How the Relationship Between Subduction and Mantle Dynamics Shapes Slab Evolution at Mid-Mantle Depths

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I, Antoniette Greta Grima, confirm that the work presented in this thesis is my own. Where information has been derived from other sources, I confirm that this has been indicated.
Abstract

Advances in seismic tomography have revealed that the Earth’s mantle is more complex than previously thought and hosts a plethora of slab morphologies. These morphologies can be broadly grouped into three main types; penetrating slabs, deflected slabs and broken, or orphan, slabs. Slab orphaning is a newly discovered phenomenon whereby slabs break directly at mid-mantle depths. This produces a flattened parent slab above 660 km and an orphan slab below it. As the orphan slab slowly sinks towards the core-mantle boundary, subduction continues through the lateral motion of the parent above 660 km. In nature, the Tonga, Arabian, Japan and Central American slabs are possible candidates for slab orphaning. Orphaning has significant implications for the interpretation of slab remnants and their inferred ages, with consequences for tectonic reconstructions.

The subduction of slabs, however, does not take place in isolation and slab dynamics must be influenced to some degree by the overriding plate. The nature of the overriding plate plays a major role in the evolution of deep slab morphologies at mid-mantle depths. 2-D numerical simulations of subduction indicate that the presence of continental lithosphere at subduction zones produces markedly different slab behaviour at depth. In particular, a continental overriding plate results in bigger orphans and encourages the mid-mantle penetration of slabs that are otherwise inclined to flatten. It is therefore clear that, in contrast to the slabs of the upper mantle, the deeper slab morphologies are the result of a complex interaction between the overriding plate forcing and the changes in the relative strength ratios of the slab and mantle.
Impact Statement

The sinking of surface lithospheric tectonic plates into the Earth’s mantle is a fundamental process on Earth. Known as subduction, this process is thought to be unique to our planet and one that enables it to host life. The surface lithospheric plates, also known as slabs once they subduct into the mantle, undergo pressure and temperature changes with depth. The interaction of slabs with the surrounding mantle can determine their behaviour and morphology at depth which, in turn, can directly shape the surface evolution of our planet.

This study focuses on the evolution of slabs at mid mantle depths. It analyses the relationship between the slab strength and that of the surrounding ambient mantle and, in doing so, identifies a new type of slab-mantle interaction; slab orphaning. Slab orphaning describes, for the first time, how slabs can break off directly at mid-mantle depths, much deeper than previously thought. This new type of slab behaviour sheds new light on the slab mantle interaction at depth. Furthermore, slab orphaning also has important implications for plate reconstructions and seismic tomography that currently assume only shallow slab break-offs. Features like Tonga, Japan, Arabian and Central American remnant slabs must, for example, likely be reinterpreted as orphan slabs that broke off at depth. This study also explores the link between surface features such as the presence of a continental lithosphere with the deep slab behaviour. Results indicate that continents are extremely influential in determining the slab behaviour at 660 km depth and more.
The outcomes of this thesis are vital for the understanding of our planet's interior dynamics and how these can shape the surface we live on. An understanding of deep slab dynamics is fundamental to provide better constraints on the mechanisms of some of the deepest earthquakes on Earth, such as those measured at the Tonga subduction zone at 700 km depth. Earthquakes can have devastating effects on populations living close by. The results of this study will provide interdisciplinary benefit to a number of fields in Earth Science, in particular, seismology and mineral physics and most importantly the study and reconstruction of past plate motions. The motions of plates many millions of years ago are the basis for the reconstruction of continental positions through time and crucial to inferences of past sea-level and paleoclimate.

This thesis has sparked several scientific discussions and invited talks at major international conferences. These have been communicated to the public via outreach events at UCL, including student engagement days and the GeoBus initiative which brings Earth Sciences learning directly to schools, inspiring students to follow up STEM subjects.
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## Contents

1 Introduction  
1.1 Subduction Dynamics ........................................... 16  
1.2 What Shapes Subduction?  ......................................... 18  
  1.2.1 The Overriding Plate ...................................... 19  
  1.2.2 Subduction and Trench Behaviour ........................... 20  
  1.2.3 Slab Mineralogy and Rheology ............................... 21  
1.3 Slab Morphologies of the Mid-Mantle  ............................ 26  
  1.3.1 Mid-Mantle Slab Remnants .................................. 28  
1.4 Outline ................................................................. 31  
  1.4.1 Main aims ..................................................... 31  
  1.4.2 Thesis structure ............................................... 32

2 Methodology  
2.1 Governing Equations ............................................... 33  
2.2 General Description of the Code ................................ 36  
2.3 Physical Model ..................................................... 37  
2.4 Rheology .............................................................. 38  
2.5 Numerical Model .................................................... 45  
2.6 Boundary and Initial Conditions ................................ 46  
2.7 Continental Overriding Plate Implementation ................. 48  
2.8 Diagnostics and Visualisations .................................. 49
3 Slab Orphaning as a Mechanism for Mid-Mantle Slab Break-Off .......................... 51
   3.1 Introduction ................................................. 51
   3.2 Methods ..................................................... 54
   3.3 Results ...................................................... 55
   3.4 Discussion ................................................... 58
      3.4.1 Orphan Slabs in Tomography ......................... 61

4 The Effect of a Continental Overriding Plate on Subduction Dynamics .............. 65
   4.1 Methods ..................................................... 67
   4.2 Results ...................................................... 69
      4.2.1 Slab Morphologies ..................................... 69
      4.2.2 Plate Velocities ....................................... 75
      4.2.3 Slab Tip Evolution ..................................... 78
      4.2.4 The Relationship between the Trench Velocities and the Slab Tip Behaviour 79
      4.2.5 The Overriding Plate and Slab Evolutionary Stages ......... 82
   4.3 Discussion ................................................... 85

5 Slab Orphaning Regimes: The Missing Link Between Penetrative and Deflective Slab Morphologies .................................................. 92
   5.1 Introduction .................................................. 92
   5.2 Methods ..................................................... 94
   5.3 Results ...................................................... 96
      5.3.1 Slab Strength ........................................... 98
      5.3.2 Subduction Angles ....................................... 101
      5.3.3 Orphaning in an Intra-Oceanic Subduction .................. 101
      5.3.4 Orphaning in a Continent-Ocean Subduction .............. 104
   5.4 Discussion ................................................... 108
6 Conclusions and Further Work

6.1 Conclusions .......................................................... 114
6.2 The Continental Overriding Plate and Deep Slab Morphologies . . . 115
6.3 Slab Orphaning and Mid-Mantle Slab Break-Off .......................... 116
6.4 Slab Orphaning Regimes .............................................. 117
6.5 Final Conclusions and Future work .................................... 118

Appendices

A Limitations, Assumptions and Evaluations

A.1 Limitations and Assumptions ........................................... 120
  A.1.1 The Numerical Simulations ....................................... 120
  A.1.2 The Vertical Viscosity Profile ................................... 121
A.2 Orphaning Tests and Evaluations .................................... 122
  A.2.1 Initial Conditions .................................................. 122
  A.2.2 Resolution Tests .................................................. 124
  A.2.3 Rheological Sensitivity .......................................... 124

B Movies

Bibliography
List of Figures

1.1 Schematic diagram illustrating the major slab morphologies of the mid-mantle. ........................................... 17
1.2 Phase proportions taken from Stixrude and Lithgow-Bertelloni (2012), showing the phase changes in a pyrolitic mantle. ........... 23
1.3 Viscosity profiles of the mantle. ......................................... 25
1.4 Tomographic cross sections of the Tonga and Japan orphan slabs... 30
2.1 Schematic model set-up for \( V_1 \) ........................................ 42
2.2 Schematic model set-up for \( V_2 \) ........................................ 42
2.3 Schematic model set-up for \( V_3 \) ........................................ 43
2.4 Schematic model set-up for \( V_4 \) ........................................ 43
2.5 Schematic showing some of the diagnostics and measurements that can be made at a subduction zone, using StagLab ........... 50
3.1 Zoomed in view of the morphological evolution of the slab for all three model set-ups \((V1 \rightarrow V3)\) ................................. 56
3.2 Slab sinking and trench retreat velocities for each model set-up \((V1-V3)\). .................................................. 57
3.3 Schematic diagrams of slab orphaning. .............................. 59
3.4 Density, stress and deformation mechanisms for \( V_3 \) ............... 60
3.5 Strain rate for \( V_3 \). Each panel represents a time slice in the model evolution. ............................................. 60
3.6 Tomographic cross-sections of the (a) Tonga, (b) Japan, (c) Arabian and (d) Central American slabs, showing a flattened slab at 660 km depth and a deeper independent fragment.

4.1 A continental overriding plate encourages slab tip penetration and anchoring below 660 km depth for model V1 (a). The same slab in V1 flattens above 660 km depth when the overriding plate is oceanic (b).

4.2 Considerable slab penetration and anchoring is also observed for V4 when the overriding plate is continental in nature (a). Similar to V1, V4 exhibits slab flattening when subduction occurs in an intra-oceanic setting (b).

4.3 Slab morphology for viscosity profile V2, showing a ‘hooked’ slab geometry in the presence of a continental overriding plate (a). Slab morphology, however, assumes a quasi-vertical shape when the upper overriding plate is oceanic (b).

4.4 Orphaning occurs for both continental (a) and oceanic (c) overriding plates. A continental overriding plate however encourages bigger slab orphans and longer orphaning times (b vs. d) compared to its oceanic counterpart.

4.5 Variations in slab sinking and trench velocities in cm year$^{-1}$ for models V1-V4 with continental (a,b) and oceanic (c,d) overriding plates.

4.6 Trench velocities in cm year$^{-1}$ (blue) and slab tip angles in degrees (orange) for V1 with continental (a) and oceanic (b) overriding plate and for V2 with continental (c) and oceanic (d) overriding plate.

4.7 Trench velocities in cm year$^{-1}$ (blue) and slab tip angles in degrees (orange) for V3 with continental (a) and oceanic (b) overriding plate and for V4 with continental (c) and oceanic (d) overriding plate.

4.8 Trench velocities in cm year$^{-1}$ (blue) and ratio of slab bending $S_b$ (orange) for models V1-V4 with a continental overriding plate.
4.9 Schematic of the slab bending ratio $S_b$.

4.10 Viscosity plots showing the induced viscous flow patterns for a continental overriding plate (a-c) and an oceanic overriding plate (d-f).

4.11 Plot of density illustrating a lens of lower density within the slab core at 660 km depth and its evolution in systems with a continental overriding plate (a-c) and an oceanic overriding plate (d-f).

5.1 Regime diagram for orphaning, penetrating and deflected slab morphologies.

5.2 Slab orphaning for slab yield stress values of 600 MPa (a), 400 MPa (b) and 200 MPa (c) with respective Clapeyron slope values of $-2.5 \times 10^6 \text{ PaK}^{-1}$, $-2.0 \times 10^6 \text{ PaK}^{-1}$, $-1.5 \times 10^6 \text{ PaK}^{-1}$.

5.3 Slab morphologies for subduction angles of 20° to 90° and oceanic overriding plate.

5.4 Slab morphologies for subduction angles 20° to 90° in the presence of a continental overriding plate.

5.5 Slab orphaning with an oceanic overriding plate and steep subduction angles of 90°, compared with a tomographic cross-section for the Marianas slab, generated using the submachine portal from Hosseini et al. (2018).

5.6 Shallow slab break-off for model $V3(80^\circ)$ with slab morphologies on the left and strain rates on the right.

A.1 Subduction of a more naturally bent slab with a 300 km bending radius, replaces the initial slab kink of our standard models.

A.2 Orphaning of a naturally more bent slab with a bending radius of 300 km. This indicates that orphaning is not a consequence of the initial slab kink implemented in our standard models.

A.3 Increasing model width and aspect ratio, does not hinder orphaning, indicating that orphaning is not the result of side boundary forcing.
List of Figures

A.4 Zoomed in view of slab orphaning at ultra-high resolution clearly indicating that slab orphaning is not the product of an under-resolved model. ................................................. 124
A.5 Slab orphaning in ultra-high resolution for a 5 km thick weak crustal layer. .................................................. 125
A.6 Decreasing our standard crustal thickness from 15 km to 10 km, reproduces the general dynamics. In this case however, the orphan slab larger than that those observed for a thicker crust. ............. 125
A.7 Slab orphaning for a crustal thickness of 15 km. Note the size of the orphan for this simulation with that orphan from Fig. A.6. ....... 126
A.8 Slab orphaning for a crustal depth of 400 km, standard for all models run and examined in this work. ................................. 127
A.9 Slab orphaning for a crustal depth of 300 km. Orphaning is unaffected, despite the relatively shallow depth of the weak crustal layer. 127
A.10 Slab orphaning for a crustal depth of 1000 km. The deep crustal layer, reduces the size of the orphan due to its role in reducing the coupling between slab and mantle. ............................... 128
A.11 Slab orphaning with a continuous crust. Orphaning occurs regardless of the depth of the weak crustal layer indicating that orphaning is independent of depth of the weak crustal layer. ................. 128
A.12 The removal of the weak crustal layer results in intense coupling between the upper and lower plates, hindering subduction. ........ 129
List of Tables

2.1 Implemented viscosity profiles ........................................ 41
2.2 Model Parameters ...................................................... 44
2.3 Continental Overriding Plate Parameters .......................... 48

4.1 The mantle viscosity profiles tested for the continental overriding plate implementation ................................................. 68
4.2 The viscosity profile and associated slab morphology in the presence of a continental overriding plate ............................... 70

5.1 Model variations and different initial conditions .................. 95
5.2 Orphan sizes, orphaning timescales and associated trench retreat compared across orphan variants with different slab strengths and Clapeyron slopes .......................................................... 101
5.3 Orphan sizes, orphaning timescales and associated trench retreat compared across all orphan variants in an intra-oceanic subduction. 103
5.4 Orphan sizes, orphaning timescales and associated trench retreat are compared across all orphan variants in a continent-ocean subduction. 107
Chapter 1

Introduction

1.1 Subduction Dynamics

The term subduction describes the sinking of cold, dense oceanic crust and lithosphere as a coherent slab into the asthenospheric mantle that underlies it (White et al., 1970; Billen, 2008). The oceanic plate makes up the topmost thermal boundary of our planet and contains most of the thermal energy that drives convection (Crameri et al., 2018). The cooling of oceanic lithospheric plates and their eventual descent into the Earth’s mantle is the planet’s most efficient heat loss mechanism (e.g. van Hunen and Moyen, 2012; Crameri et al., 2018, and references therein). The subducting slab is the vehicle that drives plate tectonics and enables the volatile exchange between the biosphere, atmosphere and the planet’s interior (Stern, 2002; Billen, 2008; Stadler et al., 2010).

Consistent and widespread deep subduction, pursuant to current scientific knowledge, is unique to our planet (King, 2001; Stern, 2004; Crameri et al., 2018). Subduction and surface recycling throughout Earth's history is considered crucial for the emergence and evolution of life (Albarède, 2009; Stern, 2002). Moreover, subduction is also the single, major mechanism for the deformation and reorganisation of our planet's surface (Tackley, 2000a). The subducting oceanic lithosphere provides ~ 90% of the force required to drive plate tectonics (Forsyth and Uyeda, 1975). Through subduction, lateral density heterogeneities are introduced into the mantle, exciting viscous flow which is then translated to the surface through shear
traction at the base of the plate (Lithgow-Bertelloni and Richards, 1998; Conrad and Lithgow-Bertelloni, 2002, 2004).

The forces driving the lithospheric motions depend on the balance between the buoyancy forces driving subduction, and the mechanisms, viscous stresses and buoyancy forces resisting subduction (Billen, 2008). Understanding how these forces change as a function of time, depth and strength is key to the understanding of the dynamic relationship between the surface, the subducting slab and the mantle. The various slab morphologies observed through seismic tomography models (c.f. Fig. 1.1) attest to the complex nature of this relationship between the subducting slab and the surrounding, ambient mantle.

![Figure 1.1](image.png)

**Figure 1.1:** Schematic diagram illustrating the major slab morphologies of the mid-mantle. With these being shallow (c) and deep deflected (a) slabs at depths of 660 km and 1000 km respectively, penetrating slabs (d) and orphaned slabs (b). Each morphology is accompanied with examples from the seismic tomography models of Fukao et al. (2009).
1.2 What Shapes Subduction?

It is still not clear when and how subduction initiated (Condie et al., 2016). Geochemical ‘arc’ signatures in Archean greenstone belts have been interpreted as evidence for Archean subduction (e.g. van Hunen and van den Berg, 2008; Halla et al., 2009; van Hunen and Moyen, 2012; Condie et al., 2016, and references therein). However, it is likely that the subduction of the thicker Archean oceanic lithosphere occurred intermittently, through delamination, peeling and drips (Silver and Behn, 2008; van Hunen and van den Berg, 2008; Halla et al., 2009; Moyen and van Hunen, 2012; van Hunen and Moyen, 2012; Foley, 2018). Geochronological and geochemical data from the Jack Hills zircons point to evidence of modern day like subduction in the Hadean (Compston and Pidgeon, 1986; Maas et al., 1992). Yet, considering the uniqueness of the data, it is not clear whether the Jack Hills zircons record isolated subduction or whether they represent evidence of a more widespread process similar to the present day plate tectonics. The recycling of the surface through subduction or subduction-like processes inherently destroys the evidence of its own existence. Therefore, it is extremely difficult to pin down when and how the Earth evolved from the intermittent proto-subduction of its early, hotter history to the modern day, long-lived subduction, that characterises plate tectonics (Moyen and van Hunen, 2012; Bercovici and Ricard, 2014). The question of how to initiate new subduction zones, and whether the planet’s current convective regime is able to support the formation of new incipient subduction, away from any pre-existing weakness, is another non-trivial point and the focus of much debate (e.g. Stern, 2002; Gerya et al., 2015; Stern and Gerya, 2018).

Despite the ubiquitous presence of subduction zones and the central role of the subducting plate in sculpting our planet, subduction dynamics are still poorly understood. To further complicate the picture, advances in seismic tomography in the last couple of decades have brought to the forefront the richness and diversity of slab morphologies at various depths in the mantle (c.f. Fig. 1.1 and e.g. Fukao et al., 2001, 2009; Fukao and Obayashi, 2013; Goes et al., 2017). Previous work (e.g. Fukao et al., 2001, 2009; Stegman et al., 2006; Torii and Yoshioka, 2007;...
1.2. What Shapes Subduction?

Čížková et al., 2007; Schellart et al., 2007; Fukao and Obayashi, 2013; Goes et al., 2017; Petersen et al., 2017) has sought to tease out patterns of slab behaviours and draw correlations with observables such as trench motion, trench width, slab age and strength, slab dip and overriding plate type, amongst others. However, there is no clear correlation between the slab morphology and these other subduction parameters (Jarrard, 1986), which suggests that slab dynamics are the result of a combination of factors rather than one single relationship (Billen, 2008).

1.2.1 The Overriding Plate

By definition, subduction occurs in the context of an upper and lower plate. Although there are few examples of ocean-ocean collisions, most subduction zones on Earth are characterised by oceanic lithosphere subducting beneath thick and buoyant continental lithosphere. The latter is highly heterogeneous with variations in thickness, buoyancy and strength across subduction zones (Sharples et al., 2014). However, previous numerical studies tend to ignore the effect of the overriding plate, despite the influence this must exert on the down-going slab. When an overriding plate is included in numerical simulations, models produce more realistic trench behaviour and improved fits to observations (King, 2001; Butterworth et al., 2012; Sharples et al., 2014; Holt et al., 2015). Modern numerical simulations of subduction that do include an overriding plate (e.g. Garel et al., 2014; Crameri et al., 2017) tend to implement an ocean-ocean convergence. However, rock deformation experiments show that oceanic and continental lithosphere have different strengths and deformation styles (Kohlstedt et al., 1995; Jackson, 2002), suggesting that the continental overriding plate impacts the subduction zone differently from its oceanic counterpart.

Limited previous work on the subject indicates that the presence of a continental overriding plate promotes slab penetration into the lower mantle (Sharples et al., 2014; Holt et al., 2015; Crameri and Lithgow-Bertelloni, 2018). Sharples et al. (2014); Holt et al. (2015) and Crameri and Lithgow-Bertelloni (2018) also find that the presence of a continental overriding plate promotes slower trench mobility and slab sinking rates. The latter note that the addition of a continental overriding
plate in 2-D numerical models does little to change the overall geometric evolution of the model in the upper mantle. However, they observe that with the addition of a continental overriding plate, important subtleties in slab behaviour at depth emerge. First and foremost, a continental overriding plate is much less mobile than its oceanic counterpart. This favours steeper slab dip angles. Secondly, the slab dynamics and deformation patterns within the transition zone are markedly different for an oceanic and a continental overriding plate.

