomalies. In figure 3.28 I show the horizontal temperature anomalies at 1400 km depth. At this depth any effect including an exothermic phase change may have upon the convection are not immediately obvious. For the most part the down-welling material is strikingly similar between each of the considered models. There are some minor differences, but the primary observation at this depth is the similarity between each of the simulations. Overall these simulations suggest that endothermic phase changes with weak, negative and larger magnitude exothermic phase changes there may be some small effects on the convective pattern but overall there is very little effect on the larger scale.

Finally for this sub-chapter on phase transitions, I have produced the radial mass flux profiles from each of the simulations containing both a 660 km deep endothermic phase change and a 410 km deep exothermic phase change. Figure 3.29 gives the radial mass flux profiles for case 109 and the five cases with increasing Clapeyron slope at 410 km depth. Based upon the temperature fields shown in figures 3.23 to 3.25 it is expected that the mass flux profiles will not be overly different from the reference case. This is generally the case, for most values of Clapeyron slope at 410 km there is very little change in the mass flux across transition zone layers. The only noticeable difference is at the largest Clapeyron slope at 410 km depth Clapeyron slope (8 MPaK$^{-1}$, case 117) where there is a slight increase in the mass flux at 410 km depth. Despite the slight increase it appears to have very little effect on the overall convection. There is no change in the mass flux across the upper mantle to lower mantle boundary.

Figure 3.30 shows similar plots for the five cases based upon case 108. The five cases have the same set of 410 km Clapeyron slopes as the set above. Again there is very
Figure 3.28: Horizontal cross sections of temperature anomaly from cases 108 & 118 - 122 at 1400 km depth in the model. Each case is labelled with the Clapeyron slope of the phase change at 410 km depth. All quoted Clapeyron slopes are in MPaK$^{-1}$.
Figure 3.29: Graphs of radial mass flux against depth for five cases based upon a simulation with a Clapeyron slope of \(-4 \text{ MPa} K^{-1}\) at 660 km depth. Clapeyron slopes at 410 km range from 0.5 to 8 MPaK\(^{-1}\). Mass flux is calculated at present day after 300 million years of mantle circulation modelling.

little difference in radial mass flux across each of the layers from one case to another. There is less evidence of an increase in mass transfer across the 410 km boundary than in the above cases and once again there is very little effect on convection deeper in the mantle. It is fairly evident from the simulations in this section that including a positive Clapeyron slope phase change at 410 km depth has very little effect on convection in mantle convection simulations using TERRA. I have shown that there is little change in the model temperature output both in the transition zone and just below it. There is also very little impact on the
Figure 3.30: Graphs of radial mass flux against depth for five cases based upon a simulation with a Clapeyron slope of $-2 \text{ MPaK}^{-1}$ at 660 km depth. Clapeyron slopes at 410 km range from 0.5 to 8 $\text{ MPaK}^{-1}$. Mass flux is calculated at present day after 300 million years of mantle circulation modelling.

deeper mantle where mantle circulation studies are compared to subducted material imaged in seismic tomography. It is possible that phase changes with larger Clapeyron slopes than considered here could affect the convection in a more severe way. However it is highly unlikely that phase changes with magnitudes that large would be possible in the Earth.
3.5.3 Phase Change Discussion

To understand the significance of mantle circulation models with assimilated plate motions and an endothermic phase transition at 660 km depth I compare my simulations with those in similar studies such as Wolstencroft & Davies (2011) and Yanagisawa et al. (2010). Figure 3.31 (courtesy: M. Wolstencroft) shows the simulations considered in section 3.5.1 (red symbols) plotted alongside the data of Wolstencroft & Davies (2011) (black symbols). It should be noted that due to slight differences in heating parameters the $\frac{Ra}{Ra_H}$ ratio is not identical to their simulations. It is, therefore, not possible to directly compare the runs, but plotting the data simultaneously is illustrative, and provides a tool to discuss some implications of the simulations. Although these simulations are all at the same Rayleigh number, thus plot in a single column on the graph, it’s possible to understand the differences between the two different classes of simulation. At $Ra = 2 \times 10^5$ I classify simulations with a Clapeyron slope of between -0.5 and -4 $MPaK^{-1}$ as whole mantle convection. Whilst simulations with Clapeyron slopes from -8 to -32 $MPaK^{-1}$ as transitional or partially layered. The first simulation to be classified as partially layered at this Rayleigh number (Case 110) matches identically with a simulation from Wolstencroft & Davies (2011) and plots on their trend line between whole mantle and transitional convection. Simulations incorporating plate motion history do not (within the parameter space considered here) cross from the partially layered convection regime to a two layer system as observed by both Wolstencroft & Davies (2011) and Yanagisawa et al. (2010). This new data suggests that simulations assimilating plate motions produce models matching well with the curve defining the boundary between transitional and whole mantle convection. It is difficult to produce a model with two layer mantle convection with a realistic Clapeyron slope and driven by plate motions at the surface. The data presented here are all simulations at a single Rayleigh number, therefore the study is limited by a lack of simulations at more Earth-like
Rayleigh number. Given that the simulations match well to the boundary between whole mantle convection and transitional convection derived by Wolstencroft & Davies (2011) it is probably not unfeasible that further simulations at higher Rayleigh number would match well with the published curves. What is unclear, however, is whether higher resolution, higher Rayleigh number simulations would generate models with convection in two independent layers given a large enough negative Clapeyron slope at 660 km depth.

Zhong & Gurnis (1994) demonstrated that in an isoviscous mantle at Ra = 2\times10^6 a phase transition with Clapeyron slope equivalent to -4.2 MPaK\(^{-1}\) results in clearly layered convection. For similar simulations with surface plates they found that smaller plates had
little effect on the convection pattern after being stalled by the phase transformation. However, for long plates they suggest that plate scale features in the mantle are not too different between models with and without an endothermic phase change, suggesting that larger plates break the layering structure. This is a similar effect to the partially layered cases described throughout this chapter. It seems that plate tectonics plays a fundamental role in how an endothermic phase transition can layer a system. In this study I have shown that in cases that would be layered under a free-slip surface are only partially layered when plate tectonics are used as the surface velocity boundary condition.

In simulations with extremely large negative Clapeyron slopes layering is strong in large parts of the globe with local, catastrophic breakdowns in the layering process. Figure 3.32 illustrates this catastrophic breakdown in layering in the three dimensional shell. I plot a vertical cross section at 90° from the Greenwich meridian showing the absolute model temperature, I also include an isosurface of hotter than average material at the base of the mantle to illustrate plumes. The clear distinction between upper and lower mantle indicates highly stratified mantle. Both cases would clearly fall in the 2 layer convection regime if not for the large down-wellings located under long-lived subduction zones. Up-welling plumes are very obviously restricted by the endothermic phase transition in both cases. Although the layering must break down for up-wellings also, so that the model conserves mass. This is most evident in the lower right quadrant of case 111.

As suggested by Zhong & Gurnis (1994) the layering for down-wellings only breaks down in regions where large quantities of ocean floor are subducted. For example in the Farallon region (case 111) or under east Asia (case 112). The difference between the locations of the strong down-welling flows is striking. Although both clearly represent regions of long-lived convergence in UNIL (2009) plate motion history, penetrative down-wellings are not produced in both regions in each case. I speculate that up-wellings have some involvement
Figure 3.32: Spherical shell visualisation of present day absolute temperature from mantle circulation cases 111 & 112. These cases have large negative Clapeyron slopes of -16 and -32 MPaK$^{-1}$. Included on the figure is a north-south vertical cross section located at 90 degrees from the Greenwich meridian and in orange an isosurface representing mantle at 2500 K. Present day coastlines are included for reference.

in controlling which down-wellings can penetrate the 660 km discontinuity. For a large avalanche of down-welling material to occur it is hypothesised that a cold thermal instability must develop below the discontinuity. As it develops the negative buoyancy draws down the cold material from above to initiate the avalanche. It is possible that when up-welling plumes are located near to regions of high convergence the initial instability is unable to develop and the large down-welling does not develop. For each individual model case a new initial condition is developed in order to prevent numerical problems during the calculations. As plumes in the models are not conditioned by plate motions they do not appear in the same locations between cases. Hence certain, but varying, regions of down-welling not breaking through the layering due to lower mantle up-welling plumes.

There may be some degree of thermal coupling between the upper and lower mantle
in cases with large, negative Clapeyron slopes (Čížková et al. 1999, e.g.). It is suggested that given an impenetrable interface thermal coupling between the two layers could produce lower mantle downwellings despite having zero mass flux across the boundary. The results of Wolstencroft & Davies (2011) demonstrate little evidence for downwellings caused by thermal coupling in cases with the largest, negative Clapeyron slopes. However, it is a proposed mechanism for catastrophic, avalanche like breakdown of layering in transition regime models. In simulations presented in this chapter there is likely to be a component of thermal coupling across the lower mantle to upper mantle boundary in cases with particularly large, negative calpeyron slopes, however, figure 3.22 clearly demonstrates that even with a Clapeyron slope of $-32 \text{ MPaK}^{-1}$ there is still a component of mass flux across the boundary. It may be difficult to distinguish thermal from mechanical coupling in these cases, particularly if an initial lower mantle down-welling is generated by thermal coupling across the boundary, which then develops into a full coupled avalanche style down-welling. If, as suggested above, hot up-wellings are partially responsible for restricting the locations of down-wellings, then there is evidence for at least some thermal coupling. The results presented here remain different from those presented in Wolstencroft & Davies (2011) and citetYana:2011 suggesting that plate tectonics still has a significant role in the convection mode of the mantle.

While it appears that it would be impossible for Earth to exist in two-layered state of mantle convection while plate tectonics has been active, it is not possible to rule out the possibility of layering prior to the onset of plate tectonics. Since Rayleigh number has a clear effect on the behaviour of phase changes it is also possible that if early convection on Earth was more vigourous than present then layering may have occurred. This raises a number of questions which are not possible to answer here but which could prove to be an exciting extension of this work. Questions such as:
1. Is it possible to model a mantle with both plate tectonics and a two layered convective regime

2. Does the on-set of plate tectonics provide the initial break down of a two layer convecting mantle?

3. Can up-welling plumes in the lower mantle prevent long-lived down-wellings breaking through the layer boundaries

### 3.6 Other factors affecting mantle circulation models

#### 3.6.1 The effect of compressibility

I have also considered the effect of compressible convection upon mantle circulation models. I compare a further run, based upon case 103 but including compressibility via a Murnaghan equation of state (Murnaghan 1951). The depth dependent parameters included in the compressible equation of state are as in Bunge et al. (1997). Chapter 2 gives a more detailed description of compressible convection in TERRA. The basic parameters and the differences in depth dependent properties are summarised in table 3.11.

Figure 3.33 presents a comparison of the modelled temperature field at 900 km depth for the two cases described above, Case 103 the incompressible and Boussinesq case and case 132 the compressible, Murnaghan case. Overall there is very little difference between the two cases at this depth. Modelled down-wellings in the compressible convection case are, perhaps, more cylindrical. An effect most noticeable under the Americas. In the eastern hemisphere, modelled down-welling generally have lower amplitude in the compressible convection case.
Table 3.11: Model parameters used for simulations comparing incompressible (case 103) convection to compressible (case 132) convection with a Murnaghan equation of state

At 1400 km depth horizontal maps of temperature anomaly demonstrate similar features to at 900 km. Figure 3.34 presents two horizontal temperature anomaly plots for the incompressible and Boussinesq case (case 103) and the compressible, Murnaghan case (case 132) at 1400 km depth. At this depth there is a smaller volume of colder than average material in the compressible convection case than in the incompressible case. As at 900 km anomalies are more dispersed in the compressible convection calculation. Overall the patterns remain very similar with colder than average mantle underlying regions of long-lived tectonic convergence.

Overall the two cases do not have too many noticeable differences, either would be suitable use in further work on mantle circulation models. For later models I tend to chose compressible convection, as parameters such as the coefficient of thermal expansion, bulk modulus and density are likely to have some pressure and depth dependence in the
real Earth. Compressible convection more accurately captures some aspects of the depth dependence of mantle properties.

3.6.2 The effect of heating mode

Finally, I investigate the effect of heating mode upon the nature of convection developed within mantle circulation models. Until now all cases have contained basal heating and internal heating to mimic heat flux from the core due to secular cooling and heating from radioactive decay respectively. Here I investigate a further two cases based up case 103, one case contains no basal heating, achieved by implementing an insulating core mantle boundary, the second contains no internal heating. Table 3.12 summarises the differences between the cases and the reference case. All other parameters are as in case 103.

Case 133 is included to model a mantle with heating from the core only. This case contains a steady rate of heating from the core, by simulating an isothermal boundary

Figure 3.33: Horizontal cross sections at 900 km of temperature anomalies for cases investigating compressibility in mantle circulation models (a) incompressible and Boussinesq (b) compressible, Murnaghan equation of state.
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Figure 3.34: Horizontal cross sections at 1400 km of temperature anomalies for cases investigating compressibility in mantle circulation models (a) incompressible and Boussinesq (b) compressible, Murnaghan equation of state.

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<td>Case 134</td>
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</tbody>
</table>

Table 3.12: Heating parameter variations for cases investigating the effect of heating mode. Other parameters are as in Case 103

condition at the base of the model and has zero internal heating. Case 134 uses an insulating boundary condition at the base of the model to remove the effect of heating from Earth’s core and includes volumetric heating only. Internal heating is included at the same rate as all other simulations in this chapter. Bunge et al. (1997) considered the effects of including bottom heating on a standard mantle convection model using the code, TERRA. They note that the convection calculation with both heating modes contains axisymmetric up-wellings akin to plumes, when previous, internally heated models did not. The conclude that the effects arise from the predictable creation of a lower mantle thermal boundary layer.

As with all the other parameter variations presented in this chapter I consider
global, horizontal temperature anomaly maps for at two mid-mantle depths, 900 km and 1400 km. Figure 3.35 presents the maps for case 103, 133 and 134 at 900 km depth.

![Figures 3.35](image)

**Figure 3.35:** Horizontal cross sections at 900 km of temperature anomalies for cases investigating heating mode. (a) basally heated only, (b) internally heated only, (c) mixed heating mode.

There are a number of interesting differences between the three cases at 900 km depth. In the basally heated only case (figure 3.35a) up-welling plumes are similar in nature and magnitude to up-wellings in the reference case, however the down-wellings in that case do not have such large amplitudes as in the reference case (figure 3.35c). The extent of the anomalies in the Farallon and Tethys regions are also smaller in the basally heated only case.

The case containing only internal heating (figure 3.35b) is different to many of the models seen previously. The temperature planform is significantly different from the basally heated case. There is more power in the down-wellings, with higher amplitude anomalies...
beneath regions of active convergence. Cold temperature anomalies in these regions are also more slab-like in case 134. Anomalies are noticeably thinner and linear in nature particularly beneath South America. As with the basally heated case there is no obvious anomaly relating to Farallon subduction beneath North America as in the mixed heating case.

The most striking difference between the internally heated case and the other two cases are the hotter than average regions. In both the basally heated and mixed heating cases cylindrical up-wellings with temperatures at up to 600 K greater than the radial average are present, most likely corresponding to mantle plumes. In the internally heated case there is no evidence for similar cylindrical up-wellings. Instead there are broad regions of mantle around 200 K hotter than the radial average beneath the Atlantic and Indian Oceans and Africa.

Figure 3.36 presents horizontal temperature anomaly maps for the same three cases at 1400 km depth.

At 1400 km depth broadly the same pattern as above is observed. In the case with basal heating only (figure 3.36a) a number of plume like up-wellings are observed, distributed across the globe. These up-wellings are very similar in nature to in the mixed heating case (figure 3.36c). The observed weaker down-welling signature observed at 900 km depth is observed at 1400 km too, there is still colder than average material in regions below long lived subduction zones, however it is very low amplitude with only subduction beneath Indian and surrounding regions standing out at 1400 km depth.

For the case with just internal heating (figure 3.36b) the temperature planform is dominated by thin linear down-wellings, which are particularly prominent beneath South America, and the entire Tethys region from the Mediterranean to the Bay of Bengal. There are also cylindrical down-wellings beneath the Philippines and Japan. Like the temperature anomaly map for case 134 at 900 km depth there is no evidence for cylindrical, plume
Figures 3.35 and 3.36 clearly demonstrate that the heating mode plays a significant role in mantle circulation. Like Bunge et al. (1997) suggest including bottom heating in the model significantly alters the nature of up-wellings in the mantle. The result is intuitive as developing a thermal boundary layer at the base of the mantle would allow for instabilities to develop, resulting in plumes of hot material.

Although the internally heated only cases produces sharper more slab-like down-wellings, the mixed heating case has an overall more Earth-like nature to convection, with cylindrical up-welling plumes instead of broad regions of warmer than average mantle. In the basally heated only case down-wellings are less slab like, and so do not match tomography
so well. For these reasons the mixed heating mode remains the preferred model for heating mode.

3.7 Concluding remarks

The parameter space investigations throughout this chapter have shown that mantle circulation models are highly sensitive to certain parameters, such as radial viscosity variations and endothermic phase transitions. Other factors such as temperature dependent viscosity and exothermic phase transitions have considerably less effect. I have shown that mantle circulation models can be tuned to produce down-welling temperature anomalies that very closely resemble the mantle structure revealed by seismic tomography. Although these simulations are conducted at Rayleigh numbers less than might be expected for Earth the effects should be similar once scaled up to Earth-like conditions. In some instances, for example phase transitions, it is likely that the increasing Rayleigh number would mean lower values than chosen here would produce better matches to seismic structure, while other factors, such as viscosity increases into the lower mantle would remain similar. In simulations at higher-resolution the reference viscosity can be lower, resulting in convection at more Earth-like Rayleigh number. Although the radial viscosity increase would still be significant, the generally lower viscosity will change the nature of convection too. In simulations of this nature temperature dependent viscosity could be more important as the overall lower mantle viscosity would not become too large.

Including a single endothermic phase change at 660 km depth can alter convection from a single layer, whole mantle convective regime to a partially layered regime. With a sufficiently large, negative phase transition, hot up-wellings are restricted at 660 km depth, often not passing from lower to upper mantle at all. Down-wellings driven by plate tectonics
at the surface are also restricted by larger, negative phase transitions, however, these are not completely restricted from passing through the boundary. At very large, negative Clapeyron slope some down-wellings are completely restricted by the phase transition, while others pass through, suggesting localised, catastrophic breakdown of layering, these conditions are extremely unlike the real Earth. This work may help to understand hypothesised changes in early Earth convective regimes, with the onset of plate tectonic contributing to the breakdown of a two-layered convective system.

Including bottom heating as well as internal heating to mimic both secular cooling of the core and heat production from radioactivity generates models with Earth-like linear down-wellings, resembling slabs and cylindrical up-wellings resembling plumes. Models with no bottom heating do not generate a thermal boundary layer at the base of the model, and therefore no plumes are produced. These models do generate sharper down-welling features.