The deep continental roots (Conrad and Lithgow-Bertelloni, 2006; Van Summeren et al., 2012; Crameri and Lithgow-Bertelloni, 2018) introduce additional forcing to the subducting system through deep, viscous coupled flow. The strong coupling of the continental lithosphere directly affects the slab behaviour at depth and encourages the impinging of the continent on the trench (Van Hunen et al., 2001). Therefore, in the presence of a continental overriding plate the trench motion is an expression of both the slab sinking behaviour and the overriding plate push force.

1.2.2 Subduction and Trench Behaviour

As a first order observable that is easily traced to the seismically imaged slab, trench migration has received considerable attention from both numerical and analogue studies (e.g. Funiciello et al., 2003; Stegman et al., 2006, 2010; Boutelier and Cruden, 2013). Trench mobility has, in fact, been considered by some (e.g. Goes et al., 2017, 2011; King, 2001) as a major determinant of slab behaviour and morphology.

Lallemand et al. (2008) propose that trench migration is controlled by the sinking velocity of the slab which, in turn, is controlled by the age of the slab at the trench. The authors suggest that due to increasing slab stiffness and resistance to bending with age, older and faster subducting slabs tend to exhibit steeper slabs which favour trench advance. Similarly, Stegman et al. (2010) note that when weak slabs interact with a barrier to their downward descent they tend to flatten if their trenches retreat at the surface. If the trench is stationary then the slabs subduct vertically to pile and fold atop the lower mantle.
1.2. What Shapes Subduction?

The type of trench motion, and whether the trench retreats or advances, has also been interpreted as indicative of the slab behaviour at depth. Relatively stable trenches in general, have been associated with steep slabs that penetrate into the lower mantle (Christensen and Yuen, 1985; Billen and Arredondo, 2018) while retreating trenches are normally associated with flattened and deflected slabs at mid-mantle depths (Goes et al., 2017). However, global compilations of subduction zones show that the slab dip angle is only very weakly correlated to trench migration (King, 2001). In nature, penetrating slabs make up approximately half of the slabs subducting into the mantle, and yet trench advance is extremely rare for the modern day Earth (Schellart, 2008; Čížková and Bina, 2015). Furthermore, numerical models show that only a very restrictive range of plates with smaller buoyancy tend to exclusively favour trench advance (Stegman et al., 2010).

1.2.3 Slab Mineralogy and Rheology

The lithosphere is perhaps one of the most obvious and important stakeholders in shaping subduction kinematics and slab dynamics. Work by Petersen et al. (2017) suggests that the strength of the lithosphere and the subducted slab exerts a singular control on the shape of the slab. Similarly, Lallemand et al. (2008) conclude that younger and slower subducting slabs are weaker and, therefore, more prone to deformation and trench retreat. Results from 3-D numerical models of free subduction show that the role of slab flexure and buoyancy is crucial in shaping slab morphology, trench motion and trench curvature (Stegman et al., 2010).

However, slab buoyancy and strength are mainly controlled by the slab’s composition and temperature. The latter is determined by the age of the subducting plate at the trench (Crameri et al., 2017) which, in turn, has been correlated by previous work to the depth of the trench and overriding plate strain (Jarrard, 1986; Zhong and Gurnis, 1994; Gurnis et al., 1996; Lallemand et al., 2005).

The subducting slab is a multimineralic package, characterised by a complex chemical structure and extreme compositional differences that undergo continuous and discontinuous phase changes, as pressure and temperature conditions change with depth (see Fig. 1.2, and Anderson, 1987; Irifune and Ringwood, 1993; Poli
and Schmidt, 2002; Stixrude and Lithgow-Bertelloni, 2012). The inherent heterogeneities of the slab are manifested as large variations in temperature, melt content, stress and grain size, resulting in a complicated rheological profile that varies both laterally and with depth (Karato et al., 2001; Billen and Hirth, 2007).

Anomalously high seismic-wave velocities indicate that the down-going, subducting slab is colder than the mantle it is sinking into (Ricard et al., 2005). The thermal contrast between the cold slab and the surrounding warm mantle is what, essentially, drives the slab’s negative buoyancy. The temperature contrast between the two implies that the equilibrium depths of chemical reactions, such as the phase transition between low and high pressure minerals (Fig. 1.2), will be different for the slab and the mantle (Bina et al., 2001). Furthermore, due to the temperature dependence of the Clapeyron slope, the thermal gradients that exist within the slab encourage the delay of the ringwoodite to bridgmanite and ferropericlase phase transitions in the cold slab interior. Such behaviour results in the metastable persistence of low pressure mineral phases well into the stability fields of their high pressure counterparts (Bina et al., 2001; Poli and Schmidt, 2002; Billen, 2008, 2010). The rheological strength of the slab is, therefore, non-uniform and best described as a depth-dependent integrated property across the viscously deforming outer portions and the plastically deforming core (Billen, 2010).

Similar to the subducting slab, the rheology of the mantle also varies with changing depth, pressure and temperature conditions (see Fig. 1.2). On large geological time-scales, the Earth’s mantle behaves as a very viscous fluid. Deformation of the mantle material under stress can occur through brittle, plastic and creep deformation (Turcotte and Schubert, 2014). The main deformation mechanisms pertinent for the mantle at large are diffusion and dislocation creep (van Hunen et al., 2005; Ricard and Lyon, 2007; Turcotte and Schubert, 2014). The first case describes the diffusion of atoms through a crystal lattice and leads to a Newtonian rheology with a viscosity that is heavily grain size dependent. Dislocation creep describes the migration of microscopic planar imperfections within the crystalline lattice. This creep mechanism is highly stress dependent and leads to a non-linear power law
viscosity. Both diffusion and dislocation creep are thermally activated processes, meaning that the mantle’s viscosity is highly temperature dependent. Moreover, similar to slab strength, the mantle viscosity is also highly dependent on pressure, melt content, water content, mineralogical phases and oxygen fugacity (Ricard and Lyon, 2007, and references there in).

![Phase proportions](image)

**Figure 1.2:** Phase proportions taken from Stixrude and Lithgow-Bertelloni (2012), showing the phase changes in a pyrolitic mantle. Abbreviations are as follows; plg = plagioclase, cpx = clinopyroxene, opx = orthopyroxene, wa = wadsleyite, ri = ringwoodite, pv = perovskite and ppv = postperovskite. The green and dotted blue and red lines represent the shear wave velocities of harzburgite, pyrolite and basalt respectively.

The viscosity and density structure of the Earth’s upper and lower mantle can be constrained by the long wavelength non-hydrostatic geoid. Using transdimensional hierarchical Bayesian inversions of the gravity field together with thermodynamically inferred velocity-density scalings, Rudolph et al. (2015) suggest that the mantle’s viscosity increases at 1000 km depth, rather than at the canonical 660 km depth (see Fig. 1.3). These results confirmed and reinforced previous work (Forte and Peltier, 1991; King and Masters, 1992; Kido and Čadek, 1997; Kido et al., 1998)
that suggested deeper depths for the viscosity interface and proposed that the viscosity increase of the lower mantle does not necessarily always coincide with the 660 km depth seismic discontinuity (e.g. Ringwood, 1975). Rudolph et al. (2015), Kido and Čadek (1997) and Kido et al. (1998), further suggest that the viscosity increase at 1000 km depth is preceded by a low viscosity layer (LVL) between 660 and 1000 km depth (see Fig. 1.3). Recent work by Mao and Zhong (2018) finds that a low viscosity layer underlying the endothermic phase change is necessary to reproduce most of the slab morphologies observed in seismic tomography.

This new mantle radial viscosity profile clearly indicates that the relative strength of the slab with respect to the mantle would change as a function of depth as would the resistance to penetration. Furthermore, the variation of the slab strength both laterally and with depth, is a clear indication that the slab-mantle strength relationship is highly variable, dynamic and controlled to a large extend by the local slab strength at a particular depth. Slab morphology at large is therefore a representation of the relative local (depth specific) strength of the slab $vis - a - vis$ that of the ambient mantle.
1.2. What Shapes Subduction?

Figure 1.3: Viscosity profiles of the mantle computed using transdimensional hierarchical Bayesian inversions of the gravity field, taken from Rudolph et al. (2015). Highlighted in red and purple is the viscosity structure preferred by and used in this work.
1.3 Slab Morphologies of the Mid-Mantle

Slab diversity is particularly rich at depths between 410 and 1000 km (Fukao et al., 2001, 2009; Fukao and Obayashi, 2013). This depth range defines the extended transition from the upper to the lower mantle (Bullen, 1963) and describes a highly complex and variable zone within the mantle (Stixrude and Lithgow-Bertelloni, 2011). These depths also record two major seismic discontinuities at 410 km depth and at 660 km depth (Ringwood, 1975). Mineralogically, these depths represent the phase transition of olivine to wadsleyite and the transition of ringwoodite to bridgmanite and ferropericlase. The former exothermic phase transition is associated with a positive Clapeyron slope that contributes to the subduction of the slab. The endothermic phase transition is, however, associated with a negative Clapeyron slope. The temperature dependence of the Clapeyron slope can delay the phase transition of ringwoodite to bridgmanite and ferropericlase. In fast subducting slabs the kinetics of the phase transition can also encourage the formation of a meta-stable olivine wedge in the slab core, contributing to the positive buoyancy anomaly within the slab relative to the surrounding mantle (Kirby et al., 1991; Rubie and Ross, 1994).

Seismic tomography models indicate that some slabs like Sumatra and Cocos are able to navigate the mantle transition zone without deflection and, thus, to penetrate directly into the lower mantle beneath it (see Fig. 1.1). These slabs tend to have steep dip angles at transition zone depths and modest trench retreats at the surface (Čížková and Bina, 2019; Goes et al., 2017; Agrusta et al., 2017; Torii and Yoshioka, 2007; Čížková et al., 2002; Christensen, 1996). On the other side of the spectrum are flat, deflected slabs (c.f. Fig. 1.1 and Billen, 2008; King et al., 2015; Goes et al., 2017; Mao and Zhong, 2018). Slabs such as Izu-Bonin and Calabria tend to pond at around 660 km to 1000 km depth and represent approximately half the subducting slabs observed in seismic tomography (Fukao and Obayashi, 2013; Goes et al., 2017; Mao and Zhong, 2018).

Work by Christensen (1996) and Zhong and Gurnis (1997) suggest that a slab’s ability to penetrate through the phase transition at 660 km depth depends on its
buoyancy and the mobility of its trench at the surface. Fast, retreating trenches encourage flattening, while fast convergence discourages it. Older slabs also tend to penetrate into the lower mantle thanks to their negative buoyancy. However an old, fast sinking slab can still deflect if it exhibits fast enough trench mobility. Zhong and Gurnis (1997) propose that trench mobility is determined by the size of the overriding plate, with smaller plates having more mobile trenches. On the other hand, Yang et al. (2016) suggest that trench retreat is directly influenced by the slab dynamics at depth. The authors propose that, as slabs deflect at 660 km depth, trench retreat increases, however once slabs become gravitationally unstable and avalanche into the lower mantle, trench retreat decreases and may even switch to trench advance. It is therefore unclear whether slab flattening contributes to trench rollback or vice versa.

Perhaps a more fruitful relationship to explore is that between the slab strength and slab behaviour. Work by Stegman et al. (2006); Čížková et al. (2007); Čížková and Bina (2019) indicates that weak slabs stagnate, regardless of the trench retreat. On the other hand, the behaviour of stronger slabs with viscosities higher than $10^{23}$ Pas will depend on the rate of trench rollback. Recent numerical studies by Garel et al. (2014) and Agrusta et al. (2017) also link slab behaviour in the transition zone with the slab’s age at the trench. However, the latter authors suggest that hot, young, slabs penetrate into the lower mantle far more easily than older, stiffer slabs. Both studies exhibit similar slab behaviour at the base of the upper mantle, where the young slab folds upon itself inside the transition zone for a period of time until its combined negative buoyancy allows it to sink into the lower mantle. Garel et al. (2014) finds that the older the down-going slab is, the more it weakens the mantle region around it, lending to the system a sort of self-lubrication which induces faster sinking, subducting and trench retreat rates. The same authors also propose that the age of the overriding plate has a direct impact on the style of subduction. For older slabs, an older overriding plate gives rise to a bent, inclined and retreating (BIR) morphology. On the other hand, combining an older overriding plate with a young subducting slab or a young overriding plate and a young slab produces in both cases, vertical folding.
Agrusta et al. (2017) note that for extremely negative Clapeyron slope values ($-3 \times 10^6 \text{ PaK}^{-1}$), all slabs, regardless of their various properties, deflect. More positive Clapeyron values promote the penetration slabs of all ages and intermediate values allow young slabs to penetrate but trap older slabs. Agrusta et al. (2017) also suggests that the slab sinking mode can be switched by altering the upper plate mobility, the Clapeyron slope or the age of the subducting slab. The authors observe that for these numerical simulations shifts from penetration to stagnation occur far more easily than the shift from stagnation to penetration (Agrusta et al., 2017).

1.3.1 Mid-Mantle Slab Remnants

Evidence from seismic tomography indicates that not all subducting slabs remain whole, or even attached to the subducting lithospheric plate. Fragmented or remnant slabs are commonly found scattered at various depths throughout the mantle (Solomon and Butler, 1974; Van Der Voo et al., 1999; Schellart and Rawlinson, 2013). Seismically, these remnant slabs are observed as isolated fast anomalies surrounded by slow mantle material. These anomalously fast blobs have been detected at various depths, from shallow depths (e.g. the Mediterranean slabs, Wortel, 2000; Faccenna et al., 2014) to deeper depths (e.g. the Farallon slab, Sigloch et al., 2008).

Seismic tomography indicates that long lived subduction zones, such as the Tethyan or the Farallon systems, are particularly rich in remnant slabs (see Fig. 1.4). These intriguing subduction zones exhibit a present day actively subducting slab overlying enigmatically oriented fragments at depths below 660 km and more. Previous work has suggested that these fragments are the product of slab break-off close to the trench and represent the cessation of a subduction regime. Fragmented slabs have therefore often been seen as markers of tectonic regime shifts, with the fragment and present day subducting slab that overlies it belonging to different tectonic regimes (e.g. Sigloch et al., 2008; Wortel, 2000; van der Meer et al., 2010, 2012, 2017). This interpretation, however, involves several non-trivial assumptions, including the assumption that trenches are stationary in time and space. If one assumes a stationary trench, and vertical subduction without any deflection of the slab, reconstructing the paleo-trench is as simple as shifting the slab fragment vertically.
1.3. Slab Morphologies of the Mid-Mantle

upwards until it impinges on the surface. This point is then adopted as the location of the paleo-trench. The paleo-trench is used a constraint for the surface paleo-plate motions (e.g. Sigloch et al., 2008; van der Meer et al., 2010, 2012, 2017). The same assumptions and approach have been used to date the ages of slabs in the mantle, the initiation, duration and cessation of paleo-subductions and to infer the upper mantle viscosity (e.g. van der Meer et al., 2010, 2012, 2017). Nonetheless, evidence from both numerical models and geology indicates that these assumptions might not always hold true. In particular, trenches are known to be highly mobile, often retreating and in some rare cases even advancing (e.g. Jarrard, 1986; Bellahsen et al., 2005; Lallemand et al., 2005; Di Giuseppe et al., 2008; Torii and Yoshioka, 2007; Di Giuseppe et al., 2009; Holt et al., 2015). Evidence from seismicity at subduction zones and seismic tomography images also indicates that slabs tend to bend and deflect as they sink downwards towards the CMB (e.g. Houseman and Gubbins, 1997; Goes et al., 2004; Green, 2007). Indeed cases where the subducting slab directly and vertically underlies its trench are extremely rare on Earth. Furthermore, the combination of vertically subducting slabs and stationary trenches often requires complex surface kinematics and reorganisation, multiple trench jumps and several subduction polarity reversals to fit in with the observed geological evidence (e.g. Sigloch et al., 2008; Shephard et al., 2017).

However, new results presented in this study indicate that slab break-off does not necessarily always entail a change in the tectonic regime and the cessation of subduction. If, instead of breaking at shallow depths (> 300 km) (e.g. Duretz et al., 2011, 2012, 2014), the slab breaks, or rather orphans, at mid-mantle depths, subduction can continue uninterrupted through its parent slab. Orphaning does not require stationary trenches, vertically sinking slabs or any special assumptions. Slab orphaning is one possible evolutionary pathway that can be adopted by the slab as it reaches mid-mantle depths. Similar to slab flattening and penetration, orphaning is viable for a wide range of slab and subduction variables. Results show that orphaning occurs in a hitherto unexplored parameter space between flattening and deflection and is one possible way a slab can switch from a penetrative mode to a deflecting one.
1.3. Slab Morphologies of the Mid-Mantle

Figure 1.4: Tomographic cross section of the Tonga (a) and Japan (b) orphan slabs obtained from the submachine portal of (Hosseini et al., 2018), showing a flattened parent slab above 660 (magenta arrow) and a vertical abandoned orphan (magenta circle)
1.4  Outline

1.4.1  Main aims

Despite the central role of subducted slabs in plate tectonics, there are still a number of unanswered fundamental questions regarding subduction dynamics and the evolution of deep slabs in the mantle. In the work presented here, an attempt is made at addressing some of this paucity through 2-D numerical simulations of subduction, using the finite difference/volume code StagYY (Tackley, 2008).

In particular this work focuses on these 5 main aims:

1. Explore the slab-mantle relationship and how this varies in response to changes in the local strength of the former \( \textit{vis-a-vis} \) that of the latter

2. Provide an alternative, sound interpretation for the enigmatic deep slab fragments/orphans observed in seismic tomography models

3. Provide a thorough understanding of slab orphaning, the mechanism behind it, the conditions under which it occurs and its role in linking penetrative and deflecting slab morphologies

4. Explore the effects of a continental overriding plate on subduction evolution and deep slab dynamics

5. Apply the insights from numerical simulations to long-lived subduction zones on Earth, e.g. the Farallon, Tonga and Tethyan subductions
1.4.2 Thesis structure

Chapter 2: Methodology

This chapter provides an in-depth overview of the numerical modelling approach taken in this study. It provides details on the mantle convection code used, the physical and numerical aspects of the models and describes the initial and boundary conditions. A description of the diagnostics, analyses, post-processing and visualisations of the results is also included in this chapter.

Chapter 3: The Effect of a Continental Overriding Plate on subduction

Chapter 3 highlights the role and contributions of a continental overriding plate to the evolution of deep slab morphologies. Results show that while the presence of a continent does not change, the overall slab dynamics in the upper mantle, its influence has a drastic first order effect on deep slab behaviour.

Chapter 4: Slab Orphaning as a Mechanism for Mid-Mantle Slab Break-Off

Chapter 4 describes the novel slab orphaning morphology, where slabs break directly at mid-mantle depth. This chapter details the mechanism through which slabs orphan and illustrates how slab break-off is not necessarily always linked to subduction cessation and tectonic regime shifts. The implications of orphaning on slab histories and plate reconstructions are discussed together with examples of orphaning in nature.

Chapter 5: Orphaning Regimes; a Link Between Slab Deflection and Penetration

This chapter explores the conditions and parameter space that define slab orphaning. Orphaning is possible for a wide range of subduction and slab parameters. It occupies an intermediary parameter space between penetrative and deflecting morphologies and describes one possible pathway for slabs to switch from a penetrative mode to a deflected one.

Chapter 6: Conclusions

This chapter wraps up the main conclusions of this study and suggests directions for future work.
Chapter 2

Methodology

2.1 Governing Equations

Over geological time scales, the Earth’s mantle behaves as a highly viscous fluid. This is to say that the mantle material flows in response to an applied stress. Flow occurs very slowly due to the mantle’s high Prandtl number. This is a non dimensional ratio of the momentum diffusivity and the thermal diffusivity, defined as \( \frac{c_p \mu}{k} \), where \( c_p \) is the specific heat, \( \mu \) the dynamic viscosity and \( k \) is the thermal conductivity. In fluids, stresses are related to rates of strain and, in turn, strain rates are a result of spatial gradients in the velocities of that same fluid (Turcotte and Schubert, 2014). By association, stresses in fluids are also related to the velocity gradients of the fluids therefore, by equating the stresses to the velocity gradients, one can obtain the rheological law for that fluid.

The fluid behaviour of the Earth’s mantle determines plate tectonics, continental drift and shapes the thermal structure of the Earth. The latter is best described by a cool boundary layer at the surface and a hot interior, heated both internally through radioactive decay and through heat flux from the core-mantle boundary (CMB) at the bottom. The thermal and density contrasts, induced by the presence of hot material at the bottom and the cold material at the surface, result in opposing buoyancy forces. These are the main drivers of mantle convection.
2.1. Governing Equations

Relating the laws of conservation of mass (2.1), momentum (2.2) and energy (2.3) together with the appropriate rheological law will describe how the mantle material will flow under an applied force.

For mass to be conserved in an incompressible fluid, under the Boussinesq approximation, viscous flow in two dimensions with components $u$ and $v$ in the $x$ and $y$ direction respectively must be equal to zero;

$$\nabla \cdot \vec{u} = 0$$

The conservation of momentum describes the balance of forces acting on a volume of fluid, with the buoyancy forces driving the system counteracted by the viscous stress. Since the Earth’s mantle is highly viscous due to its high Prandtl number, inertia forces (i.e. the resistance to a change in motion) associated with acceleration can be ignored. Newton’s second law can therefore be written as;

$$-\nabla P + \nabla \cdot [\eta (\nabla \vec{u} + \nabla^T \vec{u})] + \rho g \vec{e}_z = 0$$

Where $u$, $P$, $\eta$, $\rho$, $T$, $g$ and $\vec{e}_z$ are the velocity, pressure, viscosity, density, temperature, gravity and the unit vector in the vertical direction respectively.

Both density and rheology are highly dependent on the ambient temperature, and, thus, on the evolution of the temperature field. This can be obtained by solving the energy equation below;

$$\rho c_p \frac{\delta T}{\delta t} = k \nabla^2 T + \rho H - \rho c_p \vec{u} \cdot \nabla T$$

Where $c_p$, $k$, $H$ and $t$ are the specific heat capacity, the thermal conductivity, the internal heating production rate and time respectively (Tackley, 1994; Turcotte and Schubert, 2014).

Equations 2.1, 2.2 and 2.3 can be non-dimensionalised using the depth of the mantle $D$, the conductive time $D^2/\kappa$ (where $\kappa$ is the thermal diffusivity), the viscosity $\eta_0$ and the super adiabatic temperature drop $\delta T$ (Tackley, 1994). The non-dimensional equations are;
2.1. Governing Equations

\[ \vec{\nabla} \cdot \vec{v} = 0 \]  
(2.4)

\[ \vec{\nabla} \cdot \sigma_{ij} - \vec{\nabla} p = RaT \hat{e}_z \]  
(2.5)

\[ \frac{\partial T}{\partial t} = \nabla^2 T - \vec{v} \cdot \vec{\nabla} T + H \]  
(2.6)

Where the dimensionless Rayleigh number \( Ra \) and the internal heating rate \( H \) are given by:

\[ Ra = \frac{\rho g \alpha \Delta T D^3}{\eta_0 \kappa} \]  
(2.7)

\[ H = \frac{D^2 q}{c_p \kappa \delta T} \]  
(2.8)

Given adequate boundary and initial conditions, it is possible to solve the three governing equations (equations 2.4, 2.5 and 2.6) for pressure, velocity and temperature numerically. The finite volume/difference code StagYY (Tackley, 2008) is used here to this extent, details of which can be found in the following section.
2.2 General Description of the Code

StagYY uses a finite volume/finite difference approximation for the spatial discretization of the non-dimensionalised governing equations (equations 2.4, 2.5 and 2.6) discussed in section 2.1 above. StagYY uses a staggered grid with the temperature and pressure defined at the centres of each cell volume and the velocity components defined at the centres of the corresponding cell face. This ensures that the different unknowns to be solved by the code are not defined at the same locations and avoids artificial oscillatory solutions of the pressure field (Tackley, 2008; Ismail-Zadeh and Tackley, 2010). The solution for the pressure and the velocity components is calculated with an iterative multigrid solver, using Jacobi iterations for relaxation. During each multigrid cycle, the residues of the velocity and the pressure iterations are interpolated and relaxed on a number of sub-grids with increasing cell sizes. This simultaneously smooths out, corrects the errors and calculates the residues at all wavelengths, reducing the time required for the solution to converge (Tackley, 2008). StagYY is parallelised using the Message Passing Interface (MPI) so that simulations can be run on multiple cores using High Performance Computers (HPC). In this case, the UK national HPC facility ARCHER has been used.

It is possible to discretize various geometries in StagYY, including 3-D Cartesian, a spherical annulus and a spherical shell (Hernlund and Tackley, 2008). However, for this study, a 2-D Cartesian geometry with a sticky air layer (refer to section 2.6), self-consistent (i.e. not kinematically driven), single-sided subduction and trench motion through time (Crameri et al., 2012b; Crameri and Tackley, 2015; Crameri and Lithgow-Bertelloni, 2018) has been implemented. Non-diffusive tracer particles are used to track the different materials which, in this case, include the sticky air, weak crust, mantle and continent when the latter is implemented. Details of the specific model implementation for this study, including descriptions of the physical and numerical model and an overview of the implemented rheology together with the initial and boundary conditions, can be found in the sections that follow.
2.3 Physical Model

Similar to Crameri and Lithgow-Bertelloni (2018), this study models an incompressible mantle by solving the non-dimensional equations for the conservation of mass (eq. 2.4), momentum (eq. 2.5) and energy (eq. 2.6) under the Boussinesq approximation. The Rayleigh number used in this study is $10^7$ and the non-dimensional internal heating rate is 20.0 (equivalent to $4.92 \times 10^{-12}$ W kg$^{-1}$).

The models include a weak hydrated crustal layer on top of the subducting plate. This develops on the surface of the young downgoing lithospheric plate, away from the ridge at $\sim$10 Myrs, and is allowed to form a freely evolving weak subduction channel once the plate subducts (Crameri and Tackley, 2015). The weak crustal layer only differs from the mantle material by its lower yield strength and has a thickness of 15 km. It is converted to regular mantle once it is subducted below a depth $d > 400$ km. A melting viscosity reduction (MVR) is also implemented to simulate a low-viscosity asthenosphere underneath the overriding and subducting plates. Within the MVR, the viscosity is reduced by a factor of 10 for regions where the temperature exceeds a depth-dependent solidus (Tackley, 2000b). It has been shown in the past that the presence of an MVR aids in maintaining single sided subduction and plate-like behaviour (see e.g., Tackley, 2000b; Crameri and Tackley, 2015, 2016).
2.4 Rheology

The mantle’s rheology describes how rocks flow or deform due to an imposed stress at specific temperature and pressure conditions. Deformation can be either recoverable or permanent. Recoverable deformation describes the return of zero strain once the stress is removed, examples of which include elastic deformation and thermal expansion. On the other hand, examples of permanent deformation include plastic, viscous flow and creep. Permanent deformation is described by a time-dependent strain-stress relationship, which means it is irrevocable and cannot be undone once a material’s yield stress has been reached. The behaviour of rocks under applied loads is empirically derived from laboratory experiments (normally of dry olivine, this being the most abundant mineral in the upper mantle (Turcotte and Schubert, 2014)) and compared to theoretical constitutive equations or rheological laws.

With increasing depth, pressures and temperatures, deformation evolves from elastic to brittle to plastic and creep behaviours. The type of creep behaviour is determined by the amount of stress. At low stress levels, a solid predominantly deforms through diffusion creep. This mechanism describes the diffusion of atoms through the crystal lattice and boundaries. Diffusion creep describes a linear relationship between strain and stress (see equation diffusion 2.12) and results in a Newtonian fluid behaviour (Turcotte and Schubert, 2014). At higher stress levels, deformation predominantly occurs through dislocation creep, i.e. the migration of imperfections in the crystalline lattice structure (Turcotte and Schubert, 2014). It is important to note that dislocations are not systematically proportional to time, and dislocation creep results in non-Newtonian deformation where the strain rate is proportional to a power $n$ of the stress, with $n$ being $\approx$ equal 3. Therefore, unlike diffusion creep, dislocation creep obeys a power-law rheology. Both diffusion and dislocation creep are directly proportional to the exponential of the inverse absolute temperature. Thus, the viscosity of the mantle has a strong temperature dependence. Furthermore, the effective viscosity of dislocation creep is inversely proportional to the square of the stress (see equation dislocation 2.13), which implies that the mantle viscosity is also stress dependent.
Following from the above, the models described in this study are implemented with a visco-plastic rheology that strongly depends on temperature and pressure. Visco-plastic rheology allows the mantle material in the model to yield plastically once its yield stress value is reached. Yielding is included in the models via a yield stress limiter using a Drucker-Prager yield criterion, with the pressure-dependent yield stress, $\sigma_{y,\text{brittle}}$, based on Byerlee’s law. In other words, yielding is both pressure-dependent and rock type independent;

$$\sigma_{y,\text{brittle}} = C + p\mu$$  \hspace{1cm} (2.9)

with specified friction coefficient $\mu$ and cohesion $C$. A constant, ductile yield stress limits the effective stress and is given by the equation:

$$\sigma_{y,\text{ductile}} = \sigma_{y,\text{const}}$$  \hspace{1cm} (2.10)

where $\sigma_{y,\text{const}}$ is the surface value of the ductile yield stress. The effective yield stress is then given by:

$$\sigma_y = \min \left[ \sigma_{y,\text{brittle}}, \sigma_{y,\text{ductile}} \right]$$  \hspace{1cm} (2.11)

The cross-over depth between brittle and ductile yielding is at approximately half depth of the lithosphere which, for the models presented in this study occurs at 72 km depth. Specific values of diffusion and dislocation-creep parameters together with other physical parameter details are presented in table 2.2.

The implemented viscosity is a combination of two constitutive equations, the Arrhenius law (equation 2.12) describing the linear, temperature dependent diffusion creep and a power law creep which describes the stress dependence of dislocation creep (equation 2.13).

$$\eta(T, p) = \eta_A \cdot \exp \left[ \frac{E_{\text{act}} + pV_{\text{act}}}{RT} \right]$$  \hspace{1cm} (2.12)

where $\eta$ is the viscosity, $p$ is the pressure, $R$ is the gas constant, $T$ the temperature, $E_{\text{act}}$ the activation energy, $V_{\text{act}}$ the activation volume and $\eta_A$ is set such that
2.4. Rheology

\( \eta \) gives the reference viscosity \( \eta_0 \) at \( T = 1600 \) K and \( p = 0 \) Pa.

\[
\eta_{\text{dis}} = \frac{\sigma^{(1-n_{\eta_{\text{dis}}})}}{\sigma_{TS}}
\]  

(2.13)

where \( \sigma \) is the stress at the centre of the cell and \( \sigma_{TS} \) is the stress value at which dislocation creep is activated.

The effective viscosity at a given grid point is given by equation 2.14. Due to its pressure and temperature dependence the effective viscosity of the mantle also varies with depth (see Fig. 2.1 and table 2.1)

\[
\eta_{\text{eff}} = \min[\eta(T,p), \eta_{y}]
\]

(2.14)

Where \( \eta_{y} \) is the plastic yielding with \( \eta_{y} = \sigma_y/(2\dot{\varepsilon}) \) and \( \dot{\varepsilon} \) is the strain rate. \( \eta(T,p) \) is the combination of the diffusion creep \( \eta_{\text{diff}} \) and the dislocation creep \( \eta_{\text{dis}} \) at specific pressures and temperatures, such that \( \eta_{T,p} = \left( \frac{1}{\eta_{\text{diff}}} + \frac{1}{\eta_{\text{dis}}} \right)^{-1} \). The viscosity variation is limited to 9 orders of magnitude by applying an upper and lower cut-off of \( \eta_{\text{max}} = 10^5 \eta_0 \) and \( \eta_{\text{min}} = 10^{-4} \eta_0 \).

Four main variants of viscosity variations with depth are implemented in this study: V1 describes the traditional and until now canonical mantle viscosity profile, where the viscosity increase is concurrent with the endothermic phase change at 660 km (Hager and Richards, 1984); Profile V2, in which the phase change effects at 660 km depth are ignored and the lower mantle viscosity increase is shifted to 1000 km depth. In V2 the viscosity increase at depth is preceded by a low viscosity layer (LVL) between 660 and 1000 km depth; Profile V3, describes the phase change of ringwoodite to bridgmanite and ferropericlase at 660 km depth and a separate viscosity jump at 1000 km depth. In this case the viscosity increase at depth is separated from the phase change by an LVL that extends from 660 to 1000 km as described in Rudolph et al. (2015); Lastly, profile V4 which is identical to V3, except that the LVL between 660-1000 km depth is missing and the upper mantle viscosity extends to 1000 km, at which depth it then increases by an order of magnitude.
Also included are the effects of the ringwoodite to bridgmanite and ferropericlase endothermic phase transition on the flow, with Clapeyron slope and density jump values specified in table 2.2. This is included in all set-ups except V2.

<table>
<thead>
<tr>
<th>Model</th>
<th>LVL</th>
<th>Clapeyron Slope (Pak⁻¹)</th>
<th>Viscosity Jump Depth (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>V1</td>
<td></td>
<td>−2.5 × 10⁶</td>
<td>660</td>
</tr>
<tr>
<td>V2</td>
<td>✓</td>
<td>0</td>
<td>1000</td>
</tr>
<tr>
<td>V3</td>
<td>✓</td>
<td>−2.5 × 10⁶</td>
<td>1000</td>
</tr>
<tr>
<td>V4</td>
<td></td>
<td>−2.5 × 10⁶</td>
<td>1000</td>
</tr>
</tbody>
</table>
2.4. Rheology

Figure 2.1: Schematic diagram showing the model set-up for V1 following from Crameri and Lithgow-Bertelloni (2018). The orange dotted line represents the imposed phase change at 660 km depth. This is complimentary to the increase in viscosity at the same depth. Darker shades of grey represent increases in viscosity. Vertical viscosity variations with depth are shown in the profile on the left.

Figure 2.2: Schematic diagram showing the model set-up used for V2 following from Crameri and Lithgow-Bertelloni (2018). A low viscosity layer is implemented between 660 km and 1000 km similar to Rudolph et al. (2015). Darker shades of grey represent increases in viscosity. The profile on the left shows the vertical viscosity variations of the model with depth.
Figure 2.3: Schematic diagram showing the model set-up used for V3 following from Crameri and Lithgow-Bertelloni (2018) with a low viscosity layer between 660 km and 1000 km. The orange dotted line represents the imposed phase change at 660 km depth, while vertical viscosity variations with depth can be observed in the profile on the left. Darker shades of grey in the model schematic on the right represent higher viscosities.

Figure 2.4: Schematic diagram showing the model set-up used for V4 following from Crameri and Lithgow-Bertelloni (2018) with a viscosity jump at 1000 km. The orange dotted line represents the imposed phase change at 660 km depth, while vertical viscosity variations with depth can be observed in the profile on the left. Darker shades of grey in the model schematic on the right represent higher viscosities.
### 2.4. Rheology

Table 2.2: Model Parameters

*a*Indicates suitable free-surface approximation with sticky-air approach if $C_{Stokes} \ll 1$ (Crameri et al., 2012a).

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Non-dimensional value</th>
<th>Dimensional value</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reference viscosity</td>
<td>$\eta_0$</td>
<td>1</td>
<td>$10^{22}$</td>
<td>Pa s</td>
</tr>
<tr>
<td>Mantle depth</td>
<td>$D$</td>
<td>1</td>
<td>2890</td>
<td>km</td>
</tr>
<tr>
<td>Upper-mantle depth</td>
<td>$D_{UM}$</td>
<td>0.23</td>
<td>660</td>
<td>km</td>
</tr>
<tr>
<td>Gravitational acceleration</td>
<td>$g$</td>
<td>-</td>
<td>9.81</td>
<td>ms$^{-2}$</td>
</tr>
<tr>
<td>Thermal conductivity</td>
<td>$k$</td>
<td>-</td>
<td>3</td>
<td>Wm$^{-1}$K$^{-1}$</td>
</tr>
<tr>
<td>Thermal diffusivity</td>
<td>$\kappa$</td>
<td>1</td>
<td>$10^{-6}$</td>
<td>m$^2$s$^{-1}$</td>
</tr>
<tr>
<td>Thermal expansivity</td>
<td>$\alpha$</td>
<td>-</td>
<td>$3 \times 10^{-5}$</td>
<td>K$^{-1}$</td>
</tr>
<tr>
<td>Temperature gradient</td>
<td>$\Delta T$</td>
<td>1</td>
<td>2500</td>
<td>K</td>
</tr>
<tr>
<td>Reference density</td>
<td>$\rho_0$</td>
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<td>3300</td>
<td>kg m$^{-3}$</td>
</tr>
<tr>
<td>Heat capacity</td>
<td>$Cp_0$</td>
<td>-</td>
<td>1200</td>
<td>Jkg$^{-1}$K$^{-1}$</td>
</tr>
<tr>
<td>Internal heating rate</td>
<td>$H$</td>
<td>20</td>
<td>$4.92 \times 10^{-12}$</td>
<td>Wkg$^{-1}$</td>
</tr>
<tr>
<td>Gas Constant</td>
<td>$R$</td>
<td>-</td>
<td>8.314</td>
<td>Jmol$^{-1}$K$^{-1}$</td>
</tr>
<tr>
<td>Clapeyron slope at 660 km</td>
<td>$\gamma_{660}$</td>
<td>-</td>
<td>$-2.5 \times 10^6$</td>
<td>Pa K$^{-1}$</td>
</tr>
<tr>
<td>Density jump at 660 km</td>
<td>$\rho_{660}$</td>
<td>-</td>
<td>341</td>
<td>kg m$^{-3}$</td>
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<td><strong>PLASTICITY</strong></td>
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<td>Friction coefficient</td>
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<tr>
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<td>1577</td>
<td>$10 \times 10^6$</td>
<td>Pa</td>
</tr>
<tr>
<td>Max. yield stress</td>
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<td>$9.5 \times 10^4$</td>
<td>$600 \times 10^6$</td>
<td>Pa</td>
</tr>
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<td><strong>DIFFUSION CREEP</strong></td>
<td></td>
<td></td>
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<td><strong>DISLOCATION CREEP</strong></td>
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<td>Activation energy</td>
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<td>kJmol$^{-1}$</td>
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<tr>
<td>Activation volume</td>
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<td>$1 \times 10^{-6}$</td>
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<td>Powerlaw index</td>
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<td><strong>STICKY-AIR LAYER</strong></td>
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<td>Thickness</td>
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<td>150</td>
<td>km</td>
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<tr>
<td>Viscosity</td>
<td>$\eta_{st}$</td>
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<td>$10^{19}$</td>
<td>Pa s</td>
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<td>C-condition$^a$</td>
<td>$C_{Stokes}$</td>
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<td><strong>WEAK CRUSTAL LAYER</strong></td>
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<td>km</td>
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<tr>
<td>Viscosity</td>
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<td>$\eta_0$</td>
<td>$10^{22}$</td>
<td>Pa s</td>
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<tr>
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<td></td>
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<tr>
<td>Cohesion</td>
<td>$C_{\text{crust}}$</td>
<td>1577</td>
<td>$10 \times 10^6$</td>
<td>Pa</td>
</tr>
</tbody>
</table>
2.5 Numerical Model

The model domain consists of the whole mantle depth together with an added 150 km thick sticky layer on top (Schmeling et al., 2008; Crameri et al., 2012a). The 2-D Cartesian box has an aspect ratio of 2:1 ($x$:$z$) and the model is discretized to $512 \times 256$ grid points. The horizontal grid spacing is constant and the resolution in the box is approximately 11 km. The vertical grid spacing is refined towards the rock-air interface and at 660 km, yielding a maximum vertical resolution of 4 km. The high resolution at the surface is required to resolve the thin subduction channel described in section 2.3 and to guarantee lubrication between the converging plates (Crameri and Tackley, 2015). On average, 100 Lagrangian tracers per grid cell are used to track the composition (i.e., mantle, weak crust, continent and air) and physical interfaces (e.g., the free surface) in sub-grid resolution. These simulations have the required resolution to capture the dynamics presented here as they have been extensively tested in previous studies (see e.g., Crameri and Tackley, 2015; Crameri and Lithgow-Bertelloni, 2018). Moreover, the resolution of these models has been further verified through the rigorous testing discussed in Appendix A section A.2, and indicate that there are no under-resolved numerical issues or any geometry effects stemming from either the initial conditions or the model set-up.

The ringwoodite to bridgmanite and ferropericlase phase transition is implemented using the phase buoyancy parameter formulation, as in Tackley (1994, 1996). The buoyancy parameter describes a ratio of buoyancy due to phase change deflection to the buoyancy due to thermal effects, integrated over a column and is given by:

$$P = \frac{\gamma \delta \rho}{g \alpha \rho h}$$  \hspace{1cm} (2.15)
where $\gamma$ is the Clapeyron slope, $g$ is the gravitational acceleration, $\alpha$ is the thermal expansivity, $\delta\rho$ is the density jump and $h$ is the depth of the modelled mantle. Since the deflection due to the phase change and the thermal buoyancy are both proportional to the temperature scale, the latter is not significant. We use a sheet mass anomaly to represent the phase transition (Tackley, 1996; Xie and Tackley, 2004), which allows for an infinitely sharp (discontinuous) change, rather than one with a finite width (Christensen and Yuen, 1985).