In the next chapter of this thesis I will look at higher-resolution mantle circulation models in more detail for the Tethys region. This parameter space investigation has provided constraints on key parameters for input into the high resolution model. It would not be possible, due to available computing resources, to produce large numbers of models at the higher resolution. Parameters for the higher resolution models will be selected based upon results from this chapter, including a viscosity increase of thirty times into the lower mantle, a modest amount of temperature dependence and mixed heating mode.
Chapter 4

Mesozoic fossil subduction zones in global mantle circulation models

4.1 Introduction

Mantle circulation models incorporating recent plate motion history as the surface velocity boundary condition have been used to model a mantle with colder than average regions from subduction closely resembling the distribution of fast seismic velocity areas imaged by tomography (Bunge et al. 2002). By assimilating 120 million years of plate motion history from Lithgow-Bertelloni & Richards (1998) into a mantle convection model Bunge et al. (2002) demonstrated that mantle models could accurately generate downwellings matching imaged fast seismic regions in tomography beneath South Asia and North America. In chapter 3 of this thesis I demonstrated how such models and their comparison to seismic tomography varies by altering some fundamental mantle properties within the model. Broadly the models produce temperature fields with colder than average material in the correct geographic locations but the quality of the match to seismic tomography is
variable even across small changes in parameters. Like Bunge et al. (2002) these models were at a resolution where convection is at a Rayleigh number slightly lower than Earth-like. Using the most sophisticated version of the convection code TERRA and large cluster computers it is possible to simulate Earth’s mantle at more Earth-like Rayleigh number. By scaling up the simulations to Earth-like Rayleigh number, incorporating the best model parameters chosen from the study in chapter 3 and using longer plate motion history it is possible to recreate some of the higher resolution, more regional details of mantle convection. However, simulations at higher resolution require significantly larger amounts of computing resources therefore large scale parameter space investigations are more difficult to undertake. For this chapter I focus in particular on a few global models looking in depth at the Tethys region where high-resolution seismic tomography studies image complex seismic structure at mid-mantle depths (Amaru et al. 2008, Chang et al. 2010, Li et al. 2008, Ritsema et al. 2011). Interpretations of these and other seismic studies attribute individual fast seismic anomalies to specific tectonic processes (Hafkenscheid et al. 2006, Replumaz et al. 2010, van der Meer et al. 2010, van der Voo et al. 1999b). Despite the connection being made between surface processes and deep mantle fast anomalies there is no dynamic link to confirm the interpretations. Mantle circulation models can provide that link between tectonics and the deep mantle. The combination of higher resolution, higher Rayleigh number circulation simulations with the plate tectonic reconstructions of UNIL (2009) provides a significant improvement on most previous mantle circulation models.

In this chapter I will consider higher resolution, three-dimensional, spherical geodynamic mantle models combined with plate tracking marker particles. These particles are used to directly demonstrate the tectonic processes responsible for the imaged tomographic anomalies. In this chapter I will consider the previous interpretations of seismic tomography, briefly recap the modelling methods appropriate to this study and examine a
variety of tomography studies for robust regional anomalies in the Tethys region. Finally I present the results of the simulations and discuss what can be learned both about the mantle and tectonic history in the region.

4.1.1 Tethyan tectonics interpreted from seismic tomography

The motivation to consider the Tethys region is primarily driven by a number of seismic tomography studies of the region revealing mantle structure possibly linked to the complex tectonics in the region (Hafkenscheid et al. 2006, Replumaz et al. 2010, van der Meer et al. 2010, van der Voo et al. 1999). Although mantle structure in the Tethys region is complex suggesting varied subduction history, most interpretations of tomography are not too dissimilar. Four main fast anomalies are interpreted as the present day seismic signature of multiple phases of Tethys Ocean subduction. One possible interpretation of the seismic tomography from Hafkenscheid et al. (2006) is reproduced in figure 4.1. The figure demonstrates the four regions of faster than average mantle as imaged by tomography. Three of the four are at mid-mantle depths (1000 km - 2000 km) whilst the fourth is shallower, between the surface and around 600 km. These fast anomalies are robust across a number of tomography studies including P and S wave data and a variety of inversion techniques. The deepest anomaly is in the north of the study area. It exists from 1000 km to 1600 km depth or deeper. Hafkenscheid et al. (2006) interprets this anomaly as subducted Palaeotethys lithosphere and is given the name PT (Palaeotethys). The same anomaly is named I by van der Voo et al. (1999) and is attributed to the last remnant of Palaeotethys (also known as Mesotethys) subduction between the Lhasa block and the margin of Asia. Hafkenscheid et al. (2006) notes that in her interpretation the anomaly has considerably less volume than in van der Voo et al. (1999) and suggests the oldest portions of Palaeotethyan subduction are not detectable by tomography.
**Figure 4.1:** Schematic cross sections of Bijwaard & Spakman (2000)’s tomography model by Hafkenscheid et al. (2006). Left figure shows limits of faster than average material; IO is Indian Ocean, IC is Indian Continent, Hi is Himalaya and PT is Palaeotethys. Right figure is the interpretation of Hafkenscheid et al. (2006) in terms of tectonic processes; NT is Neotethys, PT is Palaeotethys and SPT is Spongtang. Figures represent the same geographic region. The vertical axis is depth from the surface and the transect is in degrees north of the equator.

The major anomaly in terms of volume and lateral extent is to the south of PT or I. It exists from 900 to 1600 km depth. It is a linear north-west to south-east trending feature present in four interpretations of tomography. In Hafkenscheid et al. (2006) splits this into two close but distinct regions, IC (Indian continent) and Hi (Himalaya), with IC being the deeper part and Hi being the shallower part. They hypothesise IC to be subduction of old Neotethys ocean beneath the margin of Eurasia, whilst Hi is the subduction of the Spongtang back-arc ocean in the same location. van der Voo et al. (1999b) treat this as a single anomaly, named II. It is interpreted as one of two Neotethys subduction zones of Mesozoic age. Replumaz et al. (2010) interprets this anomaly as simply subduction of Tethys lithosphere. Finally a large anomaly at mid-mantle depth beneath the present day Indian continent is present in the paper by van der Meer et al. (2010) but is not named and
Further south again is a third anomaly at similar mid-mantle depths. It is commonly attributed to a second phase of Neotethys subduction behind a back-arc ocean formed by earlier subduction at the Eurasian margin. Hafkenscheid et al. (2006) names this anomaly IO (Indian Ocean) and suggests the subduction of Neotethys lithosphere behind the Spongtang back-arc basin. It is noted that locally slow subduction would be required for this anomaly to exist at the shallowest depths at which it is observed, this is consistent with observations by van der Voo et al. (1999b) and Replumaz et al. (2004). In van der Voo et al. (1999b) this anomaly is named III and is given as stretching from the eastern Mediterranean under the Arabian peninsula to the southern tip of India. They suggest northward intra-Tethyan subduction may be responsible for the southern anomaly. van der Meer et al. (2010) highlight a Maldives slab (Md) below the north-western Indian Ocean. They agree with other interpretations suggesting Md is the result of northward subduction of Neotethys lithosphere beneath the Spongtang Ocean.

The shallower anomaly in the north of the region is called IV by van der Voo et al. (1999b) and although not explicitly mentioned, is thought to be a result of post-collisional processes as India continues to indent into Asia. This interpretation is slightly different to Hafkenscheid et al. (2006) and van der Meer et al. (2010) where the deeper part of this region corresponds with the shallowest part of the large, central anomaly. Both papers identify a Himalaya (Hi anomaly) and interpret it as the subduction of the Spongtang ocean post the arc accreting onto the Indian content. This is essentially subduction of the youngest part of the Neotethys Ocean prior to the collision of India with Asia. Hafkenscheid et al. (2006) also identify a small, very shallow anomaly underneath the Hindu Kush (HK) which is likely to be the more recent, post-collisional processes. Replumaz et al. (2010) has a broadly similar interpretation with near vertical northward subduction down to 500 km, an anomaly also
named HK and a southward dipping anomaly, India (IN) between 500 and 800 km similar to the Hi anomalies.

Broadly the interpretations of these tomography studies agree well. There are small differences in the interpretations. For example the right hand panel of figure 4.1 shows Hafkenscheid’s interpretation of the tomography study suggesting that the IO and IC anomalies originate from the same, probably intra-ocean subduction zone. For this situation to occur the subduction zone would have to begin at the Eurasian margin followed by a step southwards to a subduction zone behind the back arc. Finally after the accretion of the arc onto the northward moving continent, subduction briefly returns to the Eurasian margin. The alternative scenario is best illustrated by a figure from van der Voo et al. (1999b). The figure is reproduced in figure 4.2

Figure 4.2: Left: Schematic cross section of tomography by van der Voo et al. (1999b). Faster than average material is represented by shaded areas; I is interpreted as Meostethys subduction, II as subduction of Neotethys ocean or back-arc ocean at the Eurasian margin, III as intra-Tethys subduction and IV as post-collisional processes. Right figure is a schematic cross section of the tectonic processes required to produce these fast anomalies.
This figure illustrates van der Voo’s interpretations of tomography on the left hand side, on the right hand side is the interpreted subduction history. In this case it is suggested that simultaneous subduction behind the intra-ocean arc and at the Eurasian margin is required to produce the two anomalies III and II. This differs from the interpretation in figure 4.1 as it has anomaly IC/II originating from a different surface convergent margin. These interpretations of seismic tomography contain no dynamic information to link the surface processes to the deep mantle features. Using mantle circulation models with plate tracking markers can be a test of these interpretations. Given appropriate plate tectonic reconstructions it should be possible to test the above interpretations and even to distinguish between some of the small differences between interpretations.

### 4.1.2 Tectonic reconstructions in mantle models

As throughout this thesis the plate motion history included as the surface velocity boundary condition is the UNIL (2009) plate motion history (Ferrari et al. 2008, Hochard 2008, Stampfli & Borel 2002, 2004, Stampfli & Hochard 2009). The UNIL (2009) tectonic reconstructions span a larger timeframe than those used in previous mantle circulation studies (Lithgow-Bertelloni & Richards 1998). Reconstructions have complete global coverage for 300 million years instead of 120 million years used in Bunge et al. (2002) providing more history of the Tethys closure and more time for the model to develop from its initial condition. The UNIL (2009) reconstructions also have more plates per reconstruction including a series of smaller plates as well as the 10 to 12 major plates. The greater detail incorporated in the UNIL (2009) tectonic history should allow mantle circulation models to recreate some of the highly detailed mantle structure described in section 4.1.1. In the Indian region the history contains a wealth of detail including many of the tectonic features hypothesised from interpretations of seismic tomography. From 300 Ma the Palaeotethys ocean closes, with
northward subduction of the ocean at the plate boundary with Eurasia. As the Palaeotethys ocean closes a fragment of continent is rifted from the main super continent in the west. The rift continues to develop into the Neotethys ocean. Palaeotethyan subduction continues at the Eurasian margin until the continental fragment, Cimmeria, collides with the main body of Asia at around 210 Ma. During this time spreading between the super-continent and the Cimmerian fragment continues, the Neotethys ocean grows to take up much of the area of the now entirely subducted Palaeotethys ocean. After the collision the second phase of subduction begins, again at the Eurasian margin. This second phase is the subduction of the oldest Neotethys ocean, sometimes referred to as Mesotethys subduction to distinguish it from the more recent period of Neotethys subduction. Images of the global coastlines and tectonics from the reconstructions for these time frames are shown in figure 4.3. Subduction of the oldest Neotethys ocean continues to shortly after 155 Ma when India rifts from the supercontinent, Gondwana. India then moves rapidly northwards as the Neotethys ocean is subducted. Subduction continues at this margin until the major collision of the Indian continent with Eurasia at around 55 Ma. During the closure of the Neotethys ocean a back-arc basin, named the Spongtang Ocean opens up between Eurasia and the Neotethys from 142 Ma. This results in a second, intra-ocean convergent margin as Neotethys lithosphere is subducted beneath the Spongtang Ocean. During this time India continues to move northwards, initially colliding with the back-arc and eventually with the Asian continent. After the complete subduction of the Neotethys and the accretion of the Spongtang backarc the Indian landmass continues to indent into Eurasia. The northward motion is ongoing at present day. This convergence is included in the tectonic reconstructions. Two examples of tectonic reconstructions are included in figure 4.4. In this figure the geographic region is restricted to the Tethys realm to demonstrate the detail in this area. The tectonic reconstructions and the mantle circulation model remain global and have consistent lateral
Figure 4.3: Example reconstructions from the UNIL (2009) tectonic histories illustrating:
(a) closure of the Palaeotethys ocean as the Cimmerian continental fragment rifts off Gondwana and moves northwards. (b) Oldest Neotethys subduction after the collision of Cimmeria with Asia. These figures are generated from maps provided by C. Hochard.
resolution despite the focus being on one particular region.

### 4.1.3 Previous work

Prior to this work there have been few studies of the region at high enough resolution to capture the details observed in Tethyan mantle. A simple two dimensional model by Jarvis & Lowman (2005) attempted to reproduce the schematic overview of Tethyan mantle described above. Using a simple linear approximation of the subduction history of the Tethys oceans as a boundary condition, they show that the three mid to deep mantle seismic anomalies can be modelled in this simple case. The work does, however, pose questions about ‘slab survival times.’ In the model Jarvis & Lowman (2005) note that oldest subducted material, from a subduction zone which became extinct more than 140 Ma does not remain at mid-mantle depths as might be expected from imaged slabs in tomography. Even with a viscosity increase of 30 times into the lower mantle the slab corresponding to van der Voo et al. (1999b)’s anomaly I and Hafkenscheid et al. (2006)’s PT anomaly sinks to close to the core-mantle boundary. In figure 4.5 I reproduce the results of Jarvis & Lowman (2005)’s mantle model at present day. The figure shows that it is possible to generate three cold slabs from this plate motion history. It also shows how the oldest subduction which terminated 140 million years ago produces a slab which sinks to the core mantle boundary. The third anomaly is smaller than imaged in tomography and does not reach the depths indicated by tomographic images.

A recent article by Zahirovic et al. (2012) undertook a similar study to the one presented here. They tested two subduction histories for the region: A conventional model containing continuous Neotethys subduction at the boundary of the Eurasian continent. The alternative model contains the hypothesised back-arc ocean and therefore two phases of subduction. The results of Zahirovic et al. (2012) suggest Greater India collided with a
Figure 4.4: Example global reconstructions from the UNIL (2009) tectonic histories illustrating: (a) Neotethys subduction after a back arc (Spongtang) ocean has opened off the Eurasian continent. (b) The post collision motion of India indenting into Asia. These figures are generated from maps provided by C. Hochard.
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Figure 4.5: *The present day results of Jarvis & Lowman (2005). Illustrating that a 2D model with appropriate plate motion history can model a mantle with three distinct 'fast' anomalies as imaged by seismic tomography.* Figure reproduced from Jarvis & Lowman (2005).

Neotethyan back arc basin prior to the collision between India and Asia. The study follows a similar approach to the work in this chapter, using three dimensional, spherical models of mantle convection with imposed plate motions as the surface velocity boundary condition. Two examples of the results of this study are presented in figure 4.6. The figure shows the slab contours of the conventional model in red and the alternative model in green. The background shows the contours of two seismic tomography studies, GyPSuM-P (Simmons et al. 2010) and P-mean (Becker & Boschi 2002). The general consensus of the results presented is that the models re-create the imaged seismic structure well. To achieve a better match the 'alternative' model, containing intra-ocean subduction is preferred. This subduction scenario produces a better match in the south of the region.

The models presented in this chapter contain some significant improvements on previous studies. Passive tracers are used to further demonstrate the link between tectonic processes and the deeper mantle. Furthermore, I do not impose any slabs in the upper mantle, allowing the down-wellings to naturally develop based upon the surface boundary
Figure 4.6: The present day results of Zahirovic et al. (2012). Showing the slab contours of two alternative subduction histories overlain on seismic tomography contours. Figure reproduced from Zahirovic et al. (2012)

c-condition.
4.2 Summary of relevant methods

Mantle circulation is simulated using benchmarked three-dimensional, spherical mantle convection code, TERRA (Baumgardner 1985, Bunge et al. 1997, Davies & Davies 2009, Yang & Baumgardner 2000). Given currently available supercomputing resources TERRA can simulate the mantle at Earth-like Rayleigh numbers and incorporate 300 million years of plate motion history as the surface velocity boundary condition. The code solves the three principal equations of fluid dynamics (Stokes flow), conservation of mass, momentum and energy in a three dimensional spherical domain. The equations are solved using a finite element method on a fixed icosahedral grid made up of up to 129 individual shells at the highest resolution. To refine the grid further great circle arcs repeatedly divide the 20 triangles that make up the icosahedron. This leaves an evenly spaced mesh with mid-mantle resolution of around 23km in each direction at the highest resolution. Whilst the simulations presented here focus on the Tethys region at mid-mantle depths it is important to note the calculations are performed on a global mesh with constant lateral resolution throughout.

Boundary conditions for this simulation are isothermal and free slip at the core mantle boundary (CMB). The surface boundary condition is also isothermal, but velocity is prescribed from plate reconstructions. The mantle is assumed to be a compressible, Newtonian viscosity fluid with no mineral phase changes. Compressibility is incorporated through the Murnaghan equation of state (Murnaghan 1951) and uses parameter values from Bunge et al. (1997). Key model parameters are presented in table 4.1. Temperature dependent viscosity is included according to the law:

\[ \eta(T) = \eta_0 \exp[-E_a T] \] (4.1)
where temperature is non-dimensionalised by $\Delta T$ and $E_a$ is the fundamental variable controlling the degree of temperature dependence in the model. In this case $E_a$ is chosen to give a variation of two orders of magnitude. Prior to assimilating plate motions an initial condition is produced to attempt to mimic unknown mantle conditions at the start of the simulation, 300 Myr before present. The method used to generate this initial condition is identical to that used for mantle circulation models throughout this thesis and as described in Bunge et al. (2002). Initially TERRA is run as a standard convection model starting from a small-scale random temperature distribution. For this phase both boundary conditions are isothermal and free slip. The convection calculation is continued until such a time that heat generation in the model from secular cooling of the core and from radioactive decay is roughly matched to that exiting the model domain at the surface. A second phase of modelling is added to the simulation to generate a distribution of mantle density and temperature anomalies that may represent unknown mantle structure from 300 million years ago. As this is impossible without plate motion history prior to 300 Ma the simulation is run for approximately 100 Myr with the oldest available plate reconstruction to generate the initial condition.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reference viscosity</td>
<td>$2 \times 10^{24}$ Pa s</td>
</tr>
<tr>
<td>Lower mantle viscosity</td>
<td>up to $6 \times 10^{22}$ Pa s</td>
</tr>
<tr>
<td>Temperature dependent viscosity, $E_a$</td>
<td>4.61</td>
</tr>
<tr>
<td>Mineral phase changes</td>
<td>None</td>
</tr>
<tr>
<td>Equation of state</td>
<td>Compressible Murnaghan</td>
</tr>
<tr>
<td>Thermal conductivity</td>
<td>$4 \ W m^{-1} K^{-1}$</td>
</tr>
<tr>
<td>Internal heat production</td>
<td>$4 \times 10^{-12} \ W m^{-3}$</td>
</tr>
<tr>
<td>Surface temperature</td>
<td>$300 \ K$</td>
</tr>
<tr>
<td>CMB temperature</td>
<td>$3800 \ K$</td>
</tr>
<tr>
<td>Velocity boundary condition</td>
<td>300 Myr plate motions</td>
</tr>
</tbody>
</table>

Table 4.1: Key model parameters for the simulations presented in this paper
Once a satisfactory initial condition has been generated a forward modelling approach is used to predict present day mantle conditions. In this phase, surface plate motions are assimilated in one of two ways: The first is to prescribe the velocities for each reconstruction for the fixed time for that reconstruction. In the 300 Myr period there are 29 discrete reconstructions. Temporally reconstructions are non-uniformly spaced, with a higher number of reconstructions closer to present day. Using this method each surface node is allocated to a single plate from the given reconstruction and each plate at each time slice has a rotation pole and vector, which is used to prescribe the surface velocities. Any grid node not on the surface layer is allowed to take the velocity of the solution. This method has the advantage of having true plates at all times, however at transitions between stages may have large jumps in the location of plate boundaries and the direction of velocity vectors.

The second method is to linearly interpolate velocities between the current reconstruction and the next. At times corresponding exactly to a given reconstruction velocities are prescribed as above. In the times between reconstructions a linear interpolation is applied to both the location of the rotation pole and the magnitude of rotation vector independently. Thus the pole for each plate moves along a great circle arc between its location in the current reconstruction and the location of the corresponding pole in the next reconstruction. The second method has the advantage of not having sharp spatial movement in the locations of plate boundaries and rotation poles due to the rapid change in plate configuration between reconstructions. This method loses some of the plate-like nature of the reconstructions due to the interpolation.

In addition to the above, passive marker particles are linked to three of the surface plates. These particles have no physical properties; they only follow the flow within the domain. Markers are used to track the position of the Indian plate, Spongtang ocean plate and the Neotethys ocean plate into the mantle revealing the location of subducted material.
on those plates at present day. The chosen plates are based upon the naming convention in the UNIL (2009) plate motion history. Here I briefly explain which tectonic process each of the plates represent through geological history. The Neotethys ocean plate refers to the oldest portion of the Neotethys ocean (also known as Mesotethys) that exists from 210 Ma to 142 Ma. The Spongtang ocean is the back arc basin which exists from 142 Ma until the collision of the India landmass and the arc. The Indian plate is a little more complicated and is the portion of the Neotethys ocean subducting northwards under the Spongtang ocean between 142 Ma and 55 Ma. After the collision between India and the arc, India refers to final phase Neotethys subduction at the Eurasian margin. After 48 Ma India refers to the northwards motion of India indenting into Asia. Figure 4.7 illustrates map views of the particle initiations from 180 Ma through to present day. Throughout the chapter red particles originate on the Indian plate, green particles on the Spongtang oceans and blue particles represent the oldest part of the Neotethys ocean. The marker field can be compared to the temperature field, and thus the seismic tomography to demonstrate which anomalies arise from which surface processes.

Furthermore, the calculated temperature from the mantle circulation model can be converted to seismic velocity using data from mineral physics experiments using a thermodynamic approach (Stixrude & Lithgow-Bertelloni 2005, 2011). Density and elastic parameters are calculated using the code PerPleX (Connolly 2005) and are corrected for the effects of anelasticity (Cobden et al. 2009, Goes et al. 2004). As the models presented here are isochemical, a pyrolitic composition is assumed for the conversion. This conversion is useful to provide a more direct comparison between the model predictions and the seismic tomography data sets. It is noted that the model predictions once converted to seismic velocities over estimate the magnitude of velocity perturbations when compared to seismic tomography (Schuberth et al. 2009a, Schuberth et al. 2009b). Once the model predictions have been
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The images show a series of maps representing the subduction zones at various times in the past. The times shown are 180 Ma, 165 Ma, 142 Ma, 121 Ma, 95 Ma, 84 Ma, 57 Ma, 48 Ma, 33 Ma, and 0 Ma. The maps illustrate the movement and changes in the subduction zones over time, providing insights into the geological history and mantle circulation patterns.
converted to seismic velocity the resulting velocities can be passed through the resolution operator of the S40RTS tomography study (Ritsema et al. 2011). Filtering the model this way accounts for the differences in resolution between mantle circulation models and seismic tomography (Davies et al. 2012, Styles et al. 2011). Applying a tomographic resolution filter to model predictions has also been shown to reduce the differences in amplitude observed between the model predictions and actual tomographic values (Schuberth et al. 2009a). The limitation of this filtering is that the only valid comparison is to S40RTS, nonetheless it is still instructive.

Three models are initially considered in this chapter, they are described in table 4.2. These three cases take into account the differences in applying plate motion history and the effects of resolution and filtering the model to match tomography.

<table>
<thead>
<tr>
<th>Case</th>
<th>Lateral resolution</th>
<th>Plate motion</th>
<th>Converted &amp; filtered?</th>
</tr>
</thead>
<tbody>
<tr>
<td>Case 001</td>
<td>23 km</td>
<td>29 discrete stages</td>
<td>no</td>
</tr>
<tr>
<td>Case 002</td>
<td>23 km</td>
<td>Interpolated</td>
<td>no</td>
</tr>
<tr>
<td>Case 003</td>
<td>23 km</td>
<td>Interpolated</td>
<td>yes</td>
</tr>
</tbody>
</table>

**Table 4.2:** A summary of the three cases presented in this chapter. Highlighting the differences between the cases with very similar model parameters.

**Figure 4.7 (previous page):** Examples demonstrating the initiation of particles on the surface layer of the model based upon the plate motion history. Red particles originate on the Indian plate, green on the Spongtaq back arc ocean and blue on the oldest Neotethys/Mesotethys ocean. Grey areas represent palaeo-continents.
4.3 Results

4.3.1 Plate motion history, resolution and filtering

Figure 4.8 considers the effect of different methods of applying plate motion history as well as the effect of resolution on the outcome of the model at present day. The three models presented here all have similar mantle parameters but vary subtly (the differences are described in table 4.2). The geographical region considered in the study is between 15° and 135° east and −15° and 45° north. The data for the selected region is extracted from the global domain. The region is chosen to capture the effect of subduction of the Tethys oceans in the region around the Indian continent. Cases 001 and 002 present high-resolution mantle circulation models with identical physical properties; the only difference is the application of plate motion history. Case 001 is the simulation where plate motion is applied in discrete stages. Here reconstructions are abruptly changed at the end of each stage to the new plate configurations. Rotation pole locations and velocity vectors are also changed abruptly. Case 002 is the model with plate motions assimilated in a smoothed fashion. The rotation pole and magnitude are interpolated linearly between stages. Plate configurations are changed as in the first method. There is little obvious difference in the region between the two approaches. It is not easy to choose a preferred model from the two given the similarities between the models. Neither is likely to be a perfect fit to tomography. Since both models are similar it should not matter which method of prescribing the surface velocity boundary condition is chosen.

Case 003 considers the effect of converting model predicted temperature fields from case 002 to seismic velocity and filtering to the resolution of tomography. The conversion is done using the thermodynamic method mentioned above (Stixrude & Lithgow-Bertelloni 2005, 2011) and the calculated S-wave velocities are passed through the resolution filter of
Chapter 4: Mesozoic fossil subduction zones in global mantle circulation models

Discrete Stages

Interpolated Vel.

S4ORTS Filter

0700 km

0930 km

1070 km

1250 km

1380 km

1510 km
S-wave tomography model S40RTS (Ritsema et al. 2011), so as to see the model predictions through the eyes of tomography. The effect is to smooth the pattern of cold/fast regions, but not so much as to lose the detail seen in the temperature fields. The effect is more noticeable at shallower depths, where anomalies are significantly broadened. Between 900 and 1100 km the filtering effect reduces some of the linear nature of the anomalies, making them appear more cylindrical. The filtering clearly reduces the sharpness of features, losing some of the detail observed in the temperature field, though the resolution is not coarsened as significantly as in the low-resolution mantle circulation model. The amplitude of seismic anomalies is also reduced to values more similar to those observed in S40RTS reducing the observed problem of over-estimating seismic velocity mentioned in the methods section.

Ideally one would wish to filter the mantle circulation model predictions with the resolution operator of each tomography model individually. At this time the only available resolution operator was for S40RTS. For future work on this topic it would be useful to filter model predictions using other tomographic resolution operator to better understand the areas of weakness in tomography studies.

As I intend to ensure the tomography anomalies considered throughout this chapter are robust across a variety of tomography studies I will not continue to use the filtered model predictions as this will mean a valid comparison can only be made with S40RTS. However

Figure 4.8 (previous page): Six horizontal depth slices through a selection of models, demonstrating the effect of resolution on the system. Cases 001 & 002 show modelled temperature anomaly (radial mean subtracted) scale is ±600 K. Case 003 is modelled temperature from case 002 converted to seismic velocity and filtered to match the resolution of S40RTS. Scale is ±1.6% perturbation. Present day coastlines are overlain in thin black lines for reference.
performing the resolution filtering clearly illustrates the effect seismic inversion may have. It shows that although models match well to seismic tomography there is not a perfect fit and that at least some of the differences can be attributed to tomography inversion techniques. For the remainder of this chapter I will present the results of the model with plate motion history applied in discrete stages. I will, however, continue to use the thermodynamic conversion from modelled temperature predictions to seismic velocities, choosing between $V_p$ and $V_s$ variations where appropriate.

### 4.3.2 Validating the model with seismic tomography

To demonstrate the validity of these mantle circulation models I once again make comparisons between model predictions and seismic velocity anomalies imaged by seismic tomography. The same six tomography studies considered in chapter 3 are used here. In this section maps of velocity anomalies from the tomography studies focusing on the Tethys region in more detail are used. Since comparing temperature predictions to seismic tomography is only analogous I also make a more direct comparison between the model and seismic tomography by converting the modelled temperature field to seismic velocities using the technology described in section 4.2. To demonstrate the conversion from temperature to seismic velocities Figure 4.9 presents horizontal cross sections through case 001 at two depths, 1070 km and 1510 km. The images show the model temperature prediction alongside the conversion to $V_p$ and $V_s$ using an anelastic approximation (Goes et al. 2004).

The thermodynamic conversion in figure 4.9 illustrates that comparisons between temperature and seismic tomography are valid. It is clear that the conversion does very little to alter the planform pattern of convection. There are no obvious major differences arising from the conversion. The two velocity fields in figure 4.9 are plotted on the same scale ($\pm 2.4\%$). This illustrates the differences between the $V_p$ and $V_s$, as with the seismic
Figure 4.9: Illustrating the conversion of model predicted temperatures to seismic velocity at two mid-mantle depths. P-wave and S-wave velocities are plotted on the same scale to illustrate the differences between the two transformations.

velocities calculated in tomography the magnitude of S-wave velocity anomalies are higher the magnitude of P-wave velocity anomalies. As in tomography this is replicated in the conversion of model temperature to seismic velocity. A noticeable difference between the modelled seismic velocity perturbations and the perturbations in tomography is magnitude. As noted by Schuberth et al (2009a) applying a tomographic resolution filter to the calculated velocity perturbations can reduce the amplitudes to values similar to observed in tomography.

Figures 4.10 and 4.11 show the validation of the model results at present day for six depths through the mid-mantle. The depths chosen are non-uniformly spaced as the vertical resolution varies considerably between different tomography studies. The six depths are chosen as each of the six tomography studies and the mantle circulation model all have a horizontal shell within 10 km of the chosen depth. The figures compare horizontal cross
sections from seismic tomography in the study area from six different tomography studies including both P and S wave data with the model predictions. Figure 4.10 presents the comparison of three P-wave models, MITP08 (Li et al. 2008), GyPSuM-P (Simmons et al. 2010) and P-mean (Becker & Boschi 2002) against the model predictions converted to $V_p$ perturbations from case 001.

Assimilating plate motions into the model allows for accurate location of down-welling mantle material. However, as up-wellings from the deep mantle are only a minor contribution to plate motions and there is no prescribed velocity at the core mantle boundary, up-wellings are not as likely to be accurately located in the model. Only at the shallowest depths where mantle material is affected by mid-ocean ridge processes will seismically slow regions of mantle match with the hotter regions in the model. With enough plate motion history it is possible that down-wellings can condition the CMB and focus up-wellings into the spaces in between down-wellings. On the global scale this process produces hot regions in similar locations to currently observed hot spots, for example, the South Pacific or Indian Ocean. It is not enough to produce plumes in the exact locations observed in tomography. As a result the comparison between the model predictions and seismic tomography should only be made for faster than average seismic velocity anomalies associated with down-wellings.

Despite variations between models and with depth, I attempt to keep the colour scales as uniform as possible. For P-wave tomography studies the scale saturates at ± 0.6% velocity perturbation. The amplitude of velocity anomalies in the model predictions is larger, in this case the scale saturates at ± 2.4%. The scales are kept consistent across all depths, leading to some of the differences observed. This is noticeable particularly in the model predictions column where the plots for the two shallower depths do not come close to the colour scale saturation. Generally the observed anomalies are robust across each of the P-wave tomography models and model predictions match well with the observations.
Figure 4.10: Horizontal depth slices at six depths comparing three P wave tomography studies with model predictions converted to $V_p$ perturbations from case 001. For the three tomography models the scale is $\pm 0.6\% V_p$ perturbation. For the model predictions the scale is $\pm 2.4\% V_p$ perturbation. Present day coastlines are overlain in thin black lines for reference.
At shallower depths the linear north west to south east trending anomaly across the Indian continent fits well with the similar anomaly imaged in tomography. In the south the modelled anomaly is larger and has higher amplitude than imaged in tomography. In particular there is little in the way of a more southern anomaly in MITP08. Deeper horizontal cross sections show a slightly better match with the lateral extent of the modelled anomaly matching well to those imaged in tomography. One interesting feature is the amplitude of the modelled $V_p$ perturbations which increase with depth. This is likely to be an effect arising from the conversion from model temperature anomalies to seismic velocity perturbations where some sensitivity variations with depth may occur.

Figure 4.11 compares the same model, converted to $V_s$ perturbations to three S-wave tomography studies S40RTS (Ritsema et al. 2011) GyPSuM-S (Simmons et al. 2010) as S-mean (Becker & Boschi 2002). As with figure 4.10 I look for robust features across the S-wave tomography models and the match between the tomographic images and model predictions.

For S-wave tomography the anomalies are broader in nature but the observed patterns are similar to those in P-wave tomography. The IC (Indian Continent) north west to south east trending anomaly is once again the most obvious feature across the three tomography studies and the range of depths. The more southern anomaly in the Indian Ocean is more evident in S-wave tomography, particularly in S40RTS. There is some difference between the location of this anomaly in the model. In tomography it appears as a more cylindrical feature in the Arabian sea, where a second more linear feature is observed in the model. At shallower depths (700 km, 925 km) the fit is once again not so good. There are more evident variations between tomography studies at these depths. There are some similarities between the modelled temperature field and tomography models, at 700 km the pair of north to south striking anomalies in GyPSuM-S have some similarities to features
Figure 4.11: Horizontal depth slices at six depths comparing three S-wave tomography studies with model predictions converted to $V_s$ perturbations from case 001. For S4ORTS the scale is $\pm 1.6\% V_s$ perturbation, for GyPSuM-S and S-mean $\pm 1.2\% V_s$. For the model predictions the scale is $\pm 2.4\% V_s$ perturbation. Present day coastlines are overlain in thin black lines for reference.
in model. At 925 km some of the features in the model replicate thinner versions of the fast anomalies in S-mean. At the deeper depths the match is more uniform. There is more agreement between the three tomography studies, and there is a better match between the modelled $V_s$ perturbations. Matches to the model predictions are particularly good at 1250 km and 1370 km.

As in chapter 3 I also present the direct comparison between the tomography models and the modelled temperature predictions. I compare the model predictions to all six seismic tomography studies described above. In figure 4.12 I compare model predictions to P-wave tomography studies MITP08, GyPSuM-P and P-mean at the four depths between 1070 and 1510 km in the same geographic region as above. In the figures the threshold of subducted material in tomography is represented by the black contour. For the P-wave models the threshold value chosen is $+0.3\%$. For the temperature field mantle 250 K colder than the radial average is selected to represent slab material and is plotted in the red contour.

Figure 4.12 further indicates a good match between the mantle model and P-wave seismic tomography. Particularly below 1250 km depth. Some differences are highlighted, such as the lack of faster than average material south of India at 1070 km, here there is no evidence from tomography that the modelled feature exists. While at deeper depths there is faster than average mantle in the south of the region, however the model does not reproduce the geographic locations accurately. Generally I suggest that the match to the GyPSuM and mean models is better than to MITP08. In MITP08 no fast material beneath the Indian ocean in the south is observed.

As S-wave tomography models tend to have higher amplitude variations I choose values of $V_s 0.6\%$ faster than average for the slab threshold in figure 4.13. In this figure I keep 250 K as the threshold for the model temperature predictions. Once again the match
Figure 4.12: Horizontal maps of slab locations from three P-wave tomography models compared to the geodynamic model predictions. Black contours represent material 0.3% faster than average in tomography whilst red contours is material 250 K colder than average in the model predictions. Present day coastlines are overlain for reference.
Figure 4.13: Horizontal maps of slab locations from three S-wave tomography models compared to the geodynamic model predictions. Black contours represent material 0.6% faster than average in tomography whilst red contours is material 250 K colder than average in the model predictions. Present day coastlines are overlain for reference.
between tomography and model predictions deeper than 1250 km is shown to be good. At 1070 km there are some significant differences between the different S-wave tomography models e.g. the large feature between India and Arabia observed in S40RTS is not present in the other two studies. This feature could possibly coincide with the, slightly smaller, modelled feature to the south of India. In general the match between model predictions and S-wave tomography is very good. The model does not fit perfectly to any one tomography study but matches well with elements of each one. The fit to S-wave tomography is better than the fit to P-wave tomography.

The close match between modelled temperature field and seismic tomography indicates that the mantle properties used in the model are representing actual mantle conditions well. Given that the model accurately recreates the fast seismic tomography anomalies it is sensible to use passive tracer particles to link the deep mantle anomalies to surface processes. Tracers are placed on a single plate at the surface and as down-wellings are formed at convergent margins the particles are subducted and follow the down-going flow pattern. For this study particles are used to track cold anomalies originating from Indian and Spongtang Ocean plates. In this study the Indian plate represents the Neotethys Ocean from 155 Ma until the collision of India with Asia. Post collision, around 50 Ma, Indian plate particles represent ongoing northward convergence between India and Eurasia. The Spongtang ocean is the name given to the short lived back-arc ocean opening between Eurasia and the Neotethys.

4.3.3 Plate tracking marker particles

Figure 4.14 shows the relationship between the model predictions converted to $V_s$ perturbations from case 001 at present day and tracer particles tracking the subduction of the Neotethys Ocean. As expected, there is excellent agreement between particle locations
and cold anomalies predicted by the model. At the same depth intervals the tracer locations from two plates are presented. Tracers plotted in red represent subducted Neotethys lithosphere whilst those in green originated in the Spongtang Ocean. In the deeper mantle both particle fields overlap. This occurs as the model contains very little slab rheology, therefore subduction at certain boundaries will be symmetrical, two sided subduction. Areas where Indian and south Spongtang particles overlap are interpreted to originate at the intra-ocean subduction zone where Neotethyan lithosphere subducts behind the smaller back-arc ocean. In regions where just Indian Ocean particles exist, the anomaly is interpreted to originate from subduction at the Eurasian margin. As the majority of the particles thought to originate at the Eurasian margin are in the shallower portion of the model it is likely that these are from more recent subduction post the Indian continent colliding with the Spongtang arc.