### 2.6 Boundary and Initial Conditions

The top and bottom boundaries are free slip. The side boundaries are periodic. This is to minimise the boundary effects of the induced mantle flow on subduction, thus preventing any unnatural forcing during the model’s longer time evolution (Chertova et al., 2012). The top boundary is set to a constant 300 K, while the bottom boundary is insulating with a zero heat flux condition. A sticky air layer is implemented at surface. This essentially extends the free-slip condition vertically upwards by introducing a low density, low viscosity layer (Crameri et al., 2012a) (see table 2.2). The sticky air layer acts as free-surface proxy, and provided it is sufficiently thick, low density and low viscosity, decouples the surface of the lithosphere from the top of the model domain. The sticky air approach allows vertical movement of the physical surface without exerting stresses on it, and any deflection of the surface is able to relax isostatically without ‘feeling’ the presence of the sticky air layer on top of it (Crameri et al., 2012a). This implies that the air/crust interface is virtually free from shear and normal stresses, enabling the formation of realistic topography. The condition for a traction free surface in the isostatic limit on timescales larger than the isostatic relaxation timescale is defined by $C_{Stokes} << 1$, where $C_{Stokes} = \frac{1}{16} \frac{\delta\rho}{\rho} \left( \frac{h_{model}}{h_{st}} \right)^3 \frac{\eta_{ch}}{\eta_{st}}$, with $h_{model}$, $h_{st}$, $\eta_{st}$ and $\eta_{ch}$ describing the height of the model, the height and viscosity of the sticky air layer and the viscosity controlling the relaxation respectively (Turcotte and Schubert, 2014). The location of the free surface is determined by the spatial resolution of the sticky-air tracers and is directly measured through the Lagrangian tracking of topography on a Euclidean...
2.6. Boundary and Initial Conditions

This method allows for great surface deformation without grid distortion (Crameri et al., 2012a).

Since subduction initiation is not the focus of this work, the initial condition is one of ongoing stable single-sided subduction. The down-going straight slab is 400 km long (see Fig. 2.3) and has an initial constant thickness corresponding to the surface plate thickness at the trench. An initial slab dip angle of $30^\circ$ is defined at the shallow depth range of around 150 – 250 km and is allowed to evolve freely. Subsequently, slab tip angles are measured at a depth of 175 km to account for the effects of mantle phase- and/or viscosity transitions.

The edge of the box is assumed to be a divergent plate boundary. This is implemented by introducing an initial temperature field that follows a standard $\sqrt{\text{age}}$-law. As the lithosphere moves away from the ridge and towards the trench it cools and thickens. Thus, the introduction of an initial temperature field reproduces a thin, light and elevated plate close to the ridge, and a thicker, denser plate further away from the ridge. The buoyancy contrast across the plate supports the initial movement of the lithosphere away from the ridge and is defined by the initial plate thickness $w_{BL}(x) = w_{BL,0} \cdot \sqrt{\Delta x_{sc}}$, where $w_{BL,0}$ is a constant controlling the maximum thickness of the plate, $x$ is the horizontal coordinate and $\Delta x_{sc}$ is the distance from the spreading centre at any given position $x$. The initial temperature is related to plate age as $T_z(x) = T_0 \cdot \text{erf}[(1 - z)/w_{BL}(x)]$, with $T_0 = 0.64$ the initial non-dimensional mantle temperature and $z$ as the vertical coordinate ranging between 0 at the bottom to 1 at the top boundary. The plate’s non-dimensional initial thickness at the trench is chosen to be $\sim 0.04D$ based on the observation that this is a typical boundary layer thickness with the chosen $Ra = 10^7$ and $H = 20$ (Crameri and Tackley, 2015). Defining the initial plate thickness in this way introduces a divergent plate boundary that evolves dynamically and self-consistently during the model evolution, without any kinematically imposed forcing.
2.7 Continental Overriding Plate Implementation

A continental overriding plate that is 200 km deep and 2000 km wide, similar to Crameri and Lithgow-Bertelloni (2018), is also implemented for some cases. The depth of the continent is such that its roots extend to the bottom of the asthenosphere into the high viscosity upper mantle. The continental geometry is rectangular with an angled edge at the trench.

While the continental material is chemically and compositionally indistinct from the mantle below it, its lighter density is due to its buoyancy ratio \( B_c \). \( B_c \) is simply the density contrast of the continent, \( \Delta \rho_c \) with respect to the ambient mantle \( \Delta \rho_m \), divided by the thermal density variation \( \rho \alpha \Delta T \) such that \( B_c = \Delta \rho_c - \Delta \rho_m / \rho \alpha \Delta T \). I assume a continent density of 3250 kg m\(^{-3}\) similar to Crameri and Lithgow-Bertelloni (2018) and take the average mantle density to be 3300 kg m\(^{-3}\), this results in a continental density contrast \( \Delta \rho_c \) of -25 kg m\(^{-3}\) and an associated buoyancy ratio \( B_c \) of -0.10. The more negative the value of \( B_c \), the lighter the implemented continent. Thus, the continent is both lighter due to its lower reference density and stronger due to its high viscosity and higher friction coefficient. The continent’s high viscosity translates to high ductile strength and its higher friction coefficient means that yielding occurs for much higher stresses. The yield stress is depth-dependent and is derived through the friction coefficient. This ensures that the continental lithosphere is stronger than the mantle material at shallow and deeper depths. Further details on the model parameters can be found in table 2.3 below.

The effect of the continental overriding plate on the flow dynamics is explored in combination with the 4 different mantle viscosity profiles shown in table 2.1.

Table 2.3: Continental Overriding Plate Parameters

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Non-dimensional value</th>
<th>Dimensional value</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>Continent</td>
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<td></td>
</tr>
<tr>
<td>Thickness</td>
<td>( d_c )</td>
<td>0.071</td>
<td>200</td>
<td>km</td>
</tr>
<tr>
<td>Width</td>
<td>( W_c )</td>
<td>0.712</td>
<td>2000</td>
<td>km</td>
</tr>
<tr>
<td>Viscosity</td>
<td>( \eta_c )</td>
<td>100( \eta_0 )</td>
<td>10(^{24})</td>
<td>Pa s</td>
</tr>
<tr>
<td>Friction coefficient</td>
<td>( \mu_c )</td>
<td>0.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Density contrast</td>
<td>( \Delta \rho_c )</td>
<td>0.087</td>
<td>-25.0</td>
<td>kg m(^{-3})</td>
</tr>
</tbody>
</table>
2.8 Diagnostics and Visualisations

Model data and results are post processed and analysed using the geodynamic diagnostic, post-processing and visualisation software StagLab 3.0 (Crameri, 2018) (http://doi.org/10.5281/zenodo.1199037). Although StagLab is compatible and has been tested with other mantle convection and geodynamic codes, it has been specifically developed to handle the raw data outputs from StagYY (c.f. section 2.2).

StagLab has a series of inbuilt key diagnostics. These include: mean plate thickness, upper mantle temperature and viscosity, measurements of the regional subduction topography (e.g. trench depth, back-arc depression and island arc height), plate thickness, and plate boundary tracking, amongst others (see Fig. 2.5). StagLab also has the ability to track the trench velocity and the slab sinking rate (Crameri, 2018). The former is the velocity of the upper plate measured next to the trench, while the latter is the vertical velocity of the slab measured at ∼ 72 km depth. Through its slab-dynamics diagnostics, StagLab can track the slab-dip depth, the shallow-depth slab dip angle (measured at ∼ 175 km), the subduction polarity, the slab’s viscosity and the slab-mantle viscosity contrast (c.f. Fig. 2.5, Crameri, 2018). These measurements are particularly relevant to this study and allow for a better understanding of, and comparison across, the different models implemented. Through StagLab, one can also highlight the temporal evolution of a simulation, and enlarge and zoom in on important details in particular subplot. Furthermore, StagLab provides ‘scientific-proof’, perceptually uniform colour schemes and visualisations. Results plotted through StagLab are intuitive, easier to understand and do not distort the underlying data (Crameri, 2018).
2.8. Diagnostics and Visualisations

Figure 2.5: Schematic diagram reproduced from Crameri et al. (2017) showing some of the diagnostics and measurements that can be made at a subduction zone, using StagLab. From right to left these are; the viscous fore-bulge (FB), subduction trench (TR), island arc (IA) and back-arc depression (BAD). Other diagnostics include the shallow depth slab dip $\theta$ and the bending radius, measured by fitting a circle to the point of maximal plate bending (Crameri, 2018).
Chapter 3

Slab Orphaning as a Mechanism for Mid-Mantle Slab Break-Off

3.1 Introduction

Seismic tomography shows a plethora of slab morphologies between 410 and 1000 km depth. These depths define the transition from the upper to the lower mantle (Bullen, 1963) and describe a highly complex and variable zone within the mantle (Stixrude and Lithgow-Bertelloni, 2011). Slabs like Sumatra and Cocos seem to navigate the mantle transition zone without deflection and to penetrate directly into the lower mantle beneath it. Such slabs tend to have steep dip angles at transition zone depths and modest trench retreats at the surface (Čížková and Bina, 2019; Goes et al., 2017; Agrusta et al., 2017; Torii and Yoshioka, 2007; Čížková et al., 2002; Christensen, 1996). On the other side of the spectrum are flat, deflected slabs (Billen, 2008; King et al., 2015; Goes et al., 2017; Mao and Zhong, 2018). Slabs such as Izu-Bonin and Calabria tend to pond at around 660 km to 1000 km depth and represent approximately half the subducting slabs observed in seismic tomography (Fukao and Obayashi, 2013; Goes et al., 2017; Mao and Zhong, 2018).
3.1. Introduction

The mantle also exhibits fragmented or remnant slabs (Solomon and Butler, 1974; Van Der Voo et al., 1999; Schellart and Rawlinson, 2013). These can be observed as isolated fast seismic anomalies surrounded by slow mantle material and appear to be detached from the subducting plate at the surface. Fragmented slabs can be observed at any given depth in the mantle, from shallow examples such as the Mediterranean slabs (e.g. Wortel, 2000; Faccenna et al., 2014) to deeper lying fragments such as the Farallon, Tonga, or Japan slabs (e.g. van der Hilst, 1995; Karato et al., 2001; Sigloch et al., 2008; Schellart and Spakman, 2012; Tao et al., 2018). There are four main pathways to slab break-off. One cause of fragmentation would be the subduction of a ridge or heavily faulted slab riddled with heterogeneities (Wortel, 2000; Burkett and Billen, 2009, 2010; Faccenna et al., 2014; Crameri and Tackley, 2015). Another cause of slab fragmentation can be the arrival at the trench of buoyant lithosphere such as a continental block (Wortel, 2000). The presence of a continental keel can directly decapitate the subducting slab (Sigloch et al., 2008). The positive buoyancy of continental lithosphere can also act to plug the subduction zone, forcing the slab to tear and subduction to terminate (Crameri and Tackley, 2015). Slabs can also break as a result of a decrease in surface plate velocities (Li, 2002) leading to shearing and necking of the slab at depths of 300 km or less (Duretz et al., 2011, 2012). This study proposes that slabs can also break directly at 660 km depth through Slab Orphaning. Orphaning and slab break-off at shallow depths (through any of the first three pathways discussed above), are both the result of weakening and shear localisation. However, orphaning is intrinsically tied to the mantle transition zone, while shallow break-off is the direct result of the surface plate dynamics. Orphaning does not exclude any of the shallow break-off mechanisms. Indeed, the two can act together to shape subduction zones and slab dynamics.
Orphaning can provide a simple interpretation for some of the seismically observed remnant slabs (see e.g., Sigloch et al., 2008; Richards et al., 2011). This is particularly so for slab fragments lying at depths > 660 km. Slab orphaning allows for the presence of slab fragments in the lower mantle without invoking stationary trenches, subduction polarity reversals or trench jumps (Sigloch et al., 2008; Yoshioka et al., 2010; Sigloch, 2011; Shephard et al., 2017). In the case of orphaning, remnant slabs can have radically different orientations and be located far away from the active trench, but still be associated with the ongoing subduction. Slab orphaning challenges the assumption that remnant slabs are the by products of extinct subduction zones (Fukao et al., 2001, 2009; Simmons et al., 2012; Fukao and Obayashi, 2013; Sigloch et al., 2008; Goes et al., 2017; van der Meer et al., 2017). Unlike shallow slab break-off (see e.g., Wortel, 2000; Li, 2002; Duretz et al., 2011), orphaning suggests that subduction can continue through the parent slab, which flattens and travels horizontally at 660 km depth.
3.2 Methods

As described in Chapter 2, this study implements a series of 2-D simulations of dynamically self-consistent, single-sided subduction using the finite difference/volume code StagYY (Tackley, 2008). The rheological implementation includes linear diffusion, power-law dislocation creep and visco-plasticity with brittle and ductile yielding (see table 2.2). The slab undergoes the same viscosity reduction as the surrounding mantle between 660 and 1000 km, this weakens the slab, mimicking the effect of processes like superplasticity and grain size weakening on rheology (e.g., Christensen and Yuen, 1985; Karato et al., 1995, 2001). The radial viscosity variations of the models are based on geoid inversions (e.g., Rudolph et al., 2015). While the effects of the endothermic ringwoodite to bridgmanite and ferropericlase phase transition are implemented using a phase buoyancy parameter formulation (Tackley, 1994, 1996; Xie and Tackley, 2004), with a strongly negative Clapeyron slope of $-2.5 \times 10^6 \text{ PaK}^{-1}$ (see eq. 2.15). The initial and boundary conditions follow those stated in Chapter 2, section 2.6 and describe a 400 km long slab with an initial dip angle of $30^\circ$. The upper plate surface is free to move and plate motion is not kinematically forced, but results from a half space cooling lithospheric age profile at the top of the model (see Chapter 2 and Figs. 2.1, 2.2, 2.3). The model evolves from this initial state without any artificial kinematic forcing. A thick, weak layer mimics a weak hydrated crust and is initially imposed on the surface of the lower plate at $\sim 10$ Myr and subsequently left to evolve self-consistently. The weak layer which acts to lubricate the subduction interface, is converted to normal mantle material at 400 km depth. For an extended description of the physical and numerical model, the initial conditions and the full list of parameters, see Chapter 2 and table 2.2.
3.3 Results

Varying the viscosity of the ambient mantle with depth makes it possible to explore how the slab responds to changes in its relative strength with respect to the surrounding mantle. This study implements two different vertical viscosity profiles for the mantle. $V_1$, an until now canonical profile where the viscosity increases at 660 km depth (Hager and Richards, 1984), and $V_3$ where the viscosity increase in the lower mantle occurs at 1000 km and is overlain by a low viscosity layer (LVL) (Kido and Čadek, 1997; Kido et al., 1998; Rudolph et al., 2015).

Illustrative of these results are three end member models shown in Fig. 3.1: model $V_1$ with a viscosity increase and strong endothermic phase change at 660 km depth; model $V_2$ with an ambient viscosity structure as in Rudolph et al. (2015) in the absence of a phase transition at 660 km depth; model $V_3$ with an ambient viscosity structure as in Rudolph et al. (2015) in the presence of a phase transition at 660 km depth. With these three cases, the main slab archetypes described in Chapter 1, Fig. 1.1: deflected and flattened slabs (Fig. 3.1a-c), penetrating slabs (Fig. 3.1d-f) and orphaned slabs (Fig.3.1g-i), are reproduced. This study indicates that ambient mantle viscosity and hence the relative strength of the slab with respect to the mantle as a whole, has a first order effect on the slab morphology at depth (Fig. 3.1). This outcome is in contrast with recent work (King et al., 2015) which suggests that kinetics of the phase transitions play a more important role in slab flattening than the viscosity of the mantle into which the slab is sinking.

Subduction evolution for $V_1$ to $V_3$ is characterized by three main epochs (Fig. 3.2a-b), as in Funiciello et al. (2003). A ‘slab free-fall’ epoch in the upper mantle, and two epochs of slow down in trench and slab velocities. The latter two result from the slab tip encountering the phase change and/or the viscosity jump. In addition for $V_3$, there is a temporary stage during orphaning defined by the transient detainment of the slab tip in the LVL between 660 and 1000 km (red shaded area in Fig. 3.2c). During this stage, slab sinking and trench retreat velocities remain constant as the slab deforms (stretching and shearing). Once break-off occurs and the slab tip orphans from its parent, a jump in slab sinking velocity (Fig. 3.2c phase 4a)
coincides with the complete flattening of the parent on top of the 660 km discontinuity. Similarly to the final epoch in V1, subduction velocities also decrease for V3 (Fig. 3.2c phase 4b). The drop in velocities reflects a switch from predominantly vertical to horizontal motion of the parent on top of 660 km.

Figure 3.1: Zoomed in view of the morphological evolution of the slab for all three model set-ups (V1 – V3) showing the viscosity structure. Each panel represents a time slice. For each model I vary the vertical viscosity profile of the mantle and exclude a phase transition at 660 km depth for V2. (a-c) Model V1 with a viscosity jump and phase change at 660 km depth: the subducting slab is unable to penetrate into the lower mantle, and it deflects at 660 km depth. (d-f) Model V2 with a viscosity jump at 1000 km and an LVL above as in Rudolph et al. (2015): despite the viscosity increase beyond 1000 km, the slab manages to penetrate through the low viscosity layer and into the lower mantle. (g-i) Model V3 is similar to V2, but with the addition of a phase change at 660 km depth: the temporarily trapped slab tip between 660-1000 km depth eventually evolves into slab orphaning. Further information on sensitivity tests for orphaning can be found in the Appendix A, section A.2.
3.3. Results

Figure 3.2: Slab sinking and trench retreat velocities for each model set-up (V1-V3). The former is measured at ~ 72 km depth and the latter is the velocity of the upper plate next to the trench. Changes in slab sinking velocities in blue are correlated to trench retreat velocities in grey, by red dashed lines. Initially, velocities increase rapidly for V1-V3, representing the slab free-fall stage in epoch 1. In epoch 2, the slab reaches 660 km depth and sinking velocity is low. In epoch 3, velocities decrease further in response to the slab’s adjustment to subduction obstacles. This epoch varies greatly across models (V1-V3). (a) V1, the subducting slab is unable to penetrate into the lower mantle, instead it flattens and deflects on top of the 660 km. (b) V2, gently decreasing gradients represent the slab’s penetration into the more viscous lower mantle. (c) in V3, the slab is temporarily trapped below 660 and above 1000 km, this is reflected in the constant velocity trend highlighted by the shaded red portion of the last graph. In epoch 4a, slab orphaning results in a temporary increase in velocities, which decreases again in epoch 4b, when the parent slab flattens on top of the 660 km phase transition and eventually laterally overtakes its orphan.
3.4 Discussion

Slab orphaning is the result of a competing force balance between the up- and down-dip portions of the slab at 660 km depth in the presence of an LVL. Up dip at 660 km depth, the temperature dependence of the Clapeyron slope delays the phase transition in the slab’s cold interior, forming a lens of lower density material within the slab (Fig. 3.4a). The low density lens hinders the sinking of the slab into the lower mantle. This effect is initially offset by the viscous coupling of the slab with the surrounding mantle, which helps to drag the slab tip further down into the LVL (Fig. 3.3). The very presence of the LVL encourages higher down-dip slab sinking speeds. The discrepancy between up-dip and down-dip slab velocities leads to shearing and stretching of the slab at 660 km depth accommodated by plastic deformation (Fig. 3.4e). Plastic deformation of the slab at this depth is critically encouraged by the viscosity reduction in the LVL and produces a slab weakening effect between 660-1000 km depth. A viscosity increase at 1000 km, separate from the phase change at 660 km, is also fundamental to orphaning. The jump in viscosity at greater depths forces the coupled slab-mantle flow to switch from generally vertical to horizontal. The change in flow direction increases the strain rate the slab experiences at 660 km (see Fig. 3.5). Shear and strain localisation are naturally focused around the low density lens, whose lateral extent, and therefore its resistance to subduction, increases as subduction progresses (Fig. 3.4b-c). This forces more of the slab up-dip to deflect and flatten at 660 km, and to eventually abandon its slab portion down-dip (Figs. 3.3).
3.4. Discussion

Figure 3.3: Schematic diagrams of slab orphaning. (a) Slab penetration into the lower mantle is discouraged by the presence of a low density lens within the cold slab core, as a result of the temperature dependence of the Clapyeron slope at 660 km depth. This is countered by a net downward force resulting from the slab’s negative buoyancy and the coupling of slab with the surrounding mantle, which help to drag the slab tip into the LVL. Resistance to subduction at 660 km depth and faster subduction velocities in the LVL encourage slab internal stretching and shearing around 660 km depth. (b) A lower upper mantle (UM) slab dip angle, fostered by progressing trench retreat, increases the lateral extent of the low-density lens. The resulting increased sinking resistance overcomes the slab tip buoyancy force and, leads to an accelerated slab flattening in the upper mantle, while the slab tip continues sinking under its own weight. The differential motion between slab tip and UM slab is accommodated by increased stretching and shearing within the slab until the complete abandonment of the orphan slab.
3.4. Discussion

Figure 3.4: (V3 density plots: a-c). Each panel represents a time slice in the model evolution. (a) The temperature dependence of Clapeyron slope produces a low density lens within the slab at 660 km, resisting slab penetration into the lower mantle. (b) As more slab material accumulates at 660 km, the spatial extent of the low density lens increases. (c) The low density lens effectively hinders further penetration of the parent slab below 660 km. (V3 stress plots: d-f) Show the slab at its maximum yield stress above the LVL and at 660 km depth. (V3 deformation mechanism: g-i) Show that the slab is deforming plastically throughout. This encourages the increase in strain rates (see Fig. 3.5) as the slab sinking velocities in the LVL increase, leading to a reduction in the slab viscosity and weakening at 660 km depth.