At 600 km depth, the vast majority of the observed cold anomaly around the Indian continent trends northeast to southwest, this is most likely to be recently subducted material at the margin between India and Asia. Some of the material is probably the final remnants of Neotethys subduction post the collision of the Indian landmass with the Spongtang back-arc. The remaining shallower material probably originates from the continuing indentation of India into Asia and ongoing subduction, perhaps in a process similar to delaminating lithosphere. This faster than average mantle extends to around 1070 km depth with primarily red (Indian) particles identifying the anomaly. Beneath 925 km depth a new feature is observed beneath the Indian Ocean to the west and south of the continent. This anomaly is made up of both red (Indian) and green (Spongtang) particles thought to originate at the intra-ocean subduction zone. Between 925 km and 1070 km this is a minor component of the total fast anomaly at the depth and probably represents the final stages of Neotethys subduction at the intra-ocean subduction zone. By 1250 km depth the two anomalies are
closer together. Although they still remain separate. There is a mix of material at this depth with roughly half from each subduction zone. The northernmost of the two anomalies is material from subduction at the Eurasian margin, while the southern anomaly is composed of material from the intra-ocean subduction zone. Deeper than 1250 km depth most of the faster than average material is in one large anomaly beneath the Indian continent. At this depth the only material originating at the Eurasian margin is in the easternmost edge of the study area. A combination of the depth and the type of particles at that depth suggest older subduction solely at the intra-ocean subduction zone, probably with material subducted at around 140 Ma. The combination of tracer particles with MCMs strongly suggests the deeper, southernmost anomaly is a result of intra-ocean subduction within the Neotethys. The material in the north of the region is likely to be from conventional subduction of the Neotethys Ocean beneath Eurasia.

Figure 4.15 shows vertical south to north cross-sections of model predictions, particle field and tomography anomalies. For this figure I chose to plot the model predictions of $V_p$ as the only tomography study I have in suitable data format for plotting cross sections is MITP08. In this figure I plot the modelled $V_p$ anomalies with the contour of

**Figure 4.14 (previous page):** Plate tracking marker particles presented on top of the modelled temperature field for the chosen case. Particle locations are contoured as the thick black lines, over the modelled temperature field (column 1) and MITP08 $V_p$ perturbation (column 3). Column two shows locations of material from two separate subduction zones. Particles contoured in red are subducted at the margin of the Eurasian continent; those in green are subducted at the intra-ocean subduction zone. The overlap occurs due to two-sided subduction in the model. The back-arc plates subducted a both edges. Present day coastlines are overlain in thin black lines for reference.
particle locations overlain in column 1. Column 2 shows the particle field divided by the plate of origin. The particles presented in column 2 are similar to those in figure 4.14 with red particles originated on the UNIL (2009) Indian plate, green particles originate on the Spongtang ocean. I also include in blue particles originating on the Neotethys ocean prior to 155 Ma. Column 3 shows the contour of all particles plotted on the P-wave velocity perturbation from MITP08 (Li et al. 2008).

The tracers cover the majority of colder than average mantle in the cross-section, they match well with the modelled faster than average mantle. A notable exception is the northern most mid mantle anomaly at 85° and 90° east where the particles associated with the oldest Neotethys subduction (blue) do not reach the depths they may be expected to and are not clearly related to a region of faster than average mantle. In the two western cross sections blue particles do reach nearly 2000 km depth and correspond to colder than average mantle. The cross sections illustrate well the nature of subduction in the region. Column 2 demonstrates that the southern most anomaly at mid-mantle depths is composed of two sets of particles, those originating on the Indian plate and those originating on the Spongtang plate. Following the ideas set out above, this suggests that this anomaly is material subducted at the intra-ocean subduction zone. The shallower anomaly at around 25° to 30° north extending to around 1000 km depth is composed mainly of red particles associated with the Indian plate, this backs up the thought that it is more recent subduction post the accretion of the back arc to the Indian continent and due to ongoing northward motions. Finally the blue particles in the north of the region are associated with Neotethys subduction between 210 Ma and 140 Ma, which has also been named Mesotethys subduction.

Figure 4.15 generally demonstrates a good match between faster than average mantle predicted by mantle circulation models and the locations of passive markers. There are some noticeable differences between the two data sets.
Figure 4.15: As figure 4.14 but in vertical south to north cross-section. The cross-section extends from 15° south of the equator to 45° north. The figure shows the deepest extent of material subducted in the past 155 Myr is 2000 km.
4.4 Discussion

4.4.1 Comparing model predictions to tomography

Whilst the match between modelled mantle conditions and seismic tomography is generally good, the model predictions are not completely compatible with any single tomography model. The major robust feature at mid-mantle depths running northeast to southwest is imaged in all tomography models from around 1000 km depth to 1600 km depth. The major difference occurs in the Indian Ocean to the south and west of the Indian continent. In this region there is less agreement between tomography models, some of which (e.g. the GyPSuM and 'mean' models) show a lower amplitude anomaly, interpreted as the result of the intra-ocean subduction zone. Others such as MITP08 do not record a strong fast velocity anomaly in this region, while S40RTS images a more cylindrical anomaly to the east of the modelled seismic velocity anomalies. The circulation model predicts a fast anomaly dipping northwards in a similar location to the GyPSuM and 'mean' tomography studies. The model predicted anomaly is located slightly to the west of the location imaged in tomography, and the magnitude of the anomaly is greater.

The south-north cross sections in figure 4.15 provide further insight into this region. Cross sections at 75°, 80° and 85° east show a shallowly northward dipping cold region in the south of the area between 1000 and 1500km. The tracer particles indicate the anomaly is caused by subduction at the intra-ocean subduction zone, however it does not resemble the shape of the corresponding seismic fast region [III, Me, IO] described in section 4.1.1. This is most likely explained by the nature of subduction in the model. The model contains no slab rheology other than a small amount of temperature dependent viscosity giving a little extra slab strength. In the circulation model slabs follow Stokes flow and sink vertically. This effect combined with the southward migration of the intra-oceanic subduction zone
in the prescribed plate motion history this could result in the shallowly northward dipping anomaly observed in figure 4.14. Although geologically sound evidence for the intra-ocean Neotethys subduction zone is strong (Aitchison et al. 2007, 2000, Aitchison & Davis 2004) the geographical location is less certain. The modelled seismic velocity anomalies therefore do create a cross-section resembling the anomalies I - IV proposed by van der Voo et al. (1999b). To obtain a better fit in terms of shape a small modification to the plate motion history could be made. The suggested improvement to the tectonic reconstruction would be to make the convergent boundary between the Neotethys and Spongtang oceans more stationary instead of migrating to the south towards present day. A stationary subduction zone is more likely to produce a near vertical sinking, larger volume temperature anomaly that would better fit the observations in tomography.

The central of the three mid-mantle anomalies is the most voluminous, similar to those imaged by tomography. Tracer particles show this anomaly to be primarily formed of material originating at the intra-ocean subduction zone. In the shallower regions it is composed of material originating from the Eurasian margin subduction zone. The region matches well with the lower part of the central anomaly imaged in tomography studies [II, IC]. The particles observed in this portion of the model are best interpreted in a similar manner to Hafkenscheid et al. (2006). The deeper part of the anomaly originates from intra-ocean Neotethys subduction and forms part of a broader anomaly dipping northwards from the south of the region comprising of anomaly III and the deeper part of II from van der Voo et al. (1999b) and the IO and IC anomalies from Hafkenscheid et al. (2006). The shallower part of the central anomaly is also sensibly interpreted in the same way as Hafkenscheid et al. (2006). This anomaly is the subduction of the Spongtang back arc basin after the entire Neotethys Ocean has been consumed beneath the back arc and the Indian continent has accreted the arc. The shallow depth of the anomaly and the 'red' particles present
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there indicate younger subduction of this ocean at the Eurasian margin, again most sensibly interpreted as late subduction of the Spongtaang Ocean. Figure 4.16 shows the modelled seismic velocity perturbations with the particle field contoured on top, annotated based upon the plate of particle origin. Panels (b) and (c) are included for reference and show the particles coloured by original surface plate and Hafkenscheid et al. (2006)'s interpretation of the Bijwaard et al. (1998) tomography respectively.

The modelled tracers indicate the northernmost anomaly to originate from subduction of the oldest Neotethys ocean in the UNIL (2009) plate motion history. This plate is subduction at the Eurasian margin from 210 Ma to approximately 140 Ma. This interpretation fits well with the ideas presented in the studies of tomography above which variously suggested Palaeotethys or Mesotethys subduction to be the cause of the imaged fast anomaly. These interpretations are essentially the same, just differing based upon naming. As suggested in Jarvis & Lowman (2005) this is a result of subduction which terminated prior to 140 Ma. This study does differ from the modelling in Jarvis & Lowman (2005) as the modelled slab does not sink to the core mantle boundary as in their model but remains at mid to deep mantle depths. There is a faster than average region modelled in the circulation model at the core mantle boundary. I suggest this is a result of the very oldest Palaeotethys subduction in the region between 300 and 210 Ma.

Finally there are the shallow fast regions present from the surface down to 1000 km depth. It is challenging to separate the slab like down-welling in the circulation model from the mid-mantle anomaly corresponding to II. However, I suggest the model to be interpreted in a similar way to Hafkenscheid et al. (2006) where particles between 500 km and 1000 km depth are the final stages of Neotethys subduction prior to India’s collision with Asia but after it accretes the back arc. In the UNIL (2009) reconstructions this is technically the Indian plate subducting and therefore red particles would be expected in this
Figure 4.16: An interpreted north-south cross section at 75° east demonstrating (a) the seismic velocity perturbations from the mantle circulation model with contours of the modelled passive tracers in thick black lines. Panel (b) is the modelled particle fields contoured by colour based upon the plate of origin, where blue is oldest Neotethys, green is Spongtang back arc and red is Indian. Panel (c) shows the interpretation of tomography anomalies in the region by Hafkenscheid et al. (2006) for reference.
region. The very shallowest anomalies down to 400 km depth are interpreted to be a result of post-collisional tectonics, potentially delamination of lithosphere or continent-continent subduction. In the mantle circulation model it is challenging to distinguish between this and the deeper anomalies suggesting the applied plate motion history has continuous subduction at this boundary post-collision. This is not unreasonable for TERRA as currently there is nothing included in the model to distinguish between ocean and continental plates. If, as in this case, the surface velocities continue to be convergent subduction is likely to continue regardless of the nature of the convergence.

Some of the differences between the model predictions and the various seismic tomography studies could be a result of the resolution or resolving power of tomography in certain regions, particularly the southernmost of the three mid-mantle fast anomalies. I will discuss the effects of resolution in more detail in section 4.4.2. Here I will briefly use other geophysical evidence to speculate that the southernmost anomaly may be a true mid-mantle feature despite being the least consistent of the anomalies across the range of seismic tomography studies. Recent geoid mapping by satellites such as GRACE and GOCE reveals a significant geoid low in the Indian ocean directly south of India (Pail et al. 2011, Reigber et al. 2005). The geoid anomaly is the largest in magnitude observed by either of the satellites and could correspond to significant volumes of dense material at mid-mantle depths. It is beyond the scope of this work to calculate geoid variations from the mantle circulation model, but this could be a significant further test of the model, which could help to understand mid-mantle structure in the Tethys region, particularly when there are some doubts in the robustness of features in seismic tomography. Models of density heterogeneity derived from plate motions (Ricard et al. 1993) have been shown to produce a synthetic geoid which matches more than 80% of the observed geoid. It is important to note that this is purely speculative as it has been shown that the geoid arising from density variations is
sensitive to the depth and wavelength of the density anomaly (Ricard et al. 1984, Richards & Hager 1984). Below a certain depth the density anomalies have been shown to have an inverted effect on the geoid. For example a low density anomaly at lower mantle depth may have the same contribution to the observed geoid as a high density anomaly near to the surface. Compressible convection and radial viscosity variations provide further complexity to the sensitivity of the geoid to density anomalies (Panasyuk et al. 1996, Steinberger 2000).

4.4.2 Resolution considerations

The effect of resolution on the comparison between seismic tomography and modelled mantle conditions is pronounced. The lower resolution, therefore lower Rayleigh number case is clearly different from the high-resolution case. Both cases produce cold regions beneath India where the Neotethys Ocean subducts, however in the low-resolution case the cold anomaly is greater in magnitude and spatial extent. In reality, tomography images the anomalies in this region to be somewhere in between the two in terms of area. This hints that resolution has a large effect on the comparison between model predictions and seismic observations. Furthermore, the plate tracing particles in the lower-resolution run demonstrate very similar behaviour to those in the higher resolution study. Particles reaching deeper depths in the simulation tend to originate from the intra-ocean subduction zone, whilst the shallower particles are subducted at the Eurasian margin. A link to the surface is still present as a form of subduction continues post-collision.

Filtering model predictions using the resolution operator of S40RTS provides further insight into the effect of resolution. When applied to the high-resolution simulation the sharpness of the cold features is reduced considerably, the filtering effect produces a field resembling S-wave tomography. The conversion from calculated temperature and pressure to seismic velocity via mineral physics results in higher amplitude of anomaly than observed.
in tomography prior to filtering with the tomography resolution operator.

Although the mantle circulation models presented in this chapter produce mid-mantle structure in this region that is largely similar to structure imaged in seismic tomography, it could be argued that the modelled southernmost anomaly is not robust across all tomography studies. It is certainly true that this anomaly is least consistent in the considered tomography studies. In some tomography studies (e.g. MITP08) there is no evident faster than average mantle in the region, whilst in the GyPSuM models there are lower amplitude anomalies in the Arabian sea at shallow depths and to the south of India at deeper depths. S40RTS has a larger amplitude anomaly to the south and west of the Indian continent. Clearly, there is some disagreement between the tomography data sets. Some of these discrepancies could result from the limitations of tomography. Low ray density in certain regions could explain some discrepancies between certain tomography models. For example, in MITP08 the fast anomaly linked to intra-ocean subduction in the south and west of the region is very weak. In vertical cross section it is difficult to observe at all. This is potentially due to the lack of resolution of body wave tomography in the oceanic regions. As mentioned above, this anomaly exists at varying amplitudes in other tomography studies suggesting it is a robust feature. Filtering of model predictions with a seismic tomography resolution operator goes someway towards matching the resolution and damping. It is a challenge to the modelling process to capture the detail in plate motion history at lower resolution however. High resolution mantle circulation models are required to capture the details of the imaged mid-mantle structure at Earth-like Rayleigh number. However, at higher resolution the model predictions do not show the same spatial extent as fast regions imaged in tomography. Filtering using resolution operators can go some way to a more direct comparison but the limited availability reduce the number of tomography models that can be considered. If the resolution operators for a number of tomography
models were available a more robust comparison of the effects of filtering mantle circulation models would be possible. With this some of the weakness of various tomographic inversions could be assessed, and the physical differences between model predictions and tomography would be clearer.

4.4.3 Post collision tectonics

The shallowest anomaly present is thought to be a result of either lithospheric delamination after the collision of India with Asia, or continuing subduction as India moves northwards. In this model the anomaly between the surface and 900 km depth has a steep, southward dip, despite resulting from the northward motion of India indenting into Asia. In vertical cross-section, tracer particles show this cold region to originate from subduction of the Indian plate, which at the time of collision represent convergence between the Indian continent and Asia. The tracers match up well with MITP08, particularly at 80° and 85°E. In places tomography images similar features in the shallow mantle, and the van der Voo et al. (1999b) interpretation of the University of Utrecht tomography (Bijwaard et al. 1998) also contains a southward dipping feature in the region. One possible interpretation of this observation is that despite the collision of the two landmasses the motion of the boundary is moving northwards more rapidly than any subducted or delaminated material can sink. Rather than the slab bending backwards to produce the southward dipping anomaly it is more likely that vertical subduction occurs at all locations as the boundary migrates northwards leaving a sheet of subducted material rather than a backward dipping slab.

4.5 Conclusions

Combining mantle circulation simulations with tracer particles provides an excellent method to link modelled cold mantle regions to the surface processes that created them.
The method provides further support to analyses of seismic tomography that record the complex nature of subduction during the closure of the Tethys oceans. The dynamic model presented here broadly agrees with many interpretations made based solely upon seismic fast anomalies. Using plate tracking marker particles alongside the convection code I have shown it is possible to understand which of the seismic fast anomalies originate at which surface subduction zone. The particles reveal enough information to begin to distinguish between some of the details of the different interpretations made. Overall this model supports the interpretation of Hafkenscheid et al. (2006) that the majority of the mid-mantle seismic anomalies originate from an intra-Neotethys ocean subduction zone, with only shallower features originating at the Eurasian margin post 140 Ma. I suggest that the northernmost anomaly observed in tomography is caused by subduction of an ocean terminating at around 140 Ma before present, which is again consistent with most conclusions based upon seismology alone, although there may be some differences in terminology used. Mantle circulation models also provide a valuable test of mantle parameters and a means of testing plate tectonic reconstructions. In areas of uncertainty this technique could be used to provide extra constrains on locations and stability of convergent margins.
Chapter 5

Thoughts on heat flow and dynamic topography

5.1 Introduction

In previous chapters I have shown that mantle circulation models can be used to simulate a mantle with a distribution of temperature and density anomalies resembling Earth’s present day mantle as imaged by seismic tomography. Therefore, it is possible to use the model data to make predictions of near surface properties of economic and scientific importance. Oil industry basin models commonly take a constant value across the globe for values of mantle heat flux. In basin models, dynamic topography is a relatively new consideration despite playing an important role in sea level changes and basin formation. Although these properties are of value to the oil industry little has been done from a modelling perspective, particularly on a global scale such as the models I will present here. Dynamic topography in particular is an area of active research and there is little consensus on methods used to calculate it. Even a definition of dynamic topography is difficult to agree upon.
With studies on local, regional, continental and global scales, magnitudes and wavelengths of dynamic topography features are actively and widely discussed. In this chapter I will look at the global mantle heat flux, some of the different ways researchers have studied dynamic topography and present results from TERRA simulations. I will consider variations of mantle heat flux through time, the calculation of dynamic topography at present day and the validity of using mantle circulation models to study variations in dynamic topography through recent geological history.