Figure 3.5: (a-c) Strain rate for V3. Each panel represents a time slice in the model evolution. (a) Before orphaning, the highest strain rates can be observed in the viscously coupled mantle material surrounding the slab. (b) With the initiation of orphaning intra-slab strain rates increase at 660 km depth as the slab starts to deform, shear and stretch. (c) Upon orphaning, strain rates at 660 km depth increase even further, reflecting the intense deformation and shearing of the slab.
3.4. Discussion

While the V3 model is missing the full thermodynamics of a chemically differentiated and multi-mineralic slab and mantle as in Earth, the strongly negative Clapeyron slope of $-2.5 \times 10^6 \text{ PaK}^{-1}$ used in the model, acts as a proxy for the rheological strength of the slab at 660 km in contrast with that of the mantle immediately surrounding the slab. Thus, it is clear that the slab strength relative to the surrounding mantle exerts a major control on the slab morphologies observed in the mid-mantle, including the new orphaning phenomenon.

3.4.1 Orphan Slabs in Tomography

Seismic tomography indicates that slab fragments are common in the deep mantle. The Tonga, Arabian, Japan and Central American slabs in particular (Fig.3.6a-b) all show flattened morphologies above 660 km, underlain in some tomography models by steeper separate portions at depth (Bina et al., 2001; Funiciello et al., 2003; Green, 2007; Sigloch et al., 2008; Sigloch, 2011). Taking these results at face value, this chapter boldly reinterprets these fragments as orphan slabs overlain by flattened parent slabs.

The slab orphaning mechanism provides a more direct interpretation for the deep, enigmatic slab fragments of the mid and lower mantle. A question arises, however, as to whether it is possible for orphaning to leave a distinct signal at the surface. This study suggests that this is possible because slab stretching, shearing, and flattening prior to orphaning do show an effect on trench motion (Fig. 3.2). However, subduction zones are complex and long lived systems, and often show multiple fragments. It would therefore, be imprudent to assume that one process can monotonously account for or be distinguishable from all the surface geological observables. This work suggests that slab orphaning is an independent, but complementary mechanism for shallow mantle slab fragmentation, and possibly particularly relevant for long lived subduction systems.
3.4. Discussion

Figure 3.6: Tomographic cross-sections generated using the submachine portal from Hosseini et al. (2018), of the (a) Tonga, (b) Japan, (c) Arabian and (d) Central American slabs. These slabs show a flattened slab at 660 km depth and a deeper fragment sinking independently beyond 660 km depth. The flattened slab (magenta arrow) above the 660 km phase transition is here interpreted as the parent slab and the deeper fragment (magenta circle), as the orphan. Difference in flattening and signs of mantle upwelling between parent and orphan (e.g. Central American slab) can indicate different stages and orphaning maturities.
This work identifies a new dynamical process to create and explain the enigmatic slab fragments observed at mid and lower mantle depths. Results presented in this chapter show that a combination of the resistance to sinking provided by the ringwoodite to bridgmanite and ferropericlase phase transition with the viscosity reduction of the slab and mantle between 660 km and 1000 km, coupled with a viscosity increase at 1000 km depth breaks the slab tip causing a portion of the slab to be abandoned. I have termed this new process slab orphaning. Slab orphaning is an important facet of subduction that may explain the presence of deep (> 600 km) slab fragments underlying present day active subduction systems.

Slab orphaning has important implications for tectonic reconstructions, many of which use a slab reference frame (e.g. van der Meer et al., 2017). Such reconstructions are based on the assumption that slabs sink vertically downwards, directly underlying their trenches. This assumption is also applied to calculate remnant slab sinking speeds, to infer slab ages as well as mantle viscosity (Čížková et al., 2012; van der Meer et al., 2010, 2017). These are then used in plate reconstruction models to inform the interpretation of surface features and to constrain the timing of processes such as island arc accretion. This approach has also been applied to date the onset and cessation of subduction (van der Meer et al., 2010, 2017). The presence of intact slabs overlying remnant slabs at depth has generally been interpreted to signal the onset of a new subduction after cessation due to shallow slab break-off (Li, 2002; Sigloch et al., 2008; Sigloch, 2011; Sigloch and Mihalynuk, 2013; Richards et al., 2011; Duretz et al., 2011; Schellart and Spakman, 2012). Slab orphaning, on the other hand, is not a precursor to subduction termination. The models presented in this work show that the on-going subduction continues and results in lateral motion of the flattened parent slab above 660 km depth.
3.4. Discussion

Slab orphaning also challenges previous interpretations of deep slab fragments as necessarily resulting from shallow slab break-off (see e.g., Sigloch et al., 2008; Sigloch, 2011; Sigloch and Mihalynuk, 2013; Schellart and Spakman, 2012; van der Meer et al., 2017). Here it is shown, for the first time, that slab fragmentation can also occur both at shallow and mid-mantle depths.

This chapter further propose a reinterpretation of the Tonga, Arabian, Japan and Central American deep slab remnants as orphan slabs. This could potentially reset the current understanding of the tectonic evolution and timing of these subduction zones and their associated geology. It is suggested here that these slabs would have reached the mid-mantle as a coherent slab with orphaning and abandonment occurring only at 660 km depth due to localised slab shearing and break-off as described in section 3.4.

Orphaning is not an exclusive but rather a complementary process. It does not exclude island arc accretion, or shallow slab break-off. All these processes can act in concert to give rise to the observed morphology at the surface and in the mantle. Nonetheless, it is clear that slab orphaning simplifies slab fragmentation. It does not require complex surface kinematics, such as multiple trench polarity reversals (Sigloch et al., 2008), to explain the abundant isolated portions of cold oceanic plate segments in the deep mantle.
Chapter 4

The Effect of a Continental Overriding Plate on Subduction Dynamics

The overriding plate is a fundamental piece of the subduction puzzle. Numerical simulations show that the addition of an overriding plate results in models that produce a better fit to the observations and more realistic trench retreat velocities (e.g. King, 2001; Butterworth et al., 2012; Sharples et al., 2014; Holt et al., 2015). The nature of the overriding plate is also very important. With a few notable exceptions, most subduction zones on Earth describe the convergence of oceanic and continental lithosphere (e.g. Heuret et al., 2011, and references therein). Continents are a source of heterogeneity and asymmetry. They introduce additional lithospheric stresses at the trench and further forcing to the subducting system. Through their positive buoyancy, continents also encourage and maintain single-sided subduction (Trubitsyn and Rykov, 1997).

The role of continental lithosphere in shaping mantle convection, plate driving forces and the planet’s thermal history has been explored and detailed in previous work (Trubitsyn et al., 2003; Conrad and Lithgow-Bertelloni, 2006; Van Summeren et al., 2012; Rolf and Tackley, 2011; Rolf et al., 2018, amongst others). However, few studies have examined the effect of the continental lithosphere on subduction evolution and slab dynamics (Holt et al., 2015). In general, continental lithosphere
exhibits behaviours that are decidedly different from those of the oceanic lithosphere or mantle (Yoshida, 2010). Rock deformation experiments also show that oceanic and continental lithospheric plates exhibit different strengths and different deformation styles (Kohlstedt et al., 1995; Jackson, 2002). The contrast in rheology, composition and behaviour between these two lithospheric plates, seems to indicate that the common numerical approach to subduction modelling (i.e. that of an oceanic overriding plate) cannot fully and satisfactorily explain the most prevalent form of subduction on Earth.

Limited previous numerical studies suggest that the presence of continents at subduction zones encourages steeper subduction angles, reduced overriding plate and trench mobilities, and reduced slab sinking rates (Sharples et al., 2014; Holt et al., 2015; Crameri and Lithgow-Bertelloni, 2018). The presence of a continental overriding plate also facilitates slab penetration into the lower mantle (Sharples et al., 2014; Crameri and Lithgow-Bertelloni, 2018). Crameri and Lithgow-Bertelloni (2018) observe that, while the presence of a continental overriding plate in 2-D numerical models does little to change the overall geometric evolution of the model in the upper mantle, its presence has subtle implications for the deeper slab morphologies. Due to its deep roots the continental lithosphere is highly coupled to the lower part of the upper mantle (Crameri and Lithgow-Bertelloni, 2018; Conrad and Lithgow-Bertelloni, 2006; Van Summeren et al., 2012). The increased coupling at depth can introduce additional forcing on the slabs of the transition zone and upper lower mantle, leading to behavioural and morphological changes.

Detailed study of slab behaviour at these deeper mantle depths has insofar only been explored within the context of an oceanic overriding plate (e.g. Čížková et al., 2007; Garel et al., 2014; Agrusta et al., 2017; Crameri et al., 2017; Goes et al., 2017). Given the above discussion, however, it seems that the continental overriding plate has a role to play in shaping the deeper slab morphologies of mid-mantle. This chapter, therefore, sets out to explore the effect of the continental overriding plate on the evolution of the deep slabs, their dynamics and morphology.
4.1 Methods

Similar to the previous chapter (Chapter 3), 2-D numerical models of ongoing subduction with visco-plastic rheology are implemented using the finite difference/volume code StagYY (Tackley, 2008). For this chapter a continental implementation is also included. As discussed in Chapter 2 the continent is 2000 km wide and 200 km deep meaning that, its roots extend through to the bottom of the asthenosphere to the top of the upper mantle. Compositionally, the continental material is identical to the mantle underneath it, with the exception that the former has a significantly lower reference density. The lower density of the continental material results from its buoyancy ratio, $B_c$ of -0.1. $B_c$ is simply the density contrast of the continent, $\Delta \rho_c$, divided by the thermal density variation $\rho \alpha \Delta T$. The more negative the value of $B_c$, the lighter the implemented continent. In this case, the density contrast of the continent with respect to the ambient mantle is $\sim -25 \text{ kgm}^{-3}$. The continental lithosphere is both light and high viscosity. This produces a continent that is less dense than the underlying mantle, thick, positively buoyant and very strong. Further details on the continental implementation and the parameters used can be found in Chapter 2, section 2.7 and in table 2.3.
The continental implementation is combined with four different mantle viscosity profiles (see table 4.1). The first profile, $V1$ represents the traditional and, until now, canonical viscosity profile for the mantle, where the viscosity increase is concurrent with the endothermic phase change at 660 km depth (Hager and Richards, 1984). In model $V2$, the phase change effects at 660 km depth are ignored and the lower mantle viscosity increase is shifted to 1000 km depth. In $V2$ the viscosity increase at depth is preceded by a low viscosity layer (LVL) between 660 km and 1000 km depth. Model $V3$ describes the phase change of ringwoodite to bridgmanite and ferropericlase at 660 km depth and a separate viscosity jump at 1000 km depth. Similar to $V2$, in $V3$ the viscosity increase at depth is overlain by an LVL that extends from 660 km to 1000 km as described in Rudolph et al. (2015). Model $V4$ tests the combined impact of the LVL and the continental overriding plate. This profile is identical to $V3$, except that the LVL between 660-1000 km is missing and the upper mantle viscosity extends to 1000 km, at which depth it then increases by an order of magnitude. As discussed in Chapter 2, $V3$ is the viscosity profile preferred by the geoid inversions and, thus, the viscosity profile preferred by this study. Details of the model variations and associated viscosity profiles can be found in table 4.1 below.

Table 4.1: The mantle viscosity profiles tested for the continental overriding plate implementation

<table>
<thead>
<tr>
<th>Viscosity Profile</th>
<th>LVL</th>
<th>Clapeyron Slope (PaK$^{-1}$)</th>
<th>Depth of Viscosity Jump (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$V1$</td>
<td></td>
<td>$-2.5 \times 10^6$</td>
<td>660</td>
</tr>
<tr>
<td>$V2$</td>
<td>✓</td>
<td>0</td>
<td>1000</td>
</tr>
<tr>
<td>$V3$</td>
<td>✓</td>
<td>$-2.5 \times 10^6$</td>
<td>1000</td>
</tr>
<tr>
<td>$V4$</td>
<td></td>
<td>$-2.5 \times 10^6$</td>
<td>1000</td>
</tr>
</tbody>
</table>
4.2 Results

In an attempt to understand the contribution of the continental overriding plate to the morphology of deep slabs, this study compares set-ups with continental and oceanic upper plates and examines variations in both slab parameters and trench behaviour. The analyses covered include; variations in slab morphology, slab and trench velocities, slab tip angles and slab evolutionary stages. Variations resulting from the type of overriding plate, are observed for all viscosity profiles to varying degrees and are described in the sections below.

4.2.1 Slab Morphologies

In agreement with previous work (Crameri and Lithgow-Bertelloni, 2018), this study finds that the evolution of slab morphology in the upper mantle is independent of the type of overriding plate. Similarly, the vertical viscosity profile of the mantle does not influence the slab evolution above 660 km depth. However, it is clear that the combination of the mantle viscosity profile and the type of overriding plate can determine the slab behaviour at mid and upper lower mantle depths (that is from 660 km depth and deeper). Depending on the nature of the mantle viscosity profile, the presence of an overriding plate can result in anchored, hooked and broken or orphaned slabs (c.f. table 4.2). The latter is a newly identified morphology, described for the first time by this study. As discussed in much greater detail in Chapter 3, slab orphaning is the terminology used to describe slab break-off directly at mid-mantle depths and is an inherently mid-mantle process.
4.2. Results

Table 4.2 and Figs. 4.1 and 4.2 show that the presence of a continental overriding plate can force slabs to adopt drastically different behaviour. The models most susceptible to the overriding plate influence are V1, V2 and V4. In a continent-ocean convergence, models V1 and V4 exhibit slabs that sink and are anchored below 660 km depth. On the other hand, the same viscosity profiles, produce slabs that are flattened above 660 km depth when the overriding plate is oceanic (Figs. a vs. b in 4.1 and 4.2). In V2, slabs assume a ‘hooked’ morphology when the overriding plate is continental (see table 4.2 and Fig. 4.3 a), similar to the morphology observed by Arredondo and Billen (2017) and Billen and Arredondo (2018) for strong slabs. When the overriding plate is oceanic, the same slab in V2 adopts a quasi vertical slab geometry (see Fig. 4.3 b).

Table 4.2: The viscosity profile and associated slab morphology in the presence of a continental overriding plate

<table>
<thead>
<tr>
<th>Viscosity Profile</th>
<th>Clapeyron Slope (PaK⁻¹)</th>
<th>Depth of Viscosity Jump (km)</th>
<th>Slab Morphology</th>
</tr>
</thead>
<tbody>
<tr>
<td>V1</td>
<td>$-2.5 \times 10^6$</td>
<td>660</td>
<td>Anchored</td>
</tr>
<tr>
<td>V2</td>
<td>$0$</td>
<td>1000</td>
<td>Hooked</td>
</tr>
<tr>
<td>V3</td>
<td>$-2.5 \times 10^6$</td>
<td>1000</td>
<td>Orphaned</td>
</tr>
<tr>
<td>V4</td>
<td>$-2.5 \times 10^6$</td>
<td>1000</td>
<td>Anchored</td>
</tr>
</tbody>
</table>

In the case of V3, however, the nature of the overriding plate has little effect on the slab break-off at depth. Fig. 4.4, illustrates orphaning in both intra-oceanic and continent-ocean systems, with continental lithosphere producing orphan slabs that are considerably larger than their counterparts in intra-oceanic systems.
4.2. Results

Figure 4.1: A continental overriding plate encourages slab tip penetration and anchoring below 660 km depth for model V1 (a). The same slab in V1 flattens above 660 km depth when the overriding plate is oceanic (b).
Considerable slab penetration and anchoring is also observed for $V_4$ when the overriding plate is continental in nature (a). Similar to $V_1$, $V_4$ exhibits slab flattening when subduction occurs in an intra-oceanic setting (b).

**Figure 4.2:** Considerable slab penetration and anchoring is also observed for $V_4$ when the overriding plate is continental in nature (a). Similar to $V_1$, $V_4$ exhibits slab flattening when subduction occurs in an intra-oceanic setting (b).
4.2. Results

Figure 4.3: Slab morphology for viscosity profile V2, showing a ‘hooked’ slab geometry in the presence of a continental overriding plate (a). Slab morphology, however, assumes a quasi-vertical shape when the upper overriding plate is oceanic (b).
Figure 4.4: Being an inherently mid-mantle process, orphaning occurs for both continental (a) and oceanic (c) overriding plates. A continental overriding plate however encourages bigger slab orphans and longer orphaning times (b vs. d) compared to its oceanic counterpart.
4.2. Results

4.2.2 Plate Velocities

Variations in slab morphology for continental and oceanic overriding plates are also accompanied with variations in trench velocities. As illustrated in Fig. 4.5 a-b and c-d, trench and slab velocities register an overall decrease as the slab reaches deeper mantle depths. A comparison of trench and slab sinking velocities for ocean-ocean and continent-ocean subductions, shows that the initial, ‘slab free-fall’ stage velocities (Funiciello et al., 2003) are higher for a continental overriding plate. As the subduction model matures and the slab reaches mid-mantle depths, trench velocities for the continental overriding plate average 2.5 cm year\(^{-1}\) while velocities for the oceanic overriding plate average around 4 cm year\(^{-1}\) (see Fig. 4.5). This is comparable to the trench velocities measured for the present day Earth where trench velocities for continental lithosphere on Earth average around 2.7 cm year\(^{-1}\) as opposed to 3.9 cm year\(^{-1}\) on average for oceanic overriding plate types (Jarrard, 1986; Bijwaard et al., 1998; Heuret and Lallemand, 2005; Lallemand et al., 2005). Similarly slab sinking velocities are also lower for models with a continental overriding plate. However, compared to the estimated present day slab sinking velocities on Earth, both continent-ocean and ocean-ocean models show significantly higher slab sinking velocities, possibly due to the ‘pristine’ nature of the mantle in numerical simulations. Unlike the real Earth, the mantle in these subduction simulations is not littered with upwellings, slab fragments or other subductions that can interact and affect the velocity of the sinking slab.
Fig. 4.5 also reveals that while in an intra-oceanic setting trench velocities decrease monotonically with time, for a continental overriding plate, trench mobility shows an transient increase after an initial sharp drop (see Fig. 4.5 a and Fig. 4.5 c). Similarly slab sinking velocities also show a consistent decrease when the overriding plate is oceanic (Fig. 4.5 d). In contrast, subduction in a continent-ocean setting is generally characterised by plateaus in slab sinking velocities when the slab reaches mid-mantle depths (Fig. 4.5 b) followed by further decrease as the slab sinks in the lower mantle. V2 is the exception to this trend, exhibiting initial fast slab sinking velocities that decline abruptly as the slab reaches the high viscosity lower mantle. The abrupt decline in slab sinking velocity is also mirrored in a decrease in trench velocity. The slab behaviour in V2 is irrespective of the overriding plate with the same trend observed for both continental and oceanic lithosphere (Fig. 4.5 c and d).
4.2. Results

Figure 4.5: Variations in slab sinking and trench velocities in cm year$^{-1}$ for models V1-V4 with continental (a,b) and oceanic (c,d) overriding plates, exhibiting an overall decrease in plate velocities. Following the initial fast velocities of the ‘slab free-fall stage’ (Funiciello et al., 2003) velocities for continent-ocean systems are lower than those observed for ocean-ocean systems. Ocean-ocean subduction is characterised by steeper gradients indicating sharp declines in velocity. When the overriding plate is continental, trench velocities exhibit a period of trench velocity increase, followed by overall gentler declines that tend towards plateaus in both slab sinking and trench velocities. V2 is the only exception, exhibiting rapid slab sinking velocities accompanied by steep drops in trench velocities irrespective of the overriding plate.
4.2. Results

4.2.3 Slab Tip Evolution

A relationship between the angle at the slab tip and the nature of the overriding plate has been suggested by previous work (Lallemand et al., 2005). After analysing the progressive variations of the slab tip dip (measured directly at the tip of the slab), throughout V1-V4, this study observes a general trend of slab tip flattening for all slabs at mid-mantle depths. This behaviour is seemingly independent of the overriding plate lithosphere. Slab tip dip switches from steeply dipping values of 60° or more at the start of the simulation to shallower, gentler dips of 30° or less towards the end of the simulation (Figs. 4.6 and 4.7). V4 is the only notable exception, where the slab tip dip increases from ~65° to ~90° when its overriding plate is continental.

In contrast to Lallemand et al. (2005), this study does not observe any conclusive correlation between the generic slab tip flattening and the nature of the overriding plate for any of the four viscosity profiles implemented. Results instead suggest that in general the presence of a continental overriding plate encourages slab penetration into the lower mantle and steeper slabs overall (as opposed to just the slab tips) (Sharples et al., 2014; Holt et al., 2015; Crameri and Lithgow-Bertelloni, 2018).
4.2. Results

4.2.4 The Relationship between the Trench Velocities and the Slab Tip Behaviour

Inspired by the work of Lallemand et al. (2005), this study investigates the inter-relationship between the trench velocity, the slab tip behaviour and the overriding plate nature. While the previous section highlights the apparent lack of correlation between slab tip flattening and the overriding plate, Figs. 4.6 and 4.7 indicate a correlation between the slab tip angle and the trench velocity. The latter is strongly affected by the nature of the overriding plate. For ocean-ocean systems, decreasing slab-tip angles are correlated with decreasing trench velocities with time (see Figs. 4.6 and 4.7). These observations hold true irrespective of the viscosity profile or the slab morphology. Trench velocities and slab tip angles appear to be anticorrelated when the overriding plate is continental, with the slab tip dip increasing as trench velocity decreases and vice-versa (see Figs. 4.6 and 4.7). This trend is particularly noticeable for profiles V1 and V4, the dynamics of which are not complicated by mid-mantle processes such as orphaning. Slab tip steepening is particularly prevalent for V4 (see Fig. 4.7 g) and reflects the steepening slab tip as it anchors in the upper lower mantle and sinks vertically downwards.

The behaviour of V2 is particularly intriguing. This models exhibits similar slab and trench behaviour irrespective of the nature of the overriding plate (see Fig. 4.6). The reduced resistance to slab sinking resulting from the missing phase transition of ringwoodite to bridgmanite and ferropericlase induces high slab sinking velocities and trench motion, overprinting any signal from the overriding plate. This behaviour highlights the out-sized role of the phase transition of ringwoodite to bridgmanite and ferropericlase in shaping, viscous flow and slab dynamics (Christensen, 1984).
4.2. Results

$V_3$, similar to $V_1$ and $V_4$ also shows increases in trench velocities with a continental overriding plate. However, in this case, slab tip angles remain in consistent increase until, similar to its oceanic counterpart, slab tip steepness is abruptly reduced (see Fig. 4.7). This behaviour is independent of the overriding plate and reflects the slab tip steepening as it shears and splits away from the parent. The more extensive the shearing and stretching between parent and orphan, the more the slab tip/proto-orphan approximates a Stokes’ blob and, thus, the steeper it becomes as it sinks under its own weight.

**Figure 4.6:** Trench velocities in cm year$^{-1}$ (blue) and slab tip angles in degrees (orange) for $V_1$ with continental (a) and oceanic (b) overriding plate and for $V_2$ with continental (c) and oceanic (d) overriding plate
Figure 4.7: Trench velocities in cm year$^{-1}$ (blue) and slab tip angles in degrees (orange) for V3 with continental (a) and oceanic (b) overriding plate and for V4 with continental (c) and oceanic (d) overriding plate. Models V1-V3 show a decrease in the trench velocity and slab tip angles, signifying flattening and reduced trench mobility. V4 exhibits an increase in the slab tip angle when the overriding plate is continental (g).
4.2. Results

4.2.5 The Overriding Plate and Slab Evolutionary Stages

Further to the above analyses, this study also includes the slab bending ratio $S_b$. $S_b$ is calculated by dividing the slab tip angle by its shallow counterpart at $\sim 72$ km depth. This suggests the overall steepness of the slab morphology as opposed to just that of the slab tip. The benefit of this approach is that it provides a measure of the intra-slab bending and indicates the maturity of the subducting slab (see Fig. 4.9).

In the presence of a continental overriding plate clear slab behaviour trends emerge. When $S_b > 1$ the subducting slab is in its juvenile stage (see Fig. 4.8). This stage is characterised by rapid penetration and a steep slab that is rolling trenchwards through the upper mantle (see Fig. 4.9). When $S_b \approx 1$ (see Fig. 4.8), there is little slab tip penetration, increased trench retreat (as opposed to slab rollback) and significant intra-slab bending. At this stage the slab would have reached its maturity in the mid/upper mantle depths (see Fig. 4.9). A slab’s old age is defined by an $S_b < 1$ (see Fig. 4.9). Small values of $S_b$ indicate a flattened slab tip with limited penetration into the lower mantle (see Fig. 4.8). At this stage subduction is dominated by the lateral motion of the flattened tip. When a slab is in its old age, its overall morphology is drastically different from that of its juvenile stage.

In some cases old age is followed by a period of rejuvenation (see Fig. 4.9). Rejuvenation occurs for both $V1$ and $V4$ (see Fig. 4.8). In both cases $S_b$ values increase above 1 and reflect a renewed period of slab steepening and penetration, although this only occurs for the slab tip below 660 km depth (see Fig. 4.8). The slab portion above 660 km depth continues to flatten slowly, with its tip acting as an anchor, forcing the slab to slowly unbend, straighten out and fall into the lower mantle. Consistently high $S_b$ values for $V3$ (see Fig. 4.8), indicate that orphaning occurs for slabs in juvenile stages. Due to shearing and stretching, a slab that is undergoing orphaning cannot evolve into maturity. Orphaning dramatically forces a slab out of its juvenile stage straight into old age. $V2$ is also characterised by a prolonged juvenile stage (see Fig. 4.8). The lack of resistance to sinking promotes steep slab morphologies and rapid penetration into the upper lower mantle. In $V2$
4.2. Results

the slab’s transition into maturity and old age occurs over a short period of \( \sim 3 \) to 4 million years. In this case, slab tip flattening accompanies intra-slab bending as the slab adopts a ‘hooked’ morphology. As slab tip penetration decreases in comparison to that of its juvenile stage, subduction is accommodated by the lateral retreat of the ‘hooked’ and flattened slab tip.

![Figure 4.8: Trench velocities in cm year\(^{-1}\) (blue) and ratio of slab bending \( S_b \) (orange) for models V1-V4 with a continental overriding plate. Ratios of \( S_b \) higher than 1 indicate a juvenile slab. This stage is characterised by steep slabs, rapid penetrations, fast slab rollback and decreasing trench velocities. Ratios of \( S_b \) equal to 1 indicate slab maturity. This stage is characteristic of slabs undergoing significant intra-slab bending, decreased penetration and increased trench mobility. Ratios of \( S_b \) below 1 indicate old age. This stage represents significantly reduced penetration, flattened slab tips, prominent lateral migration of the slab tip and increased trench retreat. In some cases, old age can be followed by a period of rejuvenation. This period is characterised by \( S_b \) values above 1, steepened slab tip and renewed penetration.](image-url)
Figure 4.9: Schematic of the slab bending ratio $S_b$. $S_b$ ratio higher than 1 is associated with juvenile slabs, fast penetration, slab rollback and steep angles. $S_b$ equal to 1 represents mature slabs with significant intra-slab deformation, limited penetration and increased trench mobility. $S_b$ less than 1 is representative of old age, a flattened slab tip, little to no penetration and lateral motion of the deflected slab. In some cases a steepened slab tip allows for rejuvenation of the slab, higher $S_b$ ratios and renewed penetration.
4.3 Discussion

The results presented in this chapter clearly illustrate that within the upper mantle, it is neither the overriding plate nor the ambient viscosity of the mantle that define the slab geometry. Reaffirming previous findings, all slabs in the upper mantle, irrespective of the model set-up, undergo rapid sinking and steepening (e.g. Fucicello et al., 2003; Crameri and Lithgow-Bertelloni, 2018). However, morphological diversity peaks once the slab transitions from the upper mantle into mid-mantle depths, indicating that this is due to a complex interaction of upper plate forcing, viscous coupling (of both overriding plate and slab with the surrounding mantle), and the variation of the local slab strength $\text{vis} - a - \text{vis}$ that of the ambient mantle.

The results shown here highlight the glaringly conspicuous role of the overriding plate type on slab morphologies. Replacing an oceanic overriding plate with a continental one results in drastically different behaviour for identical slabs in identically set-up models with the same initial conditions. $V_1$ and $V_4$, in particular drive this point home (see movies $V_1$ and $V_4$ vs. $V_1\text{ Continent}, V_4\text{ Continent}$ in Appendix B). In these two cases, an oceanic overriding plate, results in deflected and flattened slabs at 660 km depth. However, the same slab interacting with the same ambient viscosity profile develops an entirely different morphology when the overriding plate is continental lithosphere. In the latter case, the slab tip subducts below 660 km depth and is anchored in the upper lower mantle (see Figs. 4.1 and 4.2). The anchoring of the slab tip is accompanied by the flattening of the slab body above 660 km depth for both cases (refer to movies $V_1\text{ Continent}, V_4\text{ Continent}$ in Appendix B). The continental overriding plate, therefore, encourages a slab that is otherwise inclined to flatten, to reach the upper lower mantle.

Further to its contributions at depth, the nature and type of the overriding plate lithosphere also determines its behaviour. Similar to what is observed on Earth (Jarrard, 1986; Heuret and Lallemand, 2005; Lallemand et al., 2005), numerical simulations with a continental overriding plate register significantly reduced mobility when compared to their oceanic overriding plate counterparts. This is due to the strong coupling between the high viscosity continent and the mantle underneath.
4.3. Discussion

it (Van Summeren et al., 2012; Crameri and Lithgow-Bertelloni, 2018). Similar to what is observed on Earth, reduced upper plate mobility results in curtailed trench retreat. While this is not necessarily the case on Earth, numerical models of subduction show that reduced overriding and trench mobility encourage steeper subducting slabs (Lallemand et al., 2005; Crameri and Lithgow-Bertelloni, 2018). In the case of V1 and V4, the slab steepening encouraged by the presence of the continental overriding plate extends the slab juvenile stage (Fig. 4.9), which leads to the slab tip penetration into the upper lower mantle. The swiftly sinking, steeply dipping slab induces viscous coupled flow with a downward, vertical orientation. This drags the slab further downwards aiding its penetration into the more viscous lower mantle. Once the slab tip sinks below 660 km depth, it essentially acts as an anchor, pulling the up-dip slab portion downwards. This rejuvenates the flattening slab and allows its tip to penetrate deeper into the mantle (see movies V1_Continent, V4_Continent in Appendix B).

While the presence of the continent encourages deeper slab tip penetration, it is the variation of the ambient viscosity of the mantle with depth that dictates the behaviour and morphology of the slab. The latter two are the physical representation of the slab’s response to changes in its relative strength with respect to that of the ambient mantle. Depending on the dynamics of this relationship the slab either flattens, penetrates (with variations in timing and style e.g. anchoring) or orphans, so much so, that the latter is entirely independent of the overriding plate type. For slab orphaning the overriding plate contribution is its influence on the orphan size and orphaning timescales. When the overriding plate is continental, orphan sizes are larger (~656 km compared to ~271 km for an intra-oceanic system) and orphaning time scales are longer (~5.5 Myr as opposed to the intra-oceanic time scale of ~2 Myr).

Moreover, as Fig. 4.10 illustrates, the difference in overriding plate translates into distinctively dissimilar viscous flow patterns (see movie V3_Continent in Appendix B). These differences in the induced flow field are a direct consequence of the thick and deep continental roots that extend to the bottom of the asthenosphere.
and into the shallow upper mantle. Due to its considerable thickness, the highly viscous continental lithosphere replaces the lubricating, low viscosity asthenosphere. Resulting, thus, in an overriding plate that is highly coupled to the upper mantle upon which it flows (Van Summeren et al., 2012). Fig. 4.10 d-f shows that the continental overriding plate dominates the viscous flow field of the upper mantle. As the continent moves trenchward, its deep roots shear and drag the shallow upper mantle (Conrad and Lithgow-Bertelloni, 2006; Van Summeren et al., 2012) towards the trench and subducting slab. This exerts a trenchward force on the up-dip portion of the slab above 660 km depth, forcing it to bend and slowly flatten. Below 660 km depth, the influence of the continental overriding plate is reduced significantly and the slab-induced viscous flow becomes more prominent. Contrary to the horizontal flow in the upper mantle, the coupled viscous flow at depth has a downward, vertical direction. This deep vertical flow induced by the sizeable slab tip, coupled with its considerable negative buoyancy, encourages the downward sinking of the down-dip slab portion.

In contrast to intra-oceanic orphaning, where the change in flow direction occurs at ∼1000 km depth induced by the viscosity increase, in a continent-ocean setting, the change in flow direction occurs at ∼660 km depth. In the latter case the change in flow direction results from the downward sinking of vertical slab tip/ proto-orphan itself. At ∼660 km depth the influence of the continental overriding plate is reduced significantly and the viscous coupling of the slab with its immediate surrounding mantle becomes more important. The coupled flow helps to drag the proto-orphan into the lower mantle, which together with its considerable size and negative buoyancy allows it to overcome the increased viscosity and resistance to sinking below 1000 km depth. Whether the change in flow direction occurs at mid-mantle depths or deeper is irrelevant to slab orphaning. What is important, however, is the establishment of two opposing flow directions in the upper and mid/lower mantle. This is a crucial aid to the slab shearing taking place at 660 km depth which results from the opposing buoyancy forces in the up and down dip portions of the slab as discussed in detail in Chapter 3.
4.3. Discussion

Figure 4.10: Viscosity plots showing the induced viscous flow patterns for a continental overriding plate (a-c) and an oceanic overriding plate (d-f). The thick continental roots induce a strong, viscous, horizontal trenchward flow across the upper mantle. When the overriding plate is oceanic (d-f) the upper mantle is dominated by the viscous coupling of the slab with the immediate surrounding mantle. Both cases produce opposing flow in the upper and lower mantle which aids the shearing, stretching and the eventual orphaning of the slab at 660 km depth.
Another noticeable difference between slab orphanging in the presence of a continental overriding plate and orphanging in an intra-oceanic setting is the steepness of the slab prior to orphanging. A comparison of the slab tip dip angles for intra-ocean and continent-ocean convergence shows that in both systems slabs reach 660 km depth with comparable tip dip angles (67° and 69° respectively). However, just prior to orphanging, slabs subducting under an oceanic overriding plate exhibit angles that are $\sim 10^\circ$ shallower than their counterparts with a continental overriding plate. Shallower slab tip angles just prior to the initiation of slab orphanging correspond to an increase in the lateral extent of the low density lens at 660 km depth (see Fig. 4.11 d-f).

The presence of a lens of lower density in the cold slab core results from the temperature dependence of the Clapeyron slope and reflects the delayed phase transition of ringwoodite to bridgmanite and ferropericlase. Its presence hinders and resists the slab’s sinking into the lower mantle. However, its effect can be offset by the viscous coupling of the slab with the surrounding mantle which helps to drag the slab tip downwards.

When the overriding plate is continental, steeper slab tip angles at depth are sustained for longer, reflecting a delay in the lateral expansion of the low density lens (see Fig. 4.11 a-c). The spatial limitation of this localised positive buoyancy force makes it possible for more slab material to sink vertically downwards and reach the upper lower mantle, resulting thus in a larger orphan. This implies that the negative buoyancy forces driving subduction dominate the system prior to orphanging. Orphaning represents a tipping point, after which the forces resisting subduction become more prominent leading to slab deflection and flattening.
Figure 4.11: Plot of density illustrating a lens of lower density within the slab core at 660 km depth and its evolution in systems with a continental overriding plate (a-c) and an oceanic overriding plate (d-f). When the overriding plate is continental the extent of low density lens is spatially limited for longer enabling more slab material to subduct to deeper depths.
4.3. Discussion

Similar to previous work (Sharples et al., 2014; Holt et al., 2015; Crameri and Lithgow-Bertelloni, 2018), this study observes that in a continent-ocean setting subduction is overall slower, trench mobility is reduced and the slab-mantle transition zone interaction is considerably longer. While the slab evolution in the upper mantle is similar across simulations with and without a continental overriding plate, at mantle transition zone depths, these change significantly. For instance, replacing an oceanic overriding plate with a continental one can induce an otherwise deflecting slab at 660 km, to anchor into the lower mantle. Moreover, this study also notes that a continental overriding plate favours prolonged orphaning timescales and, as a result, bigger orphans.

Despite the seemingly insignificant role of the continent on the geometry of a subducting system, its presence is crucial in shaping the morphologies and behaviour of deep slabs. Therefore, for a better understanding of the diversity and range of deep slab morphologies observed in seismic tomography, one must first and foremost thoroughly understand the dynamic relationship between the local slab strength and that of the ambient mantle and how this can be modified through additional forcing from the overriding plate.
Chapter 5

Slab Orphaning Regimes: The Missing Link Between Penetrative and Deflective Slab Morphologies

5.1 Introduction

As discussed in Chapter 3, seismic tomography has revealed a plethora of slab morphologies at mid-mantle depths, including flattened, penetrating and orphaned slabs. The latter is a new phenomenological slab behaviour identified and described for the first time in Chapter 3. Orphaning describes the abandonment of the slab tip by its parent directly at mid-mantle depths. In contrast to shallow slab break-off (i.e. slab fragmentation at depths < 300 km e.g. Duretz et al., 2011, 2012), slab orphaning allows for deep slab fragmentation without interruptions to the overarch- ing subduction regime. When a slab orphans, it slowly sinks vertically downwards towards the CMB behaving as a Stokes fluid parcel. Its deflected parent, however, continues to travel laterally through the mantle at 660 km depth. The latter accommodates and sustains the ongoing subduction of the plate at the trench. The resulting morphology is that of a steeply vertical slab fragment overlain by a flattened slab at 660 km depth. Similar morphologies have been observed to be particularly common at ancient, long-lived subduction zones such as the Tethys, Farallon and the Tonga.
5.1. Introduction

Previous interpretations have suggested that the presence of deep slab fragments represent a break in subduction due to a change in the tectonic regime (e.g. Wortel, 2000; Sigloch et al., 2008; Sigloch and Mihalynuk, 2013; van der Meer et al., 2017). In this interpretation the slab breaks at shallow depths close to the trench and slowly makes its way down through the upper mantle, eventually a new subduction regime re-initiates subduction and the fragmented slab is overlain by the newly subducting slab (e.g. Wortel, 2000; Sigloch et al., 2008; van der Meer et al., 2010; Richards et al., 2011; Sigloch, 2011; Schellart and Spakman, 2012; van der Meer et al., 2017). This interpretation has been used as a tool to calculate the upper mantle viscosity, to reconstruct paleo plate motions and the location of paleo trenches, to estimate the cessation, duration and initiation of subduction events and to calculate the ages of slabs in the mantle, amongst other things (e.g. van der Meer et al., 2010, 2017). The main crux of this interpretation is the assumption that trenches are stationary through time and space, despite evidence from both geology and numerical models that this is not the case (e.g. Yoshioka et al., 2010; Shephard et al., 2017). Furthermore, in order to sustain stationary trenches through time, previous work invokes subduction polarity reversals, trench jumps, rapid subduction of entire oceans, and/or complex surface plate kinematics (Sigloch et al., 2008; Yoshioka et al., 2010; van der Meer et al., 2010; Sigloch, 2011; Shephard et al., 2017; van der Meer et al., 2017).

On the other hand, as per Chapter 3, slab orphaning provides a simple, alternative to the above. Orphaning suggests that these enigmatic and deep fragments do not necessarily always represent a cessation in subduction or a shift in the tectonic regime. In this work I propose that slab orphaning is one of the possible, natural evolutionary pathways a slab take as its subduction zone matures and it reaches deeper depths. Orphaning occurs in a hitherto unexplored parameter space between slab penetration and flattening. It describes a previously missing link between these two behaviours and slab morphologies. Indeed, orphaning details one of the possible ways a subducting slab can switch from a penetrating to a flattened mode. This chapter describes the range of conditions under which orphaning occurs and
explores the effects of various subduction parameters such as subduction angles and overriding plate types. Furthermore, this chapter expands on the previous by examining in further depth the interdependence of the slab strength and the Clapeyron slope. These two parameters are proxies for the dynamic slab-mantle strength relationship and its contribution to slab dynamics and morphology.