5.2 Modelling Method

Like in previous chapters this study mostly considers the results of mantle circulation models at present day. As the goal is to find out what can be learned about heat flow and dynamic tomography from mantle circulation models, I consider the best case model thus far. I have shown in the parameter study of chapter 3 that mantle circulation models can be used to select a set of parameters which accurately reproduce the large scale seismic structure of Earth’s mantle. Higher resolution models also capture some of the finer regional scale structure as seen in chapter 4. The model presented in chapter 4 demonstrated a good match between mid-mantle seismic structures and the distribution of temperature anomalies in the mantle simulations. Therefore this model provides a good basis for studying surface effects such as the heat flow and dynamic topography. Model parameters for the simulation considered here are recapped in table 5.1. A three layer viscosity structure is used for radial viscosity variations and is shown in figure 5.1 where the solid line is the layer mean viscosity and the dashed lines represent the minimum and maximum values of viscosity arising from the inclusion of temperature dependent viscosity. Temperature dependence of viscosity is included according to equation 5.1.
\[ \eta(T) = \eta_0 \exp[-E_a T] \] (5.1)

The modelling process is the same three stage process as used throughout this thesis. Results are primarily considered at present day in order to understand model predictions at a familiar time. The advantage of these mantle circulation models is they allow for predictions to be made at any time throughout the model and to examine the time varying effects of the important parameters. I will consider predictions made for both heat flow and dynamic topography throughout geological time. Predictions are made at the intervals between plate reconstructions provided by UNIL. The full methods are described in chapter 2, while methods specific to certain sub-chapters, such as the equations for calculating dynamic topography, are included in the relevant sections of this chapter.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reference viscosity</td>
<td>$2 \times 10^{21}$ PaS</td>
</tr>
<tr>
<td>Lower mantle viscosity</td>
<td>$8 \times 10^{22}$ PaS</td>
</tr>
<tr>
<td>Ea</td>
<td>4.61</td>
</tr>
<tr>
<td>Mineral phase changes</td>
<td>None</td>
</tr>
<tr>
<td>Equation of state</td>
<td>Compressible Murnaghan</td>
</tr>
<tr>
<td>Thermal conductivity</td>
<td>$4 \text{ Wm}^{-1}\text{K}^{-1}$</td>
</tr>
<tr>
<td>Internal heat production</td>
<td>$4 \times 10^{-12} \text{ Wm}^{-3}$</td>
</tr>
<tr>
<td>Surface temperature</td>
<td>300 $K$</td>
</tr>
<tr>
<td>CMB temperature</td>
<td>3800 $K$</td>
</tr>
<tr>
<td>Velocity boundary condition</td>
<td>300 Myr plate motions</td>
</tr>
</tbody>
</table>

Table 5.1: Model parameters for the simulation used to investigate heat flow and dynamic topography.
5.3 Heat Flow

Commercial basin models often use simple or crude estimates of important mantle parameters when assessing the economic viability of a specific basin. Parameters such as the basal heat flow may only be included in basin models as the same constant value across the globe. In mantle circulation models near surface temperatures are controlled by the assimilated plate motions, therefore, it is possible to use the near surface temperature predictions to estimate the mantle heat flux across the surface boundary of the model. Heat flux is calculated from Fourier’s Law:

\[ Q = \kappa \frac{dT}{dz} \]  

(5.2)

Where \( Q \) is the heat flow across a given layer \((\text{W m}^{-2})\), \( \kappa \) is the thermal conductivity \((\text{W m}^{-1}\text{K}^{-1})\) and is specified in the model input. \( dT \) is temperature difference across the layer \((\text{K})\) and \( dz \) is the thickness of the layer \((\text{m})\). To calculate heat flow across a given layer, equation (5.2) can be written in the form:
\[ Q = k \frac{(T_{dz2} - T_{dz1})}{\Delta t} \] (5.3)

Where \( T_{dz} \) is the temperature field on a given layer, the subscripts 1 and 2 refer to the upper and lower temperature fields respectively. In some cases heat flow is calculated on the uppermost radial layer of the model. Here a uniform temperature boundary condition, specified in the model input is subtracted from the predicted temperature field \( T_{dz2} \). In other cases, a lithosphere may be simulated using an increased viscosity in the uppermost layers, in this case it is more appropriate to calculate heat flow at the base of the lithosphere. Two temperature planforms on the two radial surfaces closest to the base of the lithosphere are used in calculating heat flow.

In figure 5.2 I present plots of the mantle component of heat flux through time from 250 million years before present to present day. The heat flux in these figures is calculated from the horizontal temperature variations one model layer down from the surface at 22 km depth. It is therefore logical that the plots are dominated by plate boundaries. Most obviously, the regions of high heat flux are concentrated near mid-ocean ridges. The figures for time intervals closer to present day show differences between fast and slower spreading ridges with higher absolute heat flux and wider lateral extent of high heat flux areas near faster spreading ridges (East Pacific Rise and Indian Ocean ridges). The coldest regions are regions of active tectonic convergence although they are not significantly colder than the surrounding continental regions.

Figure 5.2 demonstrates the change in surface mantle heat flux through time. Heat flux is dominated by plate boundaries throughout the entire 250 million years. Plots for all time slices are plotted on the same scale from 0 to 300 mW m\(^{-3}\), indicating a fairly uniform heat flux throughout the model time. There are some periods of high heat flux over a
Chapter 5: Thoughts on heat flow and dynamic topography

Figure 5.2

Mantle heat flux 250 Ma - 155 Ma

Heat flow, W/m²

0.04 0.08 0.12 0.16 0.20 0.24 0.28

250 Ma

230 Ma

220 Ma

210 Ma

200 Ma

180 Ma

165 Ma

155 Ma
Figure 5.2

Mantle heat flux 131 Ma - 57 Ma

131 Ma

121 Ma

112 Ma

103 Ma

095 Ma

084 Ma

070 Ma

057 Ma

Heat flow, W / m²

0.04 0.08 0.12 0.16 0.20 0.24 0.28
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Figure 5.2: Plots of surface variations of the mantle component of heat flux throughout time as calculated from TERRA simulations.
large region, such as the early Pacific Ocean between 165 Ma and 103 Ma where large areas with heat flux of greater than 150 mW m$^{-3}$ are present. The absolute values of heat flux calculated are generally in agreement with those calculated over the last 65 million years by Loyd et al. (2007).

Loyd et al. (2007) calculate variations in oceanic heat flow, using a half space cooling model (Jaupart et al. 2009), ocean ages (Xu et al. 2006) and two different tectonic reconstructions (Gordon & Jurdy 1986, Hall 2002). They demonstrate how heat flow has reduced by a small amount during the Cenozoic. They attribute the change to the amount of oceanic crust close to mid ocean ridges. Looking at the Pacific Ocean, Loyd et al. (2007, figure 4) shows a decrease of at least 50% in the area of the oceanic crust close to the ridge (ridge-proximal) over the last 100 million years. This suggests that present day heat flow could be quite low relative to some times in the geological past. For oceanic crust, predictions of heat flow from mantle circulation models are similar to those of Loyd et al. (2007) for the Cenozoic. This suggests that, predictions of heat flow from mantle circulation models further back in time may have some value. However, predictions further back in history are limited by greater uncertainties and model dependence upon its initial condition.

For illustration I plot the total surface heat flux for the most recent 100 million years of the mantle circulation model in figure 5.3. Note in this plot the lower axis is model time, which corresponds to approximately 103 million years of Earth time when scaling is considered. the origin of the plot is 103 million years before present, heading towards present day on the right. The plot shows a number of abrupt changes in heat flux, each of which is associated with a stage change (there are 12 stage changes in the time frame considered). However a background cooling trend is present if these abrupt changes are neglected. In the last 50 million years the trend is more steady, with a roughly constant heat flux. The trend is slightly different to Loyd et al. (2007) but a general reduction in heat flow in recent
geological history is observed.

![Graph of surface heat flux against model time, for the most recent 103 Myr.](image)

**Figure 5.3:** Surface heat flux against model time, for the most recent 103 Myr

Figure 5.4 shows the heat flow predictions from Loyd et al. (2007) for comparison with the time varying plots of heat flux presented in figure 5.2. As mentioned earlier they observe a general reduction in the magnitude of heat flow over the last 65 million years, which they attribute to a reduction in ridge-proximal oceanic crust. The reductions are possibly more pronounced in the work of Loyd et al. (2007) than in the predictions of mantle circulation models, however, there is some evident decrease in areas of high heat flow since 65 Ma.

Continental regions are more challenging to understand. In the model predictions of heat flow presented in figure 5.2 continental regions clearly have considerably lower heat flow from the mantle. This is probably a result of continents being relatively stable throughout the 300 million years of model time. As radioactivity in the crust is thought to be responsible for most of the heat flux in continental regions, it is not surprising that mantle circulation models do not properly capture the surface heat flow in continental regions since only the mantle component is modelled.

When compared to the work of Loyd, model predictions of heat flow seem to be
Figure 5.4: Heat flow predictions from Loyd et al. (2007) for comparison with those predicted by mantle circulation models. Continents have a uniform value of 65 mW m$^{-3}$. 
reasonably robust in the past 65 million years, suggesting that predictions further back in
geological time have some value. The predictions further back in time must be treated with
some caution, according to the caveats mentioned previously. The heat flow calculated here
could be significant for industry basin models, it is certainly an improvement on choosing a
constant value for the mantle component of heat flux. The absolute values of heat flow may
be less useful, particularly on the smaller scales of basin models. A potential application of
the data is track the heat flow at a specific location back through time, according to the
plate motion history. This will provide changes in heat flow for a specific location as the
input to the basin models.

5.4 Dynamic Topography

5.4.1 What is dynamic topography?

Earth’s surface topography is, to the first order, representative of the tectonic
processes that shape it. The highest mountain ranges are a direct result of compressional
tectonic forces, whilst deep ocean trenches form where old ocean crust subducts beneath
younger continental crust. Some non-tectonic features such as ocean island volcanoes (e.g.
Hawaii, Reunion) are also explained by mantle processes. Up-welling plumes of hot ma-
terial are the most likely source of such features. Despite understanding the vast majority
of topographic features, some aspects of Earth’s observed topography still remain poorly
understood. A particularly prominent example is southern Africa, which is highly elevated
despite the African continent being surrounded by passive margins and, therefore, remaining
tectonically inactive during recent geological time. To understand anomalous features such
as these, Earth scientists look towards the mantle. It is hypothesised that viscous stresses
arising from the convecting mantle cause deflections in major boundary layers, such as the
crust and core-mantle boundary (CMB) as well as other boundary layers such as transition zone phase changes. The challenge for Earth scientists is to separate this anomalous topography, also known as flow induced or dynamically supported topography, from the tectonically generated topography.

Dynamic topography is a challenging concept. Many different definitions for what constitutes dynamic topography exist, varying with data sources and scale of observations. Dynamic topography is currently an active research topic with much debate on both temporal and spatial scales of anomalous elevations, as well as the magnitude of the dynamic signal. In the context of mantle modelling it may be defined as anomalous topography arising from buoyancy distributions within Earth’s mantle. It is the observed topography that cannot be explained by surface processes and isostatic effects (Hager & Richards 1989).

Essentially, it is the study of vertical motions of the Earth’s crust due to viscous forces in the mantle, hence the alternative name, flow induced topography. In the 1970’s Anderson and McKenzie first linked mantle flow to gravity anomalies and surface deformations (Anderson et al. 1973, McKenzie 1977, McKenzie et al. 1974). By this time numerical models had been used to show that upward convection of hot mantle material can be associated with a positive gravity anomaly, and thus an anomalous flow driven elevation. McKenzie’s subsequent papers further develop the idea and incorporate numerical techniques. They show, for a simple two dimensional numerical case, that the surface deformation is strongly dependent on viscosity variations (McKenzie 1977).

Figure 5.5, adapted from Davies (1999), is a simple, schematic representation of how a sphere of buoyant, viscous material can deflect the free surface of a fluid layer.

As previously mentioned, dynamic topography refers to areas of anomalous elevation that cannot be explained by simply isostasy, such as the African ‘superswell’ (Lithgow-Bertelloni & Silver 1998) which can be attributed to buoyant upwelling mantle potentially...
Figure 5.5: A sketch of how buoyant viscous spheres affect the free surface of a fluid layer, assuming a less dense fluid above, and a more dense fluid below. Sphere A is negatively buoyant, like a subducting slab, and causes a depression or basin at the surface. Sphere B is rising, and thus deflects the lower boundary locally. Sphere C also rises, and shows how a buoyant sphere in the mid mantle deflects both surfaces over a long wavelength. Adapted from Davies (1999)

from the lower mantle (Pari 2001). This is most simply explained by sphere ‘C’ in figure 5.5 which demonstrates a buoyant rising sphere deflecting both boundaries of its containing fluid. Alternatively, dynamic topography can refer to areas of negative buoyancy in the upper mantle related to palaeo-subduction zones. A sinking buoyant sphere, as demonstrated by sphere ‘A’ in figure 5.5 can deflect the upper boundary of the fluid downwards to form a trench. This commonly occurs when a continent passes over the remains of a now extinct mid-mantle slab. One area where this is potentially occurring is the North Atlantic, where Earth’s surface is pulled down relative to what might be expected from surface processes alone. This is likely to be related to the sinking of the Farallon slab in the mid-mantle (Conrad et al. 2004).

Whilst complicated to predict, variations in dynamic topography over tens of millions of years are useful for oil industry basin models. For example, they can be used to derive information on relative sea levels, and thus basin forming conditions. One motivation for this study is to understand at least some of the variations in dynamic topography predicted
by mantle circulation models over the course of the model time. However to understand variations through time it is also important to consider anomalous topography at present day to ensure predictions are physically reasonable. One example of the economic aspects of dynamic topography in this context is the New Jersey margin of North America in the Cretaceous. Cretaceous sea levels were generally high, however estimates of variations range by an order of magnitude (Müller et al. 2008). Müller et al. (2008) suggest that the New Jersey margin not only subsided due to cooling and thickening of the lithosphere away from the Mid Atlantic Ridge but also due viscous forces created by the movement of North America over the Farallon plate, thus allowing for sea levels to be higher than previously thought. Of course, knowing sea levels in the past is extremely important in resource exploration.

As the dynamic signal of topography is masked by isostatic effects and surface process it can be very challenging to directly observe dynamic topography and so there are many approaches to looking at the dynamic topography problem. They range from global estimates of anomalous topography from instantaneous flow models or dynamic mantle models over long periods of time to localised field studies of relative sea level. Here I will briefly examine some of the work on dynamic topography to date before presenting some methods of extracting dynamic topography predictions from mantle circulation models.

Dynamic topography can be estimated on a variety of scales using subsidence data. In continental regions, the stratigraphic record can be used to constrain the timing of episodes of anomalous subsidence (Wheeler & White 2002), potentially related to dynamic processes. Subsidence data are used alongside seismic reflection data to constrain temporal evolution of the strain rate in a region (in the case of Wheeler & White (2002), south east Asia). This method can be used to investigate dynamic variations in topography from the stratigraphic record. It is probably most valid at a more localised scale than the long wavelength variations hypothesised by other methods. Magnitudes of dynamic sub-
sidence calculated by this method are up to approximately 500 m at short wavelengths. The magnitudes of topography estimated from stratigraphy are less than hypothesised long wavelength dynamic topography of 1 - 2 km from the global studies discussed later in this chapter. The difference is most likely a result of the difference in the scale of observations, with more local data sources returning smaller values for dynamic topography.

On a continental scale, estimates of dynamically supported topography can be quantified spatially and temporally by inverting a network of river profiles (Pritchard et al. (2009), Roberts & White (2010) Roberts et al. 2012a, Roberts et al. 2012b). The shape of the river profiles is assumed to be controlled by uplift rate and erosion. River profiles are inverted for uplift rate history. For western North America the river profile inversions suggests around 2.5 km of uplift over the past 70 million years (Roberts et al. 2012b). The same techniques also suggest uplifts of 1 - 2 km over 15 million years for Madagascar consistent with dynamically supported topography (Roberts et al. 2012a), and suggests rapid growth of domes in Africa over the past 40 million years (Roberts & White 2010).

Dynamic topography has also been used to understand alterations in the drainage pattern of the Amazon River (Shephard et al. 2010). They suggest that significant changes in the drainage of wetland areas of northern South America, previously attributed to Andean uplift, could be caused by dynamic topography. Shephard et al. (2010) suggest South America has moved westward over cold and dense subducted slabs, resulting in subsidence and then rebound of the Amazon region. They use a coupled model of mantle convection and plate kinematics to study the uplift and subsidence of the region in the Cenozoic. These models predict initial subsidence of 40 metres per million years followed by a similar rate of uplift after 30 million years before present.

The flooding of Australia in the Cenozoic is also attributed to mantle flow induced topography (Heine et al. 2010). Heine et al. (2010) studied the process using a backward
advection of seismic tomography method, combined with plate kinematics, palaeogeography and sea level data. Their models reveal two dynamic topography lows, which the Australian continent has progressively moved over during the last 70 million years. The gradual subsidence of Australia as it moves over the dynamic topography lows results in the progressive flooding. Heine et al. (2010) predict a maximum of between 500 and 600 metres of dynamic topography affecting Australia during the Cenozoic.

These methods clearly present valuable information on aspects of local to continental scale dynamic topography. However, some of the methods are used to calculate either dynamic uplift or dynamic subsidence. It may be more useful to consider global predictions of dynamic topography with both uplift and subsidence. Global, instantaneous flow models from buoyancy and density variations within the mantle can be used to predict present day dynamically supported topography. The flow models are commonly derived from density distributions observed in seismic tomography studies. The distribution of density and buoyancy can then be used to estimate dynamic topography (e.g. Conrad et al. 2007, Lithgow-Bertelloni & Silver 1998, Ricard et al. 1993, Simmons et al. 2006). The dynamic elevations predicted from instantaneous flow models will be considered further, later in this chapter.

Dynamic modelling also provides a useful tool for studying that which cannot be directly observed. Numerical simulations of Earth’s mantle can used to produce global or local estimates of mantle stresses responsible for dynamic topography. Two approaches to calculating dynamic topography are possible. Firstly mantle circulation models (MCMs) assimilating plate motion history over recent geological time can be used to simulate a mantle with conditions representing those imaged in seismic tomography. Then the distribution of temperature, pressure and velocity from the simulation can be used to generate a stress field, and thus estimate the mantle derived component of dynamic topography. Secondly,
an inverse modelling method assimilating present day seismic structure and modelling backwards can be used (Bunge et al. 2003). This method involves calculating density variations from seismic velocities from tomography studies and running the convection calculation backwards. This adjoint method of simulating mantle convection has been used to predict dynamic topography during the Cenozoic (e.g. Conrad & Gurnis 2003). This method is computationally expensive and will not be considered further in this study, despite generating better initial conditions for the models. An example of dynamic topography calculated from adjoint method comes from Conrad & Gurnis (2003), they predict between 500 m and 700 m of uplift during the Cenozoic.

5.4.2 Calculating dynamic topography

The buoyancy of a non-accelerating, rising sphere of fluid must be balanced at any instant due to Newton’s laws of motion. When the sphere is close to the free surface of the fluid layer, viscous stresses locally transmit the force through the fluid (Davies 1999). At the free surface the force is taken up by the surface and there is a balance between the weight of the additional topography and the buoyancy of the sphere. This provides a simple situation where the magnitude of dynamic topography is related to the buoyancy or density difference and the volume of the sphere. Whilst a sphere is chosen to simply illustrate the concept, on Earth, buoyancy variations could be generated by rising plume heads or subducting slabs. These features could generate positive and negative topography respectively.

To estimate flow induced elevations derived from mantle circulation models one must first consider the stress field arising from the viscous forces within the spherical shell. In TERRA, the full stress tensor, $\sigma_{ij}$ in its general form is defined by Baumgardner (1983, equation 2.4) as
\[ \sigma_{ij} = -p \delta_{ij} + \tau_{ij} \] (5.4)

Where \( \sigma_{ij} \) is the stress tensor, \( p \) is the dynamic pressure, \( \delta_{ij} \) is the Kronecker delta and \( \tau_{ij} \) is the deviatoric (viscous) stress. To consider vertical motions of the surface due to the stresses in the fluid I only consider the normal stress acting in the radial direction upon the surface, \( \sigma_{rr} \). The need to consider \( \sigma_{rr} \) arises as in TERRA the radial velocity on the surface layer of nodes is prescribed to be zero. Thus equation 5.4 is modified to become:

\[ \sigma_{rr} = -p + \tau_{rr} \] (5.5)

Where the symbols have the same definition and the subscripts refer to the direction in which the force is acting and the plane upon which it is acting respectively, in this case the stress in the radial direction operating on the surface of the model domain. The deviatoric stress, \( \tau_{rr} \), is calculated from the gradient of velocity according to equation 5.6

\[ \tau_{rr} = \eta \left( \nabla \mathbf{u} + (\nabla \mathbf{u})^T \right) \] (5.6)

where \( \tau_{rr} \) is the deviatoric (viscous) stress, \( \eta \) is the fluid viscosity and \( \mathbf{u} \) is the fluid velocity. It should be noted, that for these calculations TERRA assumes the geological sign convention for stresses where a compressive stress is positive (Twiss & Moores 1992).

For a free surface, e.g. that of Earth, there should be no shear or normal stresses acting on that surface. However, in many models of mantle convection a free slip or no slip surface boundary condition is generally applied. In these cases some stresses will be present
at this boundary. As users of TERRA will commonly choose a free slip boundary condition, where no vertical motions are allowed, I will focus upon a free slip boundary condition. In the case of a free slip boundary condition, finite normal stresses (but no shear stresses) result at the surface boundary (Ricard 2009). The normal stress at a free surface should be equal to zero so a simple force balance, equation 5.7, can be used to predict the dynamic topography on the free slip boundary.

\[ \sigma_{rr} = \Delta \rho g z \]  

(5.7)

Where \( z \) is the magnitude of dynamic topography, \( \Delta \rho \) is the density difference between the buoyant mantle and the medium above the surface (air or water), and \( g \) is the acceleration due to gravity. Ricard (2009) notes that for long wavelength structure this is a good first order approximation and suggests for short wavelength and rapid events more precise computation (e.g. Zhong et al. 1996) of time dependent topography is required.

As a first approximation, this force balance is a good starting point for calculating dynamic topography arising from the mantle circulation models. I will generally consider large scale features across the entire globe and thus will not consider the finer scale features and short temporal events which require more precise calculations. \( \sigma_{rr} \) is easily calculated from equations 5.5 and 5.6 and so the only variable in the force balance is \( z \), the dynamic topography. A small modification is made to the equation to account for the sign conventions in TERRA, as the stress is acting in the radial direction on the surface plane the restoring force acts in the opposite direction and introduces a minus sign on the right hand side of equation 5.8. This ensures positive values of \( z \) equate to positive dynamic topography.
Chapter 5: Thoughts on heat flow and dynamic topography

\[ z = -\frac{\sigma_{rr}}{\Delta \rho g} \] (5.8)

Again, \( z \) is the dynamic topography in metres, \( \sigma_{rr} \) is the calculated normal stress in Pascals, \( g \) is gravitational acceleration (10 ms\(^{-2}\)) and \( \Delta \rho \) is the change in density between the two media (3300 kgm\(^{-3}\) for mantle to air, the simplest case).

I briefly note some limitations and considerations when using this method. Firstly, this approximation is strictly for viscous fluids only and takes no account of any elasticity in the crust. As a result the predicted dynamic topography is a point-wise estimate. Currently TERRA includes no near surface elasticity and ‘plates’ are only included via a prescribed velocity boundary condition on the surface layer of grid points. If elastic plates could be included at the surface a more realistic smooth estimate of dynamic topography would be obtained.

Secondly, as suggested in figure 5.5 buoyant material will deflect both the top and bottom surface of the fluid layer. Any weakly deformable internal boundaries may also be influenced, providing there is a sufficient density contrast between the two layers (Ricard 2009). To accurately detail the dynamic topography arising from density variations within the mantle the lower boundary and internal boundaries such as mineral phase transitions at 410 km and 660 km depth should be included. For simplicity, in this chapter I consider only anomalous elevations arising from deformation of the surface boundary in the predictions of dynamic topography. Future work may allow for more detailed predictions of dynamic topography due to other deformable boundaries but this will not be presented in this thesis. One possible benefit of predicting topography at all of these boundaries is that it becomes possible to predict a synthetic geoid based upon the density variations in the model, providing a further test of model validity by comparing to observed geoid data and providing a
potential test of some of the points raised in the discussion of chapter 4.

Finally I will examine some of the previous work done to refine predictions of flow-induced topography. Pysklywec & Hosein Shahnas (2003) note the limitations of calculating $\sigma_{rr}$ at the upper boundary of the model, as this does not take into account the response of buoyant crust to the deformation caused by $\sigma_{rr}$. They test the effect of varying rheology, and find that a deformable crust affects time and length scales of dynamic topography as well as a dependence on thickening and uplift. Much of the current literature uses the method of Zhong et al. (1993) as the basis for calculations of $\sigma_{rr}$, which in turn is used to calculate surface topography, geoid anomalies and gravity anomalies. Their method is similar in approach to the simple method detailed above. They use the Boussinesq approximation, to simulate incompressibility and solve the momentum equation directly to obtain values for surface stresses. Burgess et al. (1997) calculates dynamic topography using the method of Zhong et al. (1993) to calculate $\sigma_{rr}$ then equates it directly to local topography for a given temperature field. Although the exact equation used to obtain a value for dynamic topography is not given, the methodology is similar to the force balance method described above. This method is also used for studies of dynamic topography at convergent plate margins (Billen et al. 2003, Buiter et al. 2001). This method could be used to produce the more precise calculations of dynamic topography suggested by Ricard (2009), however, as I will only consider long wavelength features predicted by global models it is unnecessary for this study. Finally, Zhong et al. (1996) notes that calculating the normal stress at the surface and calculating dynamic topography via the force balance method is valid for long wavelength dynamic topography.
5.4.3 Dynamic topography at present day

Despite choosing a relatively simple method of equating normal stress at the surface boundary to anomalous elevation there are still a number of factors to consider when attempting to predict dynamic topography from mantle circulation models. Factors such as the depth at which normal stress is calculated, the nature of the surface boundary condition at the time of calculation and the density contrast used to calculate dynamic topography from mantle convection could all have a significant role in the predicted topography. In this sub-chapter I will consider a number of factors which may contribute to the calculation in order to produce a definition of dynamic topography as predicted by mantle circulation models using TERRA. First I will look at factors affecting the calculation of normal stress prior to conversion to dynamic topography.

Equation 5.5 shows that the normal stress used in the force balance is calculated from two fundamental variables within TERRA, the pressure and the deviatoric stress. In the simplest calculation the deviatoric stress is added to the negative of the pressure on each grid point within the whole domain to calculate a normal stress. The variation of stresses laterally can then be calculated on any given radial shell. Figure 5.6 shows these three fields as calculated at the surface of the model.

This figure shows that the calculated pressure is the dominant factor in the calculation of normal stress. In this case it is one order of magnitude larger than the deviatoric stress, $\tau_{rr}$. Both of these properties, as might be expected, are dominated by the plate boundaries. The pressure field clearly demonstrates regions of high pressure in areas of passive up-welling and low pressure in regions of convergence. Whilst the modelled $\tau_{rr}$ field has one clear region of high, positive stress beneath the Andes. Ridges are not so clear in the deviatoric stress field. Combining the two fields derived from TERRA to produce the normal stress results in a field with positive stress in regions of subduction and negative stress in
Figure 5.6: Horizontal cross sections of three model outputs at the surface of the model shell. (a) pressure, Pa; (b) deviatoric stress, Pa; (c) normal stress, Pa. Note the different scales
regions of spreading. At first, this seems unphysical as one would expect ridges to have large dynamically supported topography and so positive stress. This apparent reversal is due to the sign convention and as mentioned above the height is calculated from the negative of the normal stress in the force balance equation. With this in mind, the calculated normal stress shown in figure 5.6c seems a reasonable first estimate and should produce sensible results once converted to dynamic topography.

The first consideration to make is the depth at which to calculate the normal stress for conversion to dynamic topography. This mantle circulation model does not contain a rigid lithospheric layer or any other distinguishing near surface rheology, so it may be sensible to choose the uppermost radial layer of the simulation (22 km depth) for the calculation. Otherwise it may be sensible to take into account the near surface viscosity increase implemented to mimic a lithosphere and perform the stress calculations at the base of the highly viscous region at around 135 km depth. Figure 5.7 demonstrates that either of these approaches is valid as there is no discernible difference between the two calculations. The modelled stress fields in each case are almost identical. Visually there is little separating the different cases and the minimum and maximum values of stress show very little variation (table 5.2).

<table>
<thead>
<tr>
<th>Calculation Depth</th>
<th>Min.</th>
<th>Max.</th>
<th>RMS</th>
</tr>
</thead>
<tbody>
<tr>
<td>22 km</td>
<td>(-1.64 \times 10^8)</td>
<td>(2.21 \times 10^8)</td>
<td>(6.78 \times 10^7)</td>
</tr>
<tr>
<td>135 km</td>
<td>(-1.75 \times 10^8)</td>
<td>(2.57 \times 10^8)</td>
<td>(6.43 \times 10^7)</td>
</tr>
</tbody>
</table>

**Table 5.2:** Table demonstrating the variations in model stress prediction for the two calculation depths shown in figure 5.7. Values are all normal stress in Pa

Another consideration is the ‘time’ at which the stress calculation is performed. It is possible to calculate the stress field on the final time step of the convection calculation, in this case the boundary condition is the imposed velocity from plate tectonic reconstructions.
Figure 5.7: Horizontal cross sections of the normal stress calculated from TERRA plotted at 22 km depth (left) and at 135 km depth (right). Thick black lines indicate present day plate boundaries, thin lines represent present day coastlines.

Otherwise stress can be calculated a single time step after the end of the simulation using a free slip boundary condition instead of the tectonic boundary condition. This may be a sensible choice as it will negate any excessive effect the imposed velocity has upon the stress calculations, this also satisfies the theory suggesting that the accumulation of normal stress at the surface boundary is best balanced against dynamic topography given a free slip boundary condition (Ricard 2009). One further option is to perform the stress calculation after a period of relaxation. In this instance, the calculation is continued for up to 100 time steps with a free slip upper boundary condition post the completion of the circulation calculation. This serves to reduce the effect of the imposed velocity boundary condition on the physical properties of the model and again allow the calculation to be performed against a free slip boundary condition. It is important that the relaxation time is not so long that the convection calculation progresses away from the modelled present day mantle structure. These three options for calculating normal stress from the mantle circulation model are presented in figure 5.8. The figure illustrates that there is very little difference between the second and third methods so a small period of relaxation does not alter the pattern of the
stress field dramatically. Calculating the normal stress with the velocity boundary condition still applied increases the magnitude of the stress, particularly around the plate boundaries.

![Figure 5.8: Horizontal cross sections of the normal stress calculated from TERRA plotted at present day. Left figure is after 1 time step of free slip convection and the right figure is after a period of 100 time steps of relaxation. Thick black lines indicate present day plate boundaries, thin lines represent present day coastlines.](image)

Including a period of relaxation could potentially decrease the dominance of plate boundary features on the pattern of normal stress, however, including 100 time steps of relaxation does not appear to have a large effect on the pattern or magnitude. There is some trade off between the amount of relaxation and how much the model progresses from the present day distribution of temperature and density. Longer periods of relaxation may reduce the dominance of plate boundaries on the normal stress, allowing for easier exploration of other features of interest, however it may lead to the simulation progressing away from the present day predictions, reducing the accuracy of the calculated normal stress.

In these cases the most interesting feature arising from the three different ap-
proaches to calculating dynamic topography is the difference between the case where the plate motion surface velocity boundary condition remains in place and the cases where it does not. Alongside the obvious one order of magnitude increase in amplitude there are differences in the pattern too. Most noticeable is that the increase in magnitude seems to be restricted very locally to plate boundaries. There is very little structure away from the plate boundaries. Some of the structure observed in the two free slip cases may be present but masked by the nature of the colour scale and the more extreme amplitudes at plate boundaries. Due to conditions imposed by the theory set out in Ricard (2009), it is sensible to choose a free slip condition for the calculation of normal stress. As performing the calculation after a period of relaxation has very little effect on the predicted normal stress, it is computationally sensible to choose the standard free slip method for the calculation, illustrated in figure 5.8a, as the period of relaxation has so little effect on the calculated normal stress. I include a table (table 5.3) to demonstrate the subtle differences between the two calculations with a free slip surface boundary condition. There are some small differences in the maxima and minima but the values remain essentially similar. The order of magnitude difference between the plate tectonic boundary condition calculation is also shown.

<table>
<thead>
<tr>
<th>Calculation Depth</th>
<th>Min.</th>
<th>Max.</th>
<th>RMS</th>
</tr>
</thead>
<tbody>
<tr>
<td>Free Slip</td>
<td>-1.65x10^8</td>
<td>2.23x10^8</td>
<td>6.44x10^7</td>
</tr>
<tr>
<td>Relaxed</td>
<td>-1.76x10^8</td>
<td>2.45x10^8</td>
<td>6.84x10^7</td>
</tr>
<tr>
<td>Plate Motion</td>
<td>-6.48x10^9</td>
<td>7.91x10^9</td>
<td>3.84x10^8</td>
</tr>
</tbody>
</table>

Table 5.3: Table demonstrating the variations in model stress prediction for the three calculation methods in figure 5.8. Values are all normal stress in Pa.

Since the calculated normal stress varies very little between the methods outlined above (excluding when the imposed plate motion boundary condition remains in place), I chose to calculate the normal stress for the remaining calculations at 22 km depth (one
layer down in the model domain) and after one time step with a free slip upper boundary condition. There still remain a number of possible options for a conversion from normal stress to dynamic topography. As stated in sub-chapter 5.4.2 dynamic topography is calculated by a force balance equation, equation 5.8. The conversion from stress to dynamic topography requires a density contrast, \( \Delta \rho \). Initially I chose a uniform density contrast for the conversion, the contrast assumes a mantle to air transition resulting in a \( \Delta \rho \) of 3300 kg m\(^{-3}\). A second conversion was made using continent to ocean boundaries from the UNIL (2009) plate motion history. For this conversion an air over mantle transition is used for continental regions whilst water over mantle is used for oceanic crust, a density contrast of 2300 kg m\(^{-3}\). A comparison of the two cases is shown in figure 5.9.

**Figure 5.9:** Horizontal cross sections of the conversion from modelled normal stress to dynamic topography. (a) \( \Delta \rho = 3300 \text{ kg m}^{-3} \), assuming a mantle to air conversion globally. (b) \( \Delta \rho = 2300 \text{ kg m}^{-3} \), assuming mantle to water conversion for oceanic crust and mantle to air elsewhere. Scale is ± 6000 m. Thick black lines indicate present day plate boundaries, thin lines represent present day coastlines. The layer average is removed from the raw data to give a zero mean dynamic topography.

As might be expected, given the calculated normal stress, it is clearly observed that the dynamic topography at present day is heavily dominated by plate boundary pro-
cesses. Regions of spreading have large positive anomalous topography whilst regions of ongoing convergence demonstrate large negative dynamic topography features. Given that topographic features near plate boundaries (mid-ocean ridges and trenches) are thought to be the most prominent dynamically supported features on Earth (Hager & Richards 1989) the observed planform seems reasonable. Many study areas for more regional analysis of dynamic topography are away from plate boundaries so some of the more regional signal may be hidden by the dominant plate boundary signal observed in figure 5.9. The lateral pattern of calculated dynamic topography is not distinctly different between the two assumed density contrasts. There is no change in continental regions as the assumed density contrast is unchanged, whereas using a lower $\Delta \rho$ in oceanic regions naturally results in larger amplitude dynamic topography in these regions. One interesting and perhaps counterintuitive observation in both parts of figure 5.9 is the elevation of continental areas relative to oceanic areas. This is evidently not what is observed for Earth’s actual topography. The topography calculated by this method is, therefore, topography generated solely by viscous flow in the modelled mantle.

From the calculations presented so far in this sub-chapter I arrive at a definition of dynamic topography used within this chapter.

*Dynamic topography is the anomalous topography arising from density and buoyancy anomalies present in the whole model domain, calculated at a the surface with a free slip boundary condition.*

This definition is different to some other geophysical studies of dynamic topography, particularly some instantaneous flow calculations based upon the conversion of seismic tomography to density variations (e.g. Conrad & Husson (2009), Forte et al. (2007), Simmons et al. (2006), Moucha et al. (2008a), Moucha et al. (2008b)). Often such studies will neglect
density entirely in the upper few hundred kilometres of the mantle, arguing that unknown lithosphere effects should not be included in the definition of dynamic topography as they are likely included in isostatic effects (Conrad et al. 2007, Conrad & Husson 2009, Lithgow-Bertelloni & Silver 1998). It is argued that density variations in the uppermost mantle may be a result of chemical rather than thermal heterogeneity and so density variations in this region can be neglected (Lithgow-Bertelloni & Silver 1998). Alternatively, studies such as Husson (2006) examining the effect of subducting slabs on dynamic topography consider only point density anomalies shallower than 670 km depth to calculate dynamic topography, presumably as density anomalies below this depth have a lesser effect on the overall topography. It is evident that a variety of methods for understanding the mantle’s effect on Earth’s topography are used. Although calculating the ‘full’ dynamic topography, as I have done throughout this chapter, including the entire model domain is uncommon it remains a valid method. The method is particularly useful for mantle circulation models as they contain no lithosphere rheology or chemical variations. I note that Čadek & Fleitout (2003) suggest that variations in the top 300 km of the mantle are important for calculating geoid anomalies and dynamic topography particularly in models with imposed plate velocities.

It is common to compare dynamic topography predicted from geophysical modelling with a residual topography calculated by removing isostatic effects from observed Earth topography (e.g. Panasyuk & Hager 2000). Here I twist this slightly and apply corrections for crustal thickness and isostasy to the modelled dynamic topography to attempt to reproduce Earth’s observed topography. The first figures presented take the calculated normal stress, then convert the value to a dynamic topography as in equation 5.8. I then correct for isostasy effects assuming Airy isostasy for continental crust and Pratt isostasy for oceanic crust. The two isostasy hypotheses are illustrated in figure 5.10 adapted from Fowler (2005).
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Figure 5.10: Illustrations of the two hypotheses of isostasy (a) Airy isostasy and (b) Pratt isostasy.