### 5.2 Methods

As detailed in Chapter 2 and similar to Chapters 3 and 4, I implement a suite of 2D numerical models of visco-plastic, whole-mantle convection with self-consistent (i.e. not kinematically driven), single-sided subduction and trench motion through time (Crameri et al., 2012b; Crameri and Tackley, 2015; Crameri and Lithgow-Bertelloni, 2018). Variations are implemented to model V3 (see Chapter 2, section 2.4 and table 2.1), the ambient mantle viscosity of which follows from Rudolph et al. (2015). Other model variations (see table 5.1) reflect different initial conditions and define the parameter space explored in this chapter. Similar to previous results, these models are analysed and post processed using StagLab 3.0 (Crameri, 2018)(http://doi.org/10.5281/zenodo.1199037).
5.2. Methods

Table 5.1: Model variations reflecting the different initial conditions. Group 1 shows variations in slab strength with Clapeyron slope values of $-2.5 \times 10^6$ PaK$^{-1}$. Group 2 depicts variations in slab strength with Clapeyron slope values of $-2.0 \times 10^6$ PaK$^{-1}$. Group 3 represents variations in slab strength combined with Clapeyron slope values of $-1.5 \times 10^6$ PaK$^{-1}$. In group 4, weak slabs strengths are combined with Clapeyron slope values of $-1.0 \times 10^6$ PaK$^{-1}$. Groups 5 and 6 show variations in the subduction angle for intra-oceanic (group 5) and continental-ocean subduction (group 6) from 20° to 90°.

<table>
<thead>
<tr>
<th>Model</th>
<th>Slab Strength (MPa)</th>
<th>Clapeyron Slope (PaK$^{-1}$)</th>
<th>Slab Dip Angle (°)</th>
<th>Continental Overriding Plate</th>
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</tbody>
</table>
5.3 Results

The model variations detailed in table 5.1, constrain the bounds of the slab orphaning parameter space. This is mapped out in Fig. 5.1, which clearly defines the orphaning regime as an intermediary, transitional stage occupying the sweet spot between the two major slab endmember morphologies (i.e. flattened and penetrative slabs).

Slab orphaning is the slab’s natural response to an opposing force balance in the up- and down-dip portions of the slab at 660 km depth. Above 660 km depth, the temperature dependence of the Clapyeron slope introduces a lens of low density material within the slab core, which resists subduction (see Figs. 3.4 and 3.3). This is offset by the viscously coupled flow that drags the slab tip downwards into the LVL between 660 km and 1000 km depth, which in itself encourages faster sinking velocity due to its reduced viscosity (see Fig. 3.3). As discussed in Chapter 3, this induces shearing and stretching within the proto-orphan, at 660 km depth. Deformation of the proto-orphan is further enhanced by a change at depth in the direction of the viscously coupled flow, resulting in opposing flow directions at upper and mid-mantle depths (see Chapters 4 and 3). This forces the up-dip slab portion to flatten and encourages the down-dip portion to sink vertically downwards. Shearing and deformation within the slab are accommodated through the plastic deformation of the slab (see Fig. 3.4 e) until the slab splits into a flattened parent at 660 km depth and a steep orphan at depths > 660 km.

Slab orphaning is independent of the initial conditions or the model set-up. Indeed, the results I present here highlight the common place nature of this morphology. I find that orphaning can occur for a wide range of subduction parameters, including different slab strengths and subduction angles. Furthermore, as shown in Chapter 4, the nature of the overriding plate does not influence in anyway whether a slab orphans or not.
Figure 5.1: Regime diagram for orphaning, penetrating and deflected slab morphologies. The orphaning regime space is highlighted in orange and shows that orphaning is prevalent for a range of slab strengths and Clapeyron slope values. Slab orphaning describes an intermediary space in the regime diagram between deflected and penetrating morphologies, indicating that orphaning can be one way for slabs to switch from one mode to the other.
5.3. Results

5.3.1 Slab Strength

Fig. 5.1 and table 5.2 clearly illustrate that slab orphaning is a persistent phenomenon, common to both strong and weaker slabs and can be observed within a range of Clapeyron slope values (see Fig. 5.2). Fig. 5.1 also indicates that for strong slabs to orphan, they require a considerably negative Clapeyron slope value of \(-2.5 \times 10^6 \text{ Pa K}^{-1}\). Introducing less negative Clapeyron slope values for these strong slabs (with yield stress values of 600 and 500 MPa), pushes the slab behaviour out of the orphaning regime into a penetrative one. On the other hand, maintaining the same Clapeyron slope value of \(-2.5 \times 10^6 \text{ Pa K}^{-1}\) but reducing the slab strength to 400 MPa and below firmly places the slab into a deflected regime. Slab orphaning is possible at all slab strengths between 600 and 200 MPa, provided this is offset by the right amount of resistance provided by the effect of the Clapeyron slope. Weaker slabs (400 MPa and lower) require less negative Clapeyron slopes values of \(-2.0 \times 10^6 \text{ Pa K}^{-1}\) and \(-1.5 \times 10^6 \text{ Pa K}^{-1}\) to orphan. When the slab strength is less than 200 MPa, the slab behaviour is drastically altered, with slab yield stress values of 100 MPa producing slabs that are too weak to reproduce any of the morphologies observed in seismic tomography. Furthermore, Clapeyron slope values of \(-1.0 \times 10^6 \text{ Pa K}^{-1}\) and lower do not offer enough resistance to either flatten or orphan slabs, even when these have low yield stress values of 200 MPa.

Interestingly, at Clapeyron slope values of \(-2.0 \times 10^6 \text{ Pa K}^{-1}\), it is possible to observe all three modes of slab behaviour. At this Clapeyron slope, slabs are equally inclined to flatten, deflect or orphan. The factor that determines the morphology in this particular case is the slab strength, which in and of itself is a function of the slab age. This is in agreement with previous studies that have suggested that the age of the subducting slab can play a role in determining its morphology (e.g. van der Hilst and Seno, 1993; Goes et al., 2011; Garel et al., 2014; Agrusta et al., 2017; Goes et al., 2017).
Figure 5.2: Slab orphaning for slab yield stress values of 600 MPa (a), 400 MPa (b) and 200 MPa (c) with respective Clapeyron slope values of $-2.5 \times 10^6$ PaK$^{-1}$, $-2.0 \times 10^6$ PaK$^{-1}$, $-1.5 \times 10^6$ PaK$^{-1}$.
5.3. Results

For a better grasp on the nuances of orphaning I compare orphaning timescales, orphan sizes and the associated amount of trench retreat for the different slab strengths and Clapeyron slopes depicted in Fig. 5.1. As can be observed from table 5.2, the weakest slab from model $V3(200)_{iii}$, produces the largest orphan. This same model also exhibits the longest orphaning time lengths and the largest amount of orphan toe deformation (see Fig. 5.2 f). On the other hand, models $V3(500)_i$ and $V3(300)_{ii}$ are both associated with the smallest orphan and the shortest orphaning time scales. The amount of trench retreat across all models is comparable and is in the range of 460 km. The two outliers are models $V3(600)_i$ and $V3(300)_{ii}$, which exhibit impressive trench retreats of 689 km and 1380 km respectively. While a comprehensive correlation for all the orphaning behaviours described in table 5.2 is difficult, it is possible to tease out some important trends. The most obvious of these is the relationship between the orphan size and the amount of time it takes for the orphan to be separated from its parent. The longer the orphaning period, the bigger the orphan. This is because extended orphaning timescales allow for more slab material to sink below 660 km before the slab is abandoned by its parent. Intriguingly, the largest orphaning sizes and longest orphan timescales are associated with those slabs of medium and weakest strengths (models $V3(400)_{ii}$ and $V3(200)_{ii}$).

On the other hand there seems to be no correlation between the amount of trench retreat and the orphaning timescales, orphan sizes, slab strength or any other variable. This seems to indicate that trench retreat is a complex process (Jarrard, 1986; Tagawa et al., 2007; Torii and Yoshioka, 2007), shaped by many factors such as slab stiffness and strength, arc deformation and upper mantle flow to name a few (e.g. Funiciello et al., 2003; Schellart, 2004; Faccenna et al., 2007; Di Giuseppe et al., 2008; Čížková and Bina, 2013; Boutelier and Cruden, 2013), most of which are not taken into account here.
Table 5.2: Orphan sizes, orphaning timescales and associated trench retreat compared across orphan variants with different slab strengths and Clapeyron slopes.

<table>
<thead>
<tr>
<th>Model</th>
<th>Slab Strength (MPa)</th>
<th>Clapeyron Slope (PaK$^{-1}$)</th>
<th>Slab Dip Angle (°)</th>
<th>Orphaning Time (Myrs)</th>
<th>Orphan Size (km)</th>
<th>Trench Retreat (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$V_3(600)_i$</td>
<td>600</td>
<td>$-2.5 \times 10^6$</td>
<td>30</td>
<td>3</td>
<td>271</td>
<td>689</td>
</tr>
<tr>
<td>$V_3(500)_i$</td>
<td>500</td>
<td>$-2.5 \times 10^6$</td>
<td>30</td>
<td>1.5</td>
<td>100</td>
<td>463</td>
</tr>
<tr>
<td>$V_3(400)_{ii}$</td>
<td>400</td>
<td>$-2.0 \times 10^6$</td>
<td>30</td>
<td>3.5</td>
<td>364</td>
<td>475</td>
</tr>
<tr>
<td>$V_3(300)_{iii}$</td>
<td>300</td>
<td>$-2.0 \times 10^6$</td>
<td>30</td>
<td>1.5</td>
<td>144</td>
<td>1380</td>
</tr>
<tr>
<td>$V_3(200)_{iii}$</td>
<td>200</td>
<td>$-1.5 \times 10^6$</td>
<td>30</td>
<td>4.5</td>
<td>684</td>
<td>451</td>
</tr>
</tbody>
</table>

5.3.2 Subduction Angles

Slab orphaning is also independent of the subduction angle and can be observed for angles ranging from 30° to 90°. Steep subduction angles of 50° and higher produce curved and overturned orphans (see Figs. 5.3 d-h and 5.4 d-h). However, when the slab angle at the trench is shallow (20°), slabs either flatten above 660 km depth if the overriding plate is oceanic or result in a shallow flat slab if the overriding plate is continental (see Figs. 5.3 a and 5.4 a).

5.3.3 Orphaning in an Intra-Oceanic Subduction

Fig. 5.3 a-h depicts slab morphologies for model $V_3$ for subduction angles ranging from 20° to 90°. When the slab angle at the trench is shallow (20°), the subducting slab deflects and flattens at 660 km depth. Penetration to mid-mantle depths, is followed by orphaning for all models with subduction angles of 30° and higher. Subduction angles of 30° and 40° result in a steep quasi-vertical orphan slab (Fig 5.3 b and c). Angles of 50° and steeper, on the other hand, produce curved and overturned orphan slabs, with polarities and orientations that are the reverse of their parents. A notable exception is model $V_3(80°)$ (Fig. 5.3 g), with subduction angles of 80°. This case is characterised by a steep vertical slab, penetration into the lower mantle and shallow slab break-off. Model $V_3(70°)$, with subduction angle of 70° (Fig. 5.3 f) is also worth highlighting. Similar to its shallower (50°- 60°) and steeper (90°) counterparts, $V_3(70°)$ exhibits a curved and overturned slab, however, in this case, orphaning is also accompanied by double sided subduction.
Figure 5.3: Slab morphologies for subduction angles of 20° to 90° for an oceanic overriding plate. Shallow subduction angles encourage slab flattening at 660 km depth (a). Subduction angles of 30° and 40° (b-c) exhibit orphaning. Subduction angles of 50° and steeper favour convex slab tip curvatures and overturned orphan slabs (c-f and h). Subduction angles of 70° and 80° encourage double sided subduction and shallow slab break-off respectively.
Adding further insight to the above, table 5.3 details the orphaning statistics for the associated subduction angle. Orphaning timescales are all in the range of 3.5 to 1.5 Myrs. The longest orphaning timescale is for $V3(50^\circ)$ and the shortest is for $V3(90^\circ)$. Contrarily to the models discussed above, the longest orphaning timescales do not necessarily correspond to the largest orphan sizes, when one considers variations in the subduction angle. Table 5.3 shows that the largest orphans are produced by the models with the steepest angles ($V3(70^\circ)$ and $V3(90^\circ)$). Fig. 5.3 also shows that subduction angles of 70° and 90° produce the steepest slab tip curvature and biggest orphan overturn. Most models show trench retreat values of 629 - 690 km, with $V3(70^\circ)$ and $V3(40^\circ)$ exhibiting significantly reduced and pronounced trench retreat distances, respectively. Similar to the discussion above (see section 5.3.1), there is no obvious correlation between the amount of trench retreat and any orphaning variable recorded in table 5.3.

**Table 5.3:** Orphan sizes, orphaning timescales and associated trench retreat compared across all orphan variants in an intra-oceanic subduction.

<table>
<thead>
<tr>
<th>Model</th>
<th>Slab Strength (MPa)</th>
<th>Clapeyron Slope (PaK$^{-1}$)</th>
<th>Slab Dip Angle (°)</th>
<th>Orphaning Time (Myrs)</th>
<th>Orphan Size (km)</th>
<th>Trench Retreat (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>V3(20°)</td>
<td>600</td>
<td>$-2.5 \times 10^6$</td>
<td>20</td>
<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>V3(30°)</td>
<td>600</td>
<td>$-2.5 \times 10^6$</td>
<td>30</td>
<td>3</td>
<td>271</td>
<td>689</td>
</tr>
<tr>
<td>V3(40°)</td>
<td>600</td>
<td>$-2.5 \times 10^6$</td>
<td>40</td>
<td>3</td>
<td>357</td>
<td>772</td>
</tr>
<tr>
<td>V3(50°)</td>
<td>600</td>
<td>$-2.5 \times 10^6$</td>
<td>50</td>
<td>3.5</td>
<td>455</td>
<td>665</td>
</tr>
<tr>
<td>V3(60°)</td>
<td>600</td>
<td>$-2.5 \times 10^6$</td>
<td>60</td>
<td>2.5</td>
<td>535</td>
<td>677</td>
</tr>
<tr>
<td>V3(70°)</td>
<td>600</td>
<td>$-2.5 \times 10^6$</td>
<td>70</td>
<td>2</td>
<td>713</td>
<td>487</td>
</tr>
<tr>
<td>V3(80°)</td>
<td>600</td>
<td>$-2.5 \times 10^6$</td>
<td>80</td>
<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>V3(90°)</td>
<td>600</td>
<td>$-2.5 \times 10^6$</td>
<td>90</td>
<td>1.5</td>
<td>810</td>
<td>629</td>
</tr>
</tbody>
</table>
5.3.4 Orphaning in a Continent-Ocean Subduction

As outlined in Chapter 4, the nature of the overriding plate in itself does not significantly alter the orphaning process. However, as can be seen from Fig. 5.4, combining a continental overriding plate to the subduction angles described in section 5.3.3 above, can result in drastically different slab dynamics. The most obvious difference is for shallow subduction angles of 20°. In the presence of a continental overriding plate, this results in a shallow flat slab (see Fig. 5.4 a) that slowly travels laterally under the continental overriding plate (see movie V3_Continent20 in Appendix B). Shallow flat slab behaviour is also observed for the steepest subduction angles of 90° (see Fig. 5.4 h). In the latter case, flat subduction is contemporaneous to slab tip penetration into the mid/lower mantle depths (see Appendix B movie V3_Continent90).

Another significant difference between oceanic and continental overriding plates is the size of the orphan (Figs 5.4 and 5.3). As discussed in Chapter 4, when the overriding plate is continental the orphan produced is considerably larger. When the overriding plate is continental, acute slab tip curvature and eventual orphan overturn occurs at subduction angles of 40° (Fig. 5.4 c). In contrast the same behaviour with an oceanic overriding plate requires subduction angles of 50° and steeper (Fig. 5.3 d). One can also observe that in contrast to ocean-ocean subduction, a continental overriding plate hinders both double sided subduction (see Fig. 5.4 g vs. Fig. 5.3 g) and shallow slab break-off for subduction angles of 70° and 80°, respectively (Fig. 5.4 g).
5.3. Results

Figure 5.4: Slab morphologies for subduction angles 20° to 90° in the presence of a continental overriding plate. The presence of a continent encourages both flat slab subduction (a and h) and steeply curved and overturned orphans (c-g). Double sided subduction and shallow slab break-off are suppressed when the overriding plate is continental. No slab orphaning is observed when a subduction angle of 90° is combined with a continental overriding plate.
5.3. Results

Similar to sections 5.3.1 and 5.3.3 above, I also compare orphan slab sizes, orphaning timescales and trench retreat for various slab angles coupled with a continental overriding plate (listed in table 5.4 below). The longest orphaning timescales are observed for subduction angles of 30° in model $V3_c(30°)$, while the shortest timescales are registered for subduction angles of 80°. Compared to their oceanic counterparts these models exhibit longer orphaning timescales overall. Nonetheless, similar to those models with an oceanic overriding plate, models $V3_c(30°)$ to $V3_c(90°)$ do not show a correlation between the orphan size and the orphaning time length. Instead, as discussed in section 5.3.3 the largest orphans are associated with the steepest subduction angles up to 80°. For subduction angles of 90°, slabs do not orphan. Instead these flatten above 660 km until they avalanche into the lower mantle.

When the overriding plate is continental, orphan sizes tend to be larger than those produced in an intra-oceanic setting for the same subduction angles (see tables 5.4 and 5.3 and Chapter 4). Furthermore, a continental overriding plate, encourages overall larger amounts of trench retreat ranging from 510 to 772 km. However, once again, no correlation is observed between the amount of trench retreat and any of the orphaning parameters described in table 5.4.

These results highlight the prevalence of slab orphaning, even at steep subduction angles. Figs. 5.4 and 5.3 show that for subduction angles of 30° or higher the steeper the angle the bigger the size of the orphan slab. These results also illustrate that even small dip changes of 10° can have a big impact on the size of the orphan slab.
Table 5.4: Orphan sizes, orphaneing timescales and associated trench retreat are compared across all orphan variants in a continent-ocean subduction.

<table>
<thead>
<tr>
<th>Model</th>
<th>Slab Strength (MPa)</th>
<th>Clapeyron Slope (PaK $^{-1}$)</th>
<th>Slab Dip Angle ($^\circ$)</th>
<th>Orphaning Time (Myrs)</th>
<th>Orphan Size (km)</th>
<th>Trench Retreat (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>V3_c(20°)</td>
<td>600</td>
<td>$-2.5 \times 10^6$</td>
<td>20</td>
<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>V3_c(30°)</td>
<td>600</td>
<td>$-2.5 \times 10^6$</td>
<td>30</td>
<td>5.5</td>
<td>656</td>
<td>772</td>
</tr>
<tr>
<td>V3_c(40°)</td>
<td>600</td>
<td>$-2.5 \times 10^6$</td>
<td>40</td>
<td>3.5</td>
<td>531</td>
<td>641</td>
</tr>
<tr>
<td>V3_c(50°)</td>
<td>600</td>
<td>$-2.5 \times 10^6$</td>
<td>50</td>
<td>3</td>
<td>768</td>
<td>594</td>
</tr>
<tr>
<td>V3_c(60°)</td>
<td>600</td>
<td>$-2.5 \times 10^6$</td>
<td>60</td>
<td>4</td>
<td>878</td>
<td>546</td>
</tr>
<tr>
<td>V3_c(70°)</td>
<td>600</td>
<td>$-2.5 \times 10^6$</td>
<td>70</td>
<td>3</td>
<td>904</td>
<td>748</td>
</tr>
<tr>
<td>V3_c(80°)</td>
<td>600</td>
<td>$-2.5 \times 10^6$</td>
<td>80</td>
<td>2</td>
<td>1012</td>
<td>510</td>
</tr>
<tr>
<td>V3_c(90°)</td>
<td>600</td>
<td>$-2.5 \times 10^6$</td>
<td>90</td>
<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
</tr>
</tbody>
</table>
5.4 Discussion

Variations in the subduction parameters described in sections 5.3.1 and 5.3.2 produce an extensive diversity of orphan sizes and orphaning timescales. This is particularly so for subduction angles, where small variations can result in drastically different orphan morphology (for e.g. see Fig. 5.3 c and d). When subduction angles are 50° and steeper, orphan slabs are considerably bigger compared to their shallower counterparts (see tables 5.3 and 5.4).