Fowler (2005) states that for Airy’s hypothesis of isostasy the continental root, $r$, is calculated according to equation 5.9

$$ r = \frac{h \rho_c}{\rho_m - \rho_c} $$

(5.9)

where $h$ is the height above sea level, $\rho_c$ is the density of crust and $\rho_m$ is the density of mantle.

For continental regions I assume the modelled dynamic topography to act as the root of the continent, essentially deflecting the Moho, a sensible assumption as the mantle circulation model contains no crust and thus the upper surface is the discontinuity between mantle and crust. Therefore the modelled dynamic topography takes the place of $r$ in equation 5.9. As a result an elevation above the datum can be calculated according to equation 5.10. This is akin to ‘pouring’ crust over a surface deformed by dynamic topography, then calculating the isostatic balance.
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\[ h = -\frac{\sigma_{rr}}{\Delta \rho g} \frac{\rho_m - \rho_c}{\rho_c} \]  

(5.10)

Where \( h \) is the calculated relief, \( \sigma_{rr} \) is the modelled normal stress, \( \Delta \rho \) is the density difference, in this case from mantle to continental crust, \( \rho_m - \rho_c \), \( g \) is acceleration due to gravity. \( \rho_c \) and \( \rho_m \) are the densities of crust and mantle respectively. As \( \Delta \rho \) is the density difference between mantle and continental crust parts of equation 5.10 cancel to give:

\[ h = -\frac{\sigma_{rr}}{\rho_c g} \]  

(5.11)

Figure 5.11 shows predicted relief calculated from the mantle circulation model using the isostatic corrections mentioned. Calculated values are corrected to the approximate average relief of continent (2 km) or ocean (- 4 km). Blue colours are bathymetry, whilst browns and yellows are topography above sea level. The topography calculated is based upon the normal stress anomalies calculated within TERRA. Therefore stresses and derived values are based upon the normal stress variation from the mean value.

The figure demonstrates a remarkable match to Earth’s observed relief for such a simple approximation. High elevations are observed in mountainous regions such as the Himalayas and the Andes, while ocean trenches are the most extreme bathymetric feature. Mid-ocean ridges have significantly shallower water depth than older oceans. Some differences are notable, for example southern Africa has significant anomalous elevation which is not observed in the model. Most continental interiors show very little topography outside of convergence regions. These findings are discussed further in section 5.4.4. One noticeable difference is the magnitude of the predicted topography, which is around half as large again as observed topography. Despite this the deepest ocean trench predicted is approximately
Figure 5.11: Map of predicted relief calculated from modelled dynamic topography with the addition of an isostatic correction. Plate boundaries from UNIL (2009) are plotted in thick black lines, coastlines are plotted in thin black lines.

11 km whilst the highest mountains are 8 km. The predicted values are very close to actual values for the Marianas trench and the highest Himalayas.

This experiment serves to highlight the importance of isostasy whilst considering dynamic topography. It is, perhaps, unphysical to think of Earth in this way, i.e. mantle density variations deforming the moho, before a deformable crust is emplaced on top. However, considering just mantle processes one would expect areas of anomalous negative topography above subducting slabs associated with collision of plates. Clearly tectonic forces are important in mountain building, but looking at dynamic topography in this ways suggests the dynamic component of topography under large mountain ranges may be negative. Isostasy becomes an important factor when considering dynamic topography as crust in a basin like region of dynamic topography will undergo isostatic rebound reducing the vis-
ible effect of dynamic topography. It is important to stress that this experiment is overly simplistic and should not be treated as an accurate way of predicting topography or palaeotopography from mantle circulation models. It does however illustrate some important points about the way dynamic topography is thought about and the role of isostasy in understand dynamic topography at present day.

A further way to calculate the relief from the dynamic topography is to include a value for crustal thickness. By simply adding measured crustal thickness to modelled dynamic topography one can obtain an estimate of relief. Here I choose the crustal thickness model CRUST 2.0 (Laske & Masters 2001). Figure 5.12 illustrates the global crustal thickness from the model CRUST 2.0.

![Figure 5.12: Map of crust thickness from CRUST 2.0. From Laske & Masters (2001).](image)

Figure 5.13 presents the relief calculated from modelled dynamic topography corrected for crustal thickness from CRUST 2.0 (figure 5.12). This correction presents some further interesting results, oceanic bathymetry has a maximum amplitude of around 16 km,
an overestimate of around 50% while the highest mountains predicted by this method are approximately 4 km a 50% underestimate. The Himalayas have an appropriate high altitude, whilst other present day convergence zones such as the Andes have lower relief. The key feature here is the apparent elevation of regions such as southern Africa and the North American craton. Southern Africa in particular is known for having anomalously high topography, commonly attributed to the African super-swell (Lithgow-Bertelloni & Silver 1998) a phenomenon which could be observed in the dynamic topography calculated here. Much of the continental interior is at or around sea level. In the oceans, there is no unexpected features, mid-ocean ridges have higher elevation than surrounding ocean whilst trenches exhibit the largest negative elevation.

Figure 5.13: Map of predicted relief calculated from modelled dynamic topography with the addition of CRUST 2.0 data. Plate boundaries from UNIL (2009) are plotted in thick black lines, coastlines are plotted in thin black lines.
5.4.4 Discussion of present day dynamic topography

Currently there is little agreement on what constitutes dynamic topography and what should be included in calculations of dynamic topography from geodynamic models. I have previously mentioned how more localised studies predict small amounts of regional dynamic topography. Equally, global models of dynamic topography disagree on which density and buoyancy anomalies in the mantle contribute to surface dynamic topography. For example, some studies exclude near surface density variations in their calculation of dynamic topography. As it remains unclear what to include in the definition of dynamic topography, the dynamic topography calculated here from viscous stress for the whole model domain is analysed. Predictions of dynamic topography made in this way seem fairly robust and provide useful insights into the effect of Earth’s mantle on surface topography.

It may be useful to compare the dynamic topography calculated from geodynamic models with that calculated from instantaneous flow models. In figure 5.14 I compare the predicted dynamic topography calculated in chapter 5.4.3 with that predicted by Conrad & Husson (2009). To predict dynamic topography Conrad & Husson (2009) infer lateral density variations from seismic velocities from S-wave tomography model S20RTS (Ritsema et al. 2004). In the study, density anomalies shallower than 300 km depth are not included as fast velocity anomalies are assumed to correspond to neutrally buoyant continental roots and slow velocity anomalies correspond to decompression melting. Conrad & Husson (2009) suggest that down to 300 km the simple conversion from seismic velocity to density is not valid and more complex treatment would be required (e.g. Hernlund et al. 2008).

To first order the pattern of predicted topography is very similar between the two methods. Broadly speaking continental regions, such as Asia and the Americas exhibit negative topography in both figures 5.14a&b. Regions of high topography are observed in the oceanic regions in both studies. The main difference between the two studies is the
amplitude of dynamic topography. This study produces topography of amplitude greater than 4 km, a factor of 2.5 times greater than that of Conrad & Husson (2009). Furthermore the 4 km of modelled topography is greater than the maximum quoted magnitude of mantle supported long-wavelength dynamic topography from literature which is often around 2 km. Mantle viscosity appears to have an important role in how much topography can be supported dynamically. If the dependence is roughly linear a reduction in the model upper mantle reference viscosity may reduce the amplitudes of predicted dynamic topography by a similar amount, resulting in amplitudes similar to Conrad & Husson (2009). A factor of 2 decrease in the reference viscosity for the model presented in this chapter would change the upper mantle viscosity from $2 \times 10^{21}$ Pa s to $1 \times 10^{21}$ Pa s, which still remains a sensible choice for upper mantle viscosity.

In figure 5.14, continental regions are primarily regions of high amplitude negative topography. A perhaps counterintuitive observation considering that on Earth continental
regions are above sea level and ocean floor below. It is likely the continental crust masks the strong negative signal of dynamic topography though isostatic effects. In mantle circulation models the continents are part of the convecting system and so the negative dynamic topography observed is probably masked by buoyant lithosphere or thickened crust. In the case of the mantle circulation model continental regions have been at the surface of the domain for most of the model time. They remain colder than the surrounding oceans which are constantly being recycled by hot up-welling mantle. Of course, the mantle circulation model only considers 300 million years of history, so the effect of continental cooling could be significantly more pronounced for the real Earth.

Cratonic areas, such as southern Africa and central North America, have dynamic topography values of close to zero (yellow colours) in the mantle circulation models. Relative to other continental material cratons exhibit positive topography, this is expected for southern Africa which shows anomalous elevation unlikely to be associated with tectonic processes. When compared with Conrad & Husson (2009) (figure 5.14b) mantle circulation models do not produce the large, highly elevated region in southern Africa predicted by the instantaneous flow model. This hints at a more super-swell like feature beneath southern and eastern Africa and the Indian ocean is responsible for the large region of anomalous topography in figure 5.14b. Super-swell features are unlikely to be present in mantle circulation models. The Conrad & Husson (2009) estimate of dynamic topography is more akin to a degree 2 spherical harmonic pattern, whilst predictions from mantle circulation models have aspects of that pattern but are more dominated by near surface (plate boundary) processes. Seismic tomography studies, such as S20RTS (Ritsema et al. 2004), which forms the basis for their mantle flow model, have very evident degree 2 velocity structure in the lowermost mantle, which could be a significant contribution to the predicted dynamic topography. In mantle circulation models up-wellings are not conditioned other than in the
uppermost mantle where mid-ocean ridge processes are important. It is unlikely that mantle circulation models develop large super-swell like regions, like those observed in figure 5.14b, although some positive topography relative to the rest of the African continent is present in the circulation model predictions. Topography predictions from mantle circulation models produce results dominated by near surface processes, whilst predictions from the instantaneous flow calculation may be more affected by deeper processes. Overall this suggests the decision to include or exclude shallow density variations from models may be important, but it is difficult to decide which model is a better choice as dynamic topography is so difficult to observe.

It appears that the major differences between the two models is a result of how much mantle buoyancy is excluded from the dynamic topography calculation. For a more like for like comparison between these two models a further simulation is considered. The simulation takes the same present day mantle circulation model and includes a single extra time step, with a free slip surface boundary condition and density variations in the top thirteen model layers (equivalent to 290 km) removed. In figure 5.15 I show the result of this simulation compared to the predictions of Conrad & Husson (2009).

These two plots show a remarkable agreement between the two different modelling methods. The similarity suggests that both approaches to calculating dynamic topography have merit. It also suggests that the strong negative dynamic topography in continental regions is a real feature that is strongly masked by the isostasy of continents. Figure 5.15 confirms that the two different methods for calculating dynamic topography produce similar results and that the differences arise from including or excluding lithospheric density anomalies. Each method appears equally valid and which to choose depends of which aspects of dynamic topography are of interest. To understand all topography arising from mantle flow processes it might be more sensible to chose a calculation based upon the full mantle
density anomalies. However, if dynamic topography is defined as topography which is not understood by surface processes then calculating dynamic topography from sub-lithospheric density anomalies may be a better choice.

The only major difference between the two is the lower magnitude dynamic topography calculated from mantle circulation models in the Pacific Ocean. One might expect a strong, broad signal of positive topography to arise from the rapid spreading of the East Pacific Rise. This feature is noticeably missing throughout the calculations of dynamic topography presented in this chapter. One possible explanation for this is the ability of mantle circulation models to focus up-wellings into the correct places. Once the upper 300 km of density is extracted for this calculation, any up-welling derived from plate tectonic will be missing from the dynamic topography calculation. It is unlikely that these mantle circulation models will develop large warm regions that are able to support the large magnitude dynamic topography signal observed by Conrad & Husson (2009). To help understand these differences I plot the radial velocity at the end of the mantle convection simulation at ap-
proximately 150 km depth in figure 5.16. I chose 150 km depth as it is below the effect of the high viscosity 'lithosphere' layer and so the velocity will not be affected. At this depth the radial velocity matches quite well with the predicted dynamic topography. The dynamic topography calculated with the uppermost mantle density anomalies removed seems to be predominately a result of radial velocities. Dynamic topography calculated from the full set of modelled density anomalies is likely to have sources of stress from other sources, such as the imposed horizontal velocities.

![Map of radial velocity calculated for the mantle circulation model in this chapter at 150 km depth. Red represents positive, radially outward velocity. Blue represents negative, radially inward velocity.](image)

**Figure 5.16:** Map of radial velocity calculated for the mantle circulation model in this chapter at 150 km depth. Red represents positive, radially outward velocity. Blue represents negative, radially inward velocity.

Given that predictions of dynamic topography are sensible when compared to predictions from another type of modelling it is worth looking at the significance of the relief calculated in figures 5.11 and 5.13. In chapter 5.4.3 I highlighted some of the differences between the isostatic correction method and the addition of crust method for calculating relief from predicted dynamic topography. Figure 5.17 reproduces the estimated relief data (on the same scale for simplicity) alongside observed topography data set, ETOPO1, a 1 arc minute global relief model combining topography and ocean bathymetry (Amante & Eakins 2009).
Figure 5.17: Maps of predicted and observed relief: (a) predicted relief using isostatic correction, (b) predicted relief using CRUST2.0 correction (c) observed Earth topography, ETOPO1. (d) map of ocean age from Müller et al. (2008)

When plotted together the differences between the two methods for estimating relief become more apparent, although generally there are a lot of similarities. Both methods result in continents elevated above sea level with the highest elevations in regions of convergent plate boundaries (e.g. Himalayas, Andes). In ocean regions the deepest ocean is in trenches whilst shallow oceans are in regions of active spreading. By comparing the predicted ocean bathymetry to ETOPO1 (figure 5.17c) it seems that the models predict a larger range of bathymetry than is observed (Note: due to data processing, some of the deepest trench like features are not present in ETOPO1). In the Atlantic ocean, for example, a reasonably broad
area of the ocean is at elevations only slightly below 0 m (figure 5.17a), this is not replicated in ETOPO1. One feature that is captured in the model is the difference between fast and slow spreading mid-ocean ridges. ETOPO1 shows a broader region of elevated bathymetry for the East Pacific Rise when compared to the Mid-Atlantic Ridge. Interestingly, the bathymetry predicted from mantle circulation models, particularly for the case corrected for isostasy, demonstrates a good fit to the ocean age data of Müller et al. (2008), (figure 5.17d). This is an important observation as the cooling of oceanic crust with age may often not be classified as dynamic topography, as it is easily explained by an age versus depth relationship. Clearly the cooling that leads to this topography is part of the mantle convective system in these mantle circulation models and thus is included in the dynamic topography signal. When dynamic topography is defined to be anomalous topography which cannot be explained by surface process it is easy to see why ocean topography is not included in the dynamic signal. The observation that isostatic subsidence is inversely proportional to the square root of age (Davis & Lister 1974) adequately explains most ocean bathymetry. However, if oceanic lithosphere is considered part of the convecting system, as in the mantle circulation model presented here then any topography generated is mantle derived.

For continental regions there are a few more discrepancies between modelled relief and ETOPO1. For relief calculated using an isostatic correction the vast majority of continental regions have high, plateau-like, relief. Asia and South America are generally around 4 km elevation. Cratonic regions, southern Africa and central North America, are at lower relief when compared to surrounding continents. The predicted relief using a CRUST2.0 correction appears to contain more of the dynamic signal for continents with areas such as southern Africa, central North America and Antarctica having high relief. For relief predicted by the isostatic correction these areas correspond to regions of lower topography than the immediate surrounding continents. The CRUST2.0 corrected relief (figure 5.17b) is in
better agreement with ETOPO1 for continental regions.

Originally calculating relief from values of dynamic topography predicted from mantle circulation models was a test of the accuracy of the predictions. As it seems that the predicted dynamic topography allows relief to be estimated reasonably accurately the method could be used to estimate palaeo-topography during the Mesozoic. Obviously, a degree of caution would be required as figure 5.17 shows that the calculated relief is not perfect. There are still some unknowns, such as crustal thickness which would hinder such predictions, but using the full dynamic topography calculated from mantle circulation models would remove some uncertainties regarding unknown lithospheric structure in the geological past.

5.4.5 Mesozoic and Cenozoic dynamic topography

To understand how dynamic topography varies temporally, this sub-chapter looks at dynamic topography throughout the model run. As the uncertainties in calculating dynamic topography are large I chose the simplest method to calculate the topography, the force balance between the calculated normal stress and the weight of rock above. The density difference is taken to be that of mantle to air.

Figure 5.18 shows the calculated dynamic topography from the model from 250 Ma to present day. On the figure I plot the magnitude of topography with the radial average subtracted. Plotted on top in thick black lines are the corresponding plate boundaries at the time. In thin black lines I plot present day coastlines for reference. The primary feature of these plots is how the lateral pattern of topography is dominated by plate boundaries. Also of note is the large magnitude of calculated topography. Each of the figures is presented on the same scale from -5 km to +5 km. Other estimates of long wavelength dynamic topography suggest dynamic elevations of between 1 and 2 km amplitude. Although it is
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Figure 5.18
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Figure 5.18
Figure 5.18: Plots of surface variations of dynamic topography throughout time as calculated from TERRA simulations. Magnitudes are ± 5 km. This figure is over three pages.
worth noting that the highest magnitude anomalies are present at plate boundaries and many 
 studies of dynamic topography are at more regional scales and away from plate boundaries. 
 In these regions, estimates of dynamic topography presented here are of lower magnitude, 
 which may be roughly equivalent to published estimates of dynamic topography. It may 
 be useful to consider variations through time. This concept is similar to a virtual well 
 where a location of interest, for example the Gulf of Mexico, is rotated back to its previous 
 locations using the same plate motion history as is used for the modelling. At each time 
 interval the magnitude of dynamic topography for the palaeo-location of interest, and the 
 variation through time is calculated. The change in dynamic elevation is more useful for 
 industrial applications than the absolute values and partially negates any issues arising from 
 over-estimating the magnitude of dynamic topography.

5.5 Concluding Remarks

This chapter has demonstrated the potential of using mantle circulation models to 
calculate values of significance for industry, particularly in basin modelling a sea level calcula-
tions. Calculating values for the mantle component of heat flow may not be particularly 
novel on their own, but could prove useful in the context of plate reconstructions. Basin 
models which currently use a single constant value across the globe for their mantle heat 
flow could be significantly improved using a database of global heat flow values, particu-
larly in oceanic regions. Combining these predicted values with the tectonic reconstructions 
used in mantle circulation models is a powerful tool, as individual locations of interest can 
be rotated back to their palaeo-positions and variations of heat flow through time can be calculated.

Similarly, dynamic topography has important economic significance. Understand-
ing the mantle’s contribution to present day topography improves understanding of anomalous features in Earth’s topography, which provides a gateway to investigating palaeotopography and sea-levels. What contributes to dynamic topography is an area of active research. Debate still exists on the scales at which to calculate dynamic topography as well as what classes as dynamic topography. In this chapter I have shown that calculations from whole mantle density anomalies produce reasonable estimates of dynamic topography, that can be tested in a number of ways. It is difficult to have complete faith in any predictions of dynamic topography with so few measurements to compare to. However, with a few simple corrections and additions it is possible to produce a model of relief which is remarkably similar to Earth’s measured topography. This suggests there is significant merit to the approach of calculating dynamic topography presented here.

Frequently throughout this study of dynamic topography magnitudes are overestimated when compared to many other predictions of long wavelength dynamic topography. It seems viscosity may play a considerable role in the magnitudes of dynamic topography that can be supported. Further investigation into how mantle viscosity structure affects the topography generated from mantle density anomalies would provide significant insight into how much dynamic topography can be supported.

Length scales also seem to be a significant contributor to the magnitudes of dynamic topography predicted. Global scale models predicted larger absolute magnitude of dynamic topography, of the order of kilometres. Other studies on local to regional scales consider tens to hundreds of metres a more accurate estimate for the magnitude of mantle supported topography. Clearly this study only presents global scale models with large amplitudes of dynamic topography at long wavelengths and over long periods of geological time. It seems that further integration between these sorts of models and the more localised studies may begin to reconcile some of the differences in predicted dynamic topography.
Chapter 6

Summary

With modelling techniques becoming ever more sophisticated, new classes of models can investigate complex and detailed, small scale heterogeneity in inaccessible portions of our planet. Seismic tomography now provides a detailed snap shot of present day mantle conditions, imaging plumes of hot material rising from the core mantle boundary and subducted slabs sinking into the lower mantle from zones of tectonic convergence at the surface. Tomography of the core mantle boundary region reveals large, long lived regions of slower than average mantle, potentially related to chemical heterogeneity in the lowermost mantle. High-resolution tomography also images complex mid-mantle structure hypothesised to be caused by equally complex surface tectonics. For example, van der Meer et al. (2010) highlights 29 seismic fast anomalies related to subduction zones across the globe. More sophisticated models can be used to investigate some of these complex heterogeneities. In this thesis I have used mantle circulation models to further understand the interaction of surface plate tectonic processes and mantle convection and attempt to reconcile some of the imaged mid-mantle structure.

Mantle circulation models have proved a useful tool for studying Earth’s mantle. I
have investigated the effects of lateral and radial viscosity variations, mineral phase changes, heating mode and compressibility on the behaviour of mantle convection in a domain conditioned by plate tectonics. These global models focus on generating down-wellings which accurately model the faster than average mantle imaged by tomography. At higher resolution, models examining areas of tectonic convergence on a regional scale can be used to test detailed plate tectonic reconstructions. The dynamic link between complex ocean closures, such as the Neotethys Ocean can be explored by tracing particles from past subduction zones to their present day locations. Mantle circulation models can also be used to contribute to the ongoing debate on dynamic topography. In this thesis I have shown how estimates of present day topography can be made from the predictions of mantle circulation models. The predictions of dynamic topography proved to match reasonably well with other predictions made using different modelling techniques and a variety of tests showed the value of predictions both at present day and in the past.

Radial variations in viscosity have a considerable effect upon the nature of convection in mantle circulation models. An increase in viscosity into the lower mantle significantly alters the rate at which subducted material sinks. Model results presented in chapter three of this thesis indicate that the match between model temperature predictions and seismic tomography is highly sensitive to radial viscosity variations. In isoviscous simulations subducted material sinks quickly, reaching deeper into the mantle than seismology images suggest, therefore a more viscous lower mantle is required. I find that a viscosity increase of a factor of thirty produces a good match between model predictions and seismic velocity perturbations. A factor of thirty increase in viscosity is consistent with many published studies of mantle viscosity. A two order of magnitude, lower mantle viscosity increase results in a significant reduction in sinking velocity. As a result slabs do not reach the depths indicated by seismic tomography.
Lateral variations in viscosity, generated by including a temperature dependent viscosity law, have a less significant role in creating mantle circulation models that accurately match seismic tomography than might initially have been thought. Although mantle viscosity is likely to be temperature dependent, the simulations presented in this thesis do not differ hugely based upon lateral variations in viscosity. Instead, including temperature dependence in the model viscosity serves to modify the radial viscosity profile, which is clearly demonstrated to have a large effect on the rate of sinking of slabs and the match between model predictions and seismic tomography. These conclusions are similar to those made by Zhong et al. (2000), who also observed that convection models including both temperature and depth dependent viscosity generate shorter wavelength structures than other models containing one element or the other. This is confirmed in mantle circulation models, where there is no significant improvement in the match between seismic tomography and the model predictions when temperature and depth dependent viscosity is included.

Studies of mineral phase changes in the mantle transition zone generally suggest some degree of partial layering by an endothermic phase transition at 660 km depth. Over the course of Earth’s history the mantle may have transitioned from a more layered state to its present day convective regime. Phase changes affect mantle circulation models in different ways to standard mantle convection models. Wolstencroft & Davies (2011) showed that in Rayleigh number / phase buoyancy parameter space, models could develop into one of three convection modes: Whole mantle convection, two layer mantle convection or a transitional, partially layered regime. With mantle circulation models it is almost impossible to produce a model in the two layer convection regime, as down-wellings associated with plate tectonics eventually break through the layering, even at unrealistically high Clapeyron slope. The difference between the two classes of model indicates a fundamental difference in convection in a mantle with and without plate tectonics. The difference between modelled up-wellings
and down-wellings in mantle circulation models is also telling. At more negative Clapeyron slopes, up-welling flow is severely restricted by the phase transition, whilst down-wellings can eventually penetrate into the deep mantle, perhaps after a period of accumulation. Although temporal changes are not considered in this study, the fact that plate tectonics seems to contribute to the break down of layering could be an important consideration, particularly for the early Earth. In models with an exothermic phase transition included at 410 km depth alongside the endothermic phase change there was very little difference between the reference case and the modelled cases. In mantle circulation models containing phase transitions the negative Clapeyron slope transition is the dominant factor.

High resolution seismic tomography images complex mantle structure at mid-mantle depths under India and the surrounding oceans. Interpretations of faster than average mantle have hypothesised that the closure of the Neotethys ocean is as complex as the mantle structure. Regional geology of the India - Asia collision suggests that a back arc ocean opened off the Eurasian margin during the Mesozoic era, allowing for subduction at two individual convergent boundaries. One boundary at the Eurasian margin and a second subduction zone beneath the back arc ocean. Tomography reveals three north west to south east striking fast velocity anomalies. Interpretations of these seismic tomography studies suggest that a second intra-ocean subduction zone is required to produce the imaged mantle structure. However there remains some disagreement on the configuration and duration of each subduction. Mantle circulation models at high Rayleigh number, with parameters selected based upon the results of the parameter space investigation, produce mantle structure closely resembling seismic tomography and can be used to better understand the nature of Neotethys ocean closure.

Mantle circulation models with a surface velocity boundary condition containing a multi-phase closure of the Tethys ocean accurately reproduce the mantle structure imaged in
seismic tomography. Three north west to south east striking anomalies in the model predictions indicate that plate reconstructions including a Neotethys back arc ocean can produce slabs that match well to tomographic images. The differences between the model predictions and seismic tomography, as well as differences between different tomography studies highlight some key points. Mantle circulation model predictions match no single tomography study perfectly. Many differences can be attributed to resolution and data sources, some tomography models may have poor data coverage in oceanic areas, resulting in poor agreement between tomography studies. Filtering model predictions with the resolution operator of a specific tomography model allows for a good fit to that particular study and highlights the significant effect of the tomographic inversion process. Converting model predicted temperatures to seismic velocity results in an overestimate of the magnitude of seismic velocity anomalies, hinting that damping in tomography is an important consideration. Despite these limitations, mantle circulation models produce mantle structure similar to the patterns observed across a variety of tomography studies. This demonstrates the strength of using mantle circulation models to understand the interaction between plate tectonics and Earth’s mantle. Improvements can be made to account for some of the limitations of the method. For example, synthetic tomography models, accounting for bias and resolution issues in the inversion process, could be made from the predictions of mantle circulation models. Creating synthetic tomography models would allow a more direct comparison between the model predictions and seismic tomography images.

Tracer particles are used to track the location of mantle material subducted at convergent boundaries. This provides an important link between the surface processes and the deep mantle. Previously, the convergence zone of origin for each seismic fast anomaly could only be inferred from the geographical locations of plate boundaries and modelled anomalies. This method provides a new way to examine each fast anomaly in terms of tectonic
processes. Marker particles demonstrate a large quantity of faster than average mantle at mid-mantle depths originates at the intra-ocean subduction zone. This material is Neotethys Ocean lithosphere subducted beneath the Spongtang back arc. Neotethys lithosphere creates the majority of the imaged southern and central fast tomographic anomalies, referred to throughout chapter four as II and III. This interpretation is consistent with Hafkenscheid et al. (2006)’s interpretation based solely upon the tomography. Tracer particles reveal the northernmost anomaly to be comprised of two phases of subduction at the Eurasian margin, one prior to the opening of the back arc-ocean and one post the accretion of the arc onto the northward moving Indian continent. Although two subduction zones are required for mantle circulation models to match with imaged seismic structure, simultaneous subduction does not appear to be necessary. Further work could develop the tracer particle method to understand the sinking rate of slabs at different subduction zones by tracking particles based upon the age of the subduction zone. This could provide a test for the simple view that subducting material sinks at a fairly uniform rate throughout the mantle (e.g. Čížková et al. 2012, van der Meer et al. 2010) regardless of duration and geographical extent of convergent margins.

The northernmost imaged anomaly in the region is commonly hypothesised to be a results of Palaeotethys subduction. For one of the three anomalies to be associated with Palaeotethys subduction it would require slab material from a subduction zone which ceased around 140 million years ago to be present at mid-mantle depths. Mantle circulation models suggest that each of the anomalies at mid-mantle depths originate at subduction zones active after 140 Ma with a deeper colder than average region representing older subduction. The sinking of material subducted more than 140 million years before present beyond mid-mantle depths is compatible with both the surface tectonics and hypothesised slab sinking rates.

Global predictions of mantle flow induced topography indicate up to ±5 km of long
wavelength topography arising from density and buoyancy anomalies predicted by mantle circulation models. The models predict the dynamic component of topography to be heavily dominated by plate boundary processes, where the largest magnitude anomalies are found. The dynamic topography predicted by mantle circulation models is clearly influenced by near surface density differences, which many calculations of dynamic topography do not include. Despite this, the pattern of predicted dynamic topography to first order matches well to that predicted from instantaneous models of mantle flow based upon density variations derived from seismic tomography (e.g. Conrad & Husson 2009). It is difficult to resolve the issue of the magnitude of long-wavelength dynamic topography as mantle circulation models predict up to 5 km of flow induced topography whilst instantaneous models predict values closer to 2 km. Mantle viscosity is a considerable contributor to the magnitude of topography predicted. Instantaneous flow models including a low viscosity asthenosphere, or generally lower upper mantle viscosity may predict significantly lower amplitudes of dynamic topography (e.g. Conrad & Husson 2009, Lithgow-Bertelloni & Silver 1998). Predictions of dynamic topography from TERRA are limited by the lowest viscosity the code can use. Currently it is not possible to model mantle circulation with low viscosity upper mantle layers, as this would require higher resolution grids and in turn greater computational resources. The reference viscosity chosen for the upper mantle in these mantle circulation models is higher than used for some instantaneous flow calculations.

Dynamic topography predictions from mantle circulation models demonstrate patterns clearly influenced by near surface density variations. Other predictions from instantaneous flow models resemble a degree two pattern similar to that imaged in the lowermost mantle by S-wave seismic tomography studies, suggesting that instantaneous flow models are more influenced by lower mantle structure. A common process when calculating dynamic topography is to disregard density variations in the upper 300 km of the mantle, therefore
explaining some of the observed differences. The topography calculated from mantle circula-
tion models includes density anomalies for the whole mantle, so the observed significance of
plate boundary processes to the calculated topography is explained. Despite the differences
in method, the general pattern of dynamic topography is similar across both methods.

The definition of dynamic topography is important for work on the phenomenon to
continue. If one was to define dynamic topography as any topography which is unexplained
by surface processes then it is sensible to exclude near surface density variations. In this
case depth to the ocean floor is adequately explained by the depth versus age relationship
and would not be considered dynamic. However, if one defines dynamic topography as
any topography arising from convection within the mantle, including near surface density
variations is important. The buoyancy of young oceanic lithosphere and its subsequent
cooling is considered part of the mantle convection process, thus the resulting topography
can be considered to be mantle derived.

Length scales are also important in dynamic topography. Field studies and in-
versions of seismic reflection data suggest magnitudes of dynamic topography an order of
magnitude or more lower than global predictions of dynamic topography (e.g. Pritchard et al.
2009, Roberts & White 2010, Wheeler & White 2002). Further work is clearly required to
reconcile the differences in predicted dynamic topography at different length scales. Further-
more, predictions of dynamic topography back in time could provide a useful constraint on
palaeo-topography and sea-level. Of course it is crucial to understand dynamic topography
at present day, before relying too heavily on estimates for palaeo-topography.

Collecting the suggestions for future work from throughout this summary, I suggest
a number of future studies which could develop out of this work. This study considered a
limited number of models with endothermic phase transitions as part of a larger parameter
space investigations. Since endothermic phase transitions appear to have a significant effect
on mantle circulation models it would be useful to examine more models, particularly ones at higher Rayleigh number. This would help to define curves, similar to those of Wolstencroft & Davies (2011), between the different convective regimes. Further models of this nature could prove to have significant implications for the study of Earth, particularly if plate tectonics could be shown to have a significant impact of the nature of layered convection.

The combination of high-resolution mantle circulation models and plate tracking marker particles could be used for studying a number of other areas. The most obvious example would be the Farallon region, where the Farallon Ocean subducted beneath North America. At regional scale, the tectonics of the Farallon region seems more straightforward than the Tethys regions, however the particle tracking method could be used to confirm the mid-mantle slabs are related to the expected surface processes. Liu & Stegman (2011) show that in the shallow mantle the Farallon slab is segmented. Careful application of the mantle circulation and marker particle method for this region could provide further useful information on the tectonics creating this complex mantle structure. Tracer particles could also be used to track the sinking rate of subducted material. A modification of the particle method could be used to label particles on a plate according to the age of the ocean crust subducted. This would be important to understand if older oceanic lithosphere subducts faster than younger lithosphere. The method could also investigate differences in the sinking rate of material subducting at boundaries with different geometry.

Clearly, the work on dynamic topography presented in this thesis is preliminary. Although it contributes to the ongoing debate on the nature of mantle supported topography, there is plenty of potential for further study on this topic. One of the most obvious is to attempt to reconcile the differences between global and regional scales for calculating dynamic topography. There is often an order of magnitude difference in topography observed by field methods and global modelling methods. It may be that localised studies are
observing the changes of individual basins or domes within a larger background dynamically supported feature. Therefore it seems important to reconcile the differences between global and regional studies of dynamic topography. Mantle viscosity also appears to be a significant contributor to the magnitude of dynamic topography predicted from mantle circulation models. A detailed investigation of how mantle viscosity, particularly radial variations and absolute values would be useful to further understand the magnitude of mantle flow supported topography.
Bibliography


## Appendix A

## Glossary

### Mathematical Symbols

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Units</th>
</tr>
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<tbody>
<tr>
<td>$C_p$</td>
<td>Specific heat at constant pressure</td>
<td>$J kg^{-1} K^{-1}$</td>
</tr>
<tr>
<td>$g$</td>
<td>Gravitation acceleration</td>
<td>$ms^{-2}$</td>
</tr>
<tr>
<td>$H$</td>
<td>Radiogenic heat production</td>
<td>$W m^{-3}$</td>
</tr>
<tr>
<td>$k$</td>
<td>Thermal conductivity</td>
<td>$W m^{-1} K^{-1}$</td>
</tr>
<tr>
<td>$P$</td>
<td>Pressure</td>
<td>$Pa$</td>
</tr>
<tr>
<td>$T$</td>
<td>Temperature</td>
<td>$K$</td>
</tr>
<tr>
<td>$h$</td>
<td>height above sea level</td>
<td>$m$</td>
</tr>
<tr>
<td>$r$</td>
<td>thickness of continental root</td>
<td>$m$</td>
</tr>
<tr>
<td>$z$</td>
<td>dynamic topography</td>
<td>$m$</td>
</tr>
<tr>
<td>$u$</td>
<td>fluid velocity</td>
<td>$ms^{-1}$</td>
</tr>
<tr>
<td>$V_p$</td>
<td>Velocity of P-wave</td>
<td>$ms^{-1}$</td>
</tr>
<tr>
<td>$V_s$</td>
<td>Velocity of S-wave</td>
<td>$ms^{-1}$</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>Volume co-efficient of thermal expansion</td>
<td>$K^{-1}$</td>
</tr>
<tr>
<td>$\gamma$</td>
<td>Grüneisen parameter</td>
<td></td>
</tr>
<tr>
<td>$\eta$</td>
<td>Dynamic viscosity</td>
<td>$Pa S$</td>
</tr>
<tr>
<td>$\kappa$</td>
<td>Thermal diffusivity</td>
<td>$m^2 s^{-1}$</td>
</tr>
<tr>
<td>$\rho$</td>
<td>Density</td>
<td>$kg m^{-3}$</td>
</tr>
<tr>
<td>$\Delta \rho$</td>
<td>Density difference</td>
<td>$kg m^{-3}$</td>
</tr>
<tr>
<td>$\tau_{ij}$</td>
<td>Deviotoric stress</td>
<td>$Pa$</td>
</tr>
<tr>
<td>$\sigma_{ij}$</td>
<td>Normal stress</td>
<td>$Pa$</td>
</tr>
<tr>
<td>$\delta_{ij}$</td>
<td>Kronecker delta</td>
<td></td>
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</table>
Abbreviations

CMB The core-mantle boundary
GDU Geodynamic Unit
MCM Mantle circulation model
TERRA Convection code used to simulate Earth’s mantle in three dimensional spherical geometry
UNIL Universite de Lausanne, the university where the plate motion history used in many of the presented models was developed
Appendix B

The Geologic Time Scale

Earth is over 4.5 billion years old. To help geologists the time between formation and present day is divided into a number of Eons, Eras, Periods and Epochs. The following graphic illustrates the divisions and shows the locations of the plate tectonic reconstructions used in this study. Major geologic events such as mass extinctions mark the boundaries between eras (e.g. Permian - Triassic, Cretaceous - Tertiary). Smaller extinctions or changes in the fossil record mark boundaries between periods (e.g. Jurassic - Triassic).
Figure B.1: Graphic representation of the geologic time scale, red arrows represent the timings of UNIL reconstructions
Appendix C

Palaeogeography and tectonics

The following figures illustrate the plate tectonic reconstructions used as the surface velocity boundary condition for mantle circulation models throughout this thesis. Yellow colours represent continental lithosphere and blue colours oceanic lithosphere. Lithospheric plates are separated by tectonic boundaries are marked with lines of various colours, pale blue lines are mid ocean ridges, thicker blue lines represent passive margins and purple lines are convergence zones. Continents are made up from geodynamic units (GDUs), described in chapter 2, the boundaries between GDUs are marked on the maps. Present day geography is marked on each map, rotated back in time for reference. Reconstructions are provided courtesy of the University of Lausanne, the maps used to generate these figures were provided by C. Hochard.
Figure C.1
Appendix C: Palaeogeography and tectonics

Figure C.1
Figure C.1