Steep subduction angles encourage a steeply dipping, vertical slab, which sinks downwards with little to no deflection. As the slab sinks, its own negative buoyancy induces the slab to assume a convex shape. Upon reaching mid-mantle depths the slab curvature increases as the resistance to its subduction also increases due to the ringwoodite to bridgmanite and ferropericlase phase transition, and the increased viscosity the lower mantle. At 1000 km depth, the slab adopts a ‘hook’-like morphology with an overturned slab tip, similar to the slab morphologies observed in previous work (e.g. Schellart, 2005; Arredondo and Billen, 2016, 2017). Further subduction results in an increase in the slab tip curvature and encourages the flattening of the overturned slab toe. Due to the increased resistance encountered by the slab tip at 1000 km depth, some component of subduction is also accommodated via increased trench retreat rates. Increasing trench retreat encourages the up-dip portion of the slab above 660 km depth to flatten. This induces viscous coupled flow with a predominantly horizontal direction in the upper mantle. On the other hand, below 1000 km depth, the considerable negative buoyancy of the slab tip/proto-orphan induces viscously coupled flow that is predominantly vertical. Similar to orphaning with shallower subduction angles, this encourages the shearing and stretching of the slab, that results from the opposing forces acting on the up-and down-dip portions of the slab (see Chapter 3 and section 5.1 above).
When the subduction angles are steeper than 40°, orphaning is also aided by the intense intra-slab deformation as the slab bends and folds over itself. The convex nature of the slab prior to orphaning produces an ‘overturned’ or ‘flipped over’ orphan slab with an orientation that is the reverse of its parent. In seismic tomography, this could be easily mistaken for the remnant of some reverse dipping, extinct subduction. Deep slab fragments with reverse polarities from that of present day subducting slab, abound at mid-mantle depths and include amongst others the Tonga, Farallon and Central American slabs. Similarly, steep subductions with angles of 30° (Turcotte and Schubert, 2014) or more are quite common and include amongst others Sumatra, Scotia, Aleutians, New Britain, India and Marianas (Schellart, 2005; Li et al., 2011; van der Meer et al., 2017). These subduction zones are littered with underlying, mysteriously oriented deep slab fragments. Taking the above results at face value, this study therefore suggests that some of the deep slab fragments observed at these steep subduction zones can also result from orphaning (see Fig. 5.5). Orphaning implies that the subduction at these locations is one long lived process as opposed to one that is constantly interrupted and renewed following tectonic reorganisation (e.g. van der Meer et al., 2017).

**Figure 5.5:** Slab orphaning with an oceanic overriding plate and steep subduction angles of 90°, compared with a tomographic cross-section for the Marianas slab, generated using the submachine portal from Hosseini et al. (2018). Orphaning in steeply subducting slabs could easily explain the overturned fragments at locations such as the Marianas, India and Cascadia, as part of one long lived subduction process.
As discussed in Chapter 3, slab orphaning does not exclude other processes such as shallow slab break-off or double-sided subduction. The latter can occur contemporaneously with orphaning as shown in Fig. 5.3 f. In this case, the overriding plate is slowly dragged downwards by the strongly coupled viscous flow that is induced by the subducting slab. Prior to its subduction, the overriding plate experiences increasing trenchward tilt (Crameri and Lithgow-Bertelloni, 2018), until eventually it sinks into the trench in conjunction with the lower plate. The two slabs are separated by the weak crustal layer on top of the lower plate, which prevents the two slabs from interacting and stops the thicker overriding plate from decapitating the lower plate. Beyond 400 km depth, the two slabs are swept together to form a thick and negatively buoyant slab. This introduces additional forcing at 660 km depth and speeds up the later stages of slab orphaning, cutting down on the separation time (see 5.3 model V3(70°)). When the overriding plate is replaced by a continent, double sided subduction is inherently discouraged due to the strong, low density, and positively buoyant nature of the overriding plate, which prevents it from subducting.

As shown in section 5.3, steep subduction angles can also induce shallow slab break-off when the overriding plate is oceanic (see Fig. 5.3 g). In this case, subduction is characterised by fast slab sinking speeds throughout the mantle. This leads to the efficient transfer of negatively buoyant and dense material to the lower mantle rather quickly, drastically increasing the slab pull force and decreasing the trench retreat. The excessive negative buoyancy of the stronger, thicker and older material forming the slab tip exerts a strong downward pull on the younger, warmer and thinner slab material close to the trench. This introduces a necking instability, which leads to slab break-off at depths shallower than 300 km (Duretz et al., 2011, 2012) (see Fig. 5.6).
Figure 5.6: Shallow slab break-off for model V3(\(80^\circ\)) with slab morphologies on the left and strain rates on the right, showing the initial vertical sinking of the slab and high strain rates at the slab tip where folding is taking place (a-b). Rapid sinking of the folded, dense slab into the lower mantle (c) results in intense intra-slab deformation at mid-mantle depths (d). The negative buoyancy of the slab eventually leads to shearing and necking instabilities in the slab at shallow depths (e-f), leading to shallow slab break-off.
5.4. Discussion

The diversity of orphan slabs is tied to the orphaning regime space and represents the dynamic balance between the negative buoyancy forces that drive subduction and those that resist it. The above results clearly show that slab orphaning, similar to other deep slab behaviour such as penetration and flattening, is viable for a range of subduction parameters. Orphan slabs persist in different tectonic settings, and a change in the nature of the overriding plate does not determine whether a slab orphans or not. Furthermore orphaning is observed for both shallow and steep subduction angles as well as strong and weak slabs. Slab strength is often considered a proxy for slab age, where thick, cold slabs are generally understood to be the oldest and warm, thin slabs interpreted as the youngest (Garel et al., 2014; Agrusta et al., 2017; Goes et al., 2017). Previous studies have attempted to draw a correlation between the slab age at the trench and slab behaviour at depth (e.g. Goes et al., 2017, and references therein). However, Jarrard (1986) shows that there is no direct correlation between slab morphologies and any one subduction parameter. This is in agreement with the results presented in this chapter, which show no correlation between slab strength and orphaning. Instead it is clear that slab morphologies are shaped by the relationship between the relative, local strength of the slab $\text{vis} - a - \text{vis}$ that of the ambient mantle. This conclusion is strengthened by the observation that slab orphaning can occur for both strong and weak slabs as long as the negative buoyancy of the slab is offset by some amount of resistance through the endothermic phase change of ringwoodite to bridgmanite and ferropericline. It is the balance between these two forces that ultimately determines whether slabs orphan, flatten or penetrate (Fig. 5.1). This point is made particularly clear for Clapeyron slope values of $-2.0 \times 10^6 \text{PaK}^{-1}$. At this Clapeyron slope, all three major slab behaviours are equally possible. In this case, whether a slab deflects, orphans or penetrates, depends on its strength. If the slab has high yield stress values, then it is strong enough to overcome the resistance to its sinking. If the slab has low yield stress values then, upon encountering the endothermic phase change, it will be forced to flatten and deflect. Intermediary slab strengths result in orphaning (c.f. Fig. 5.1). However, a change in the Clapeyron slope quickly upsets this balance.
and leads to different slab behaviours, indicating that the slab behaviour at deeper mantle depths is more complex than a simple reflection of the slab age.

It is clear beyond any shadow of doubt that orphaning morphologies are representative of an intermediary slab behaviour between deflected, flattened slabs and deep, penetrative ones. Orphaning fills the gap in the subduction regime diagram between these two end member slab behaviours. Slab orphaning also provides a possible answer to two important questions in slab dynamics; can slabs switch from one behaviour to the other? and how can this be accomplished? A slab that has split into a parent and orphan is one that has metamorphosed through all three slab morphologies. Prior to orphaning the slab adopts a penetrative morphology. It then undergoes orphaning which allows it to adjust from a steep morphology to a flattened, deflected one. Orphaning, therefore, provides one pathway through which slabs can switch modes and is available to all slabs, irrespective of their strength, their overriding plate or their subduction angle.
Chapter 6

Conclusions and Further Work

6.1 Conclusions

The aim of this thesis was to investigate the evolution and dynamics of the mid-mantle slab morphologies observed in seismic tomography. This project also sought to explore the effect of variations in ambient mantle strength on slab dynamics, and to understand how the changing slab strength with respect to that of the ambient mantle can lead to the plethora of slab morphologies observed at mid-mantle depths. To do so, I simulated the subducting slab and its interaction with the surrounding mantle using 2-D numerical models (detailed in Chapter 2). The results obtained throughout this study highlight the complex dynamics that shape slab behaviour and geometry at mid-mantle depths. In particular, this thesis discusses the effect of a continental overriding plate on deep slab dynamics and describes, for the first time, the slab orphaning process. This type of slab morphology, newly identified in this work, bridges the gap between slab deflection and penetration and fills a hitherto unexplored parameter space between these two major slab behaviours.
6.2 The Continental Overriding Plate and Deep Slab Morphologies

The nature of the overriding plate exerts considerable influence on the slab dynamics of the mantle transition zone and upper lower mantle (Chapter 3). Variations in the slab behaviour at deeper depths are observed to some extent or other for all the mantle viscosity profiles implemented in this study. However, it is particularly evident for models with viscosity profiles $V_1$ and $V_4$ whose slab evolution is uncomplicated by mid-mantle processes such as slab orphaning. Both $V_1$ and $V_4$ show slab penetration and upper lower mantle slab anchoring when the overriding lithosphere is continental. When the overriding plate is oceanic, the slab in both models deflects and flattens at 660 km depth. The markedly different slab behaviour for models exhibiting the same set-up, slab properties and mantle viscosity profiles, indicates that the forcing introduced by the continental lithosphere encourages the deeper penetration of a slab otherwise inclined to flatten. Variations in slab morphology as a result of different overriding plates can also be observed for the viscosity profile $V_2$. This model, characterised by its overall reduced resistance to subduction, (due to it missing the endothermic phase transition of ringwoodite to bridgmanite and ferropericlase), produces a ‘hooked’ slab morphology in a continent-ocean setting. In contrast, the same model produces a quasi-vertical slab when the upper oceanic plate is replaced by a continental one. However, in this case the slab behaviour at depth is shaped more by the fast subduction rates of the slab than by its overriding plate type. Similarly, the type of overriding plate exerts limited influence on the deep slab evolution of $V_3$. In this case the slab dynamics are controlled by the mid-mantle process of slab orphaning, which occurs independently and irrespective of whether subduction occurs in a continent-ocean or ocean-oceanic setting.
Slab Orphaning and Mid-Mantle Slab Break-Off

Slab orphaning is a new slab morphology identified and defined for the first time in Chapter 3 of this thesis. Orphaning introduces the concept of deep slab break-off, where the slab abandonment takes place directly at mid-mantle depths. Orphaning results from the initial subduction and penetration of the slab tip into the lower mantle. An opposing force balance in the up and down-dip portions of the slab at this depth, encouraged by mid-mantle slab weakening and opposing viscous flow directions, promotes intra-slab deformation. As the slab continues to shear and stretch the slab tip eventually breaks off and is abandoned by its parent, which flattens and deflects at 660 km depth. Orphaning, therefore, bears witness to a changing, dynamic relationship between the local slab strength at depth and that of the overall ambient mantle.

As opposed to shallow slab break-off (Duretz et al., 2011, 2012), when a slab orphans at 660 km depth, subduction continues and is accommodated through the flattened parent slab. The latter flattens and travels horizontally at 660 km depth, allowing for new material at the trench to sink into the upper mantle. Slab orphaning suggests that not all slab fragments observed in seismic tomography represent a subduction extinction event. Instead, these slab fragments could also be the orphans of a slab break-off that occurred directly at mid-mantle depths. Orphaning does not exclude shallow slab break-off, however. Indeed, the two processes can work together to shape subduction zones, with the former manifesting transition zone processes, while the latter reflects changes in the surface plate dynamics.

Slab orphaning can, however, provide a simpler explanation for some of the mysterious, deep, quasi-vertical slab fragments underlying some of the longest lived subduction zones on Earth (e.g., Farallon, Tonga, Tethys). Unlike the shallow slab break-off interpretation often adopted to explain the presence of these fragments, orphaning does not necessitate subduction cessation, complex trench jumps, subduction polarity reversals, stationary trenches or an overhaul in the overall tectonic regime. Chapter 3 suggests that orphaning can be observed at several long lived subduction zones and proposes the reinterpretation of the remnant slabs underlying
the Tonga, Central American, Japan and the Arabian slabs as orphan slabs.

6.4 Slab Orphaning Regimes

Similar to slab deflection and penetration, slab orphaning occurs for a wide range of subduction and slab parameters and is independent of the initial condition or model set-up. Slabs can orphan, irrespective of their strength, subduction angle, type of overriding plate and the Clapeyron slope value (see discussion in Chapter 5). However, the orphaning timescales and orphan sizes vary, reflecting the variations in these subduction parameters. A continental overriding plate and steep subduction angles produce bigger orphan slabs and longer orphaning timescales when compared to intra-oceanic subduction with shallower subduction angles. For subduction angles of 50° to 90°, slabs exhibit convex geometries similar to the morphologies observed in Arredondo and Billen (2017) and Billen and Arredondo (2018). The convex nature of these slabs encourages ‘turned over’ slab tips which lead to overturned orphan slabs, whose orientations are dramatically different from those of the parent slabs. Chapter 5 suggests that, in nature, slab orphaning can also be observed at deeply dipping systems such as the Cascadia, India and Marianas subduction zones.

Slab orphaning describes behaviour that falls into a hitherto unexplored parameter space between slab deflection and penetration. The orphan slab is an intermediary, transitional morphology that bridges the gap between penetration and deflection, and can provide insight into the connections between the two. Orphaning is also representative of one way a slab can switch from a penetrative mode to a deflecting one and indicates that changes in slab morphology do not always reflect surface changes.
6.5 Final Conclusions and Future work

The aim of this thesis, was to investigate the abundant diversity in slab behaviour and morphology at mid-mantle depths. In doing so, this work identifies and describes a new type of slab morphology - the orphan slab and defines its parameter space, a hitherto unexplored region between deflecting and penetrating morphologies. The role of the continental overriding plate in shaping the deeper slab morphologies of the mid-mantle was also identified for the first time in this thesis. The results presented herein illustrate some of the complex contributions that ultimately determine the mid-mantle slab morphologies observed in seismic tomography models. The outcomes of this work bring to the forefront the need for a thorough understanding of the deep slab dynamics and their interactions with the surrounding mantle. This thesis cautions against the over-interpretation of seismic tomography models, which are often used as bona fide evidence to propose slab and subduction evolutionary histories. It is of some concern that the simple observation of a slab fragment in seismic tomography is used without further considerations to infer paleo plate motions and to interpret the surface geology (e.g. van der Meer et al., 2010, 2012, 2017, amongst others).

A finer appreciation of the interrelationship between deep slab dynamics, the surrounding mantle and any additional forcing introduced by the overriding plate, can be enriched with additional constraints on the buoyancies acting on the slab. This can be obtained by adopting a similar approach to Houseman and Gubbins (1997) and measuring the buoyancy parameter of different slab morphologies. The variation of the buoyancy parameter across deflecting, penetrating and orphaning morphologies can delimit the depth varying relationship between the local slab strength and that of the overall ambient mantle. Further exploration of the parameter space for the continental overriding plate will also provide insight into forcing introduced by the continental overriding plate. Changes in the geometry, buoyancy and density of the continental overriding plate can have a significant contribution to slab dynamics at mid-mantle depths and deeper. Another important contribution to these results, and to the field of subduction dynamics in general, is the numeri-
6.5. Final Conclusions and Future work

cal implementation of a lithologically stratified and multimineralic slab and mantle. Currently the conclusions presented in this thesis are based on the behaviour of olivine and exclude the variations of other minerals with temperature and pressure changes. However, for a realistic in-depth understanding of deep slab dynamics, it is necessary to model the effects of all phase transitions for the entire slab and mantle make-up, and not just for its olivine components. Lastly, the next natural progression of this work is to expand on the insights gained from the slab-mantle interaction at depth and explore how this can influence the evolution of the surface topography, which can provide a direct constraint on the conclusions of this work.
Appendix A

Limitations, Assumptions and Evaluations

A.1 Limitations and Assumptions

A.1.1 The Numerical Simulations

Inherently, numerical models strip away the complexities of the natural world. In this thesis I chose to model subduction in a simple two dimensional cartesian box to isolate the dynamics related to one slab from the rest of the mantle convection. The internal heating rate in the model is lower than expected from the Earth’s radioactive element concentration; as a result, the overall temperature is lower than expected. Again, this choice was dictated by the desire to examine the effect of rheology on slab morphology independently from large and small scale upwellings. The lower temperatures however, hinder diffusion creep in the lower mantle but without any consequence for the model morphology. Perhaps a bigger simplification from the real mantle is that slabs are chemically differentiated and multiminerallic, which in these models are not. As a result, one would expect many phase transitions in the mantle, whose effects on slab dynamics are difficult to model properly.
A.1.2 The Vertical Viscosity Profile

The most common approach to inferring the radial viscosity structure of the mantle are inversions of post glacial rebound and the gravity field (e.g., Mitrovica and Forte, 2004; Hager and Richards, 1984; Ricard et al., 1993). Such studies have shown that viscosity increases with depth in the mantle. The depth at which the viscosity jump occurs however, remains uncertain. Most studies have prescribed the thickness and depth of viscosity layers and naturally used 660 km as the boundary between upper and lower mantle. It is from this choice that the idea stands that the viscosity increases at the upper to lower mantle boundary. However, studies in which the depth was not a priory prescribed to be at 660 km show a viscosity increase deeper in the mantle (see e.g., Vigny and Ricard, 1989; King and Masters, 1992; Kido and Čadek, 1997; Kido et al., 1998; Mitrovica and Forte, 2004; Rudolph et al., 2015). Where the inversions see the viscosity increase is a function of the seismic structure of the tomographic model used in the inversions. The novelty in Rudolph et al. (2015) is that by using a transdimensional, hierarchical Bayesian inversion the number of layers and their depths are a natural outcome of the inversions. Given the uncertainties in tomographic models and the inherent limitations of gravity inversions, the viscosity structure adopted in this thesis is no less valid than previous ones. Indeed, one could argue that it is the most parsimonious interpretation warranted by the data.
A.2 Orphaning Tests and Evaluations

To evaluate the sensitivity of slab orphaning to the numerical set-up, I show below an extensive suite of tests on the initial condition and model set-up.

A.2.1 Initial Conditions

I test the influence of a sharp initial slab kink by instead using a more naturally bent radius of 300 km (Fig. A.1). Orphaning still persists as can be seen in Fig. A.2.

Figure A.1: Subduction of a more naturally bent slab with a 300 km bending radius, replaces the initial slab kink of our standard models.
A.2. Orphaning Tests and Evaluations

Figure A.2: Orphaning of a naturally more bent slab with a bending radius of 300 km. This indicates that orphaning is not a consequence of the initial slab kink implemented in our standard models.

I also demonstrate that orphaning still occurs in a model domain with an aspect ratio of 4:1, clearly showing no influence of side boundary forcing (Fig. A.3).

Figure A.3: Increasing model width and aspect ratio, does not hinder orphaning, indicating that orphaning is not the result of side boundary forcing.
A.2.2 Resolution Tests

I increased the resolution from $\sim 4$ km to $\sim 2$ km both at the surface and at 660 km depth. In the rest of the box I refined the resolution from 11 km to 5 km. No changes in overall dynamics were visible (see Fig. A.4), indicating that the standard results are well resolved.

![Slab orphaning in ultra-high resolution](image)

**Figure A.4:** Zoomed in view of slab orphaning at ultra-high resolution clearly indicating that slab orphaning is not the product of an under-resolved model.

A.2.3 Rheological Sensitivity

The presence of a weak crustal layer atop the subducting slab is essential to reproduce Earth like single sided subduction (e.g., Crameri et al., 2012a; Čížková and Bina, 2019). I test the effects of a thinner crustal layer (5 km and 10 km respectively). For the 10 km thick crustal layer the orphan is larger than the standard case (Figs. A.6 and A.7). I observe the same result for the 5 km case if this is run at double the resolution (Fig. A.5). At the standard resolution, the 5 km weak crust is not resolved.
A.2. Orphaning Tests and Evaluations

**Figure A.5:** Slab orphanning in ultra-high resolution for a 5 km thick weak crustal layer.

**Figure A.6:** Decreasing our standard crustal thickness from 15 km to 10 km, reproduces the general dynamics. In this case however, the orphan slab larger than that those observed for a thicker crust.
Figure A.7: Slab orphaning for a crustal thickness of 15 km. Note the size of the orphan for this simulation with that orphan from Fig. A.6.
Finally I vary the depth of the weak crustal layer from the standard 400 km to 300 km and 1000 km. I also run a case with a continuous weak crust (i.e. the crustal material is never converted to mantle material unlike the other cases). Orphaning persists irrespective of the weak crust depth.

**Figure A.8:** Slab orphaning for a crustal depth of 400 km, standard for all models run and examined in this work.

**Figure A.9:** Slab orphaning for a crustal depth of 300 km. Orphaning is unaffected, despite the relatively shallow depth of the weak crustal layer.
A.2. Orphaning Tests and Evaluations

Figure A.10: Slab orphaning for a crustal depth of 1000 km. The deep crustal layer reduces the size of the orphan due to its role in reducing the coupling between slab and mantle.

Figure A.11: Slab orphaning with a continuous crust. Orphaning occurs regardless of the depth of the weak crustal layer indicating that orphaning is independent of depth of the weak crustal layer.
A.2. Orphaning Tests and Evaluations

Figure A.12: The removal of the weak crustal layer results in intense coupling between the upper and lower plates, hindering subduction.
Appendix B

Movies

Movies accompanying this thesis can be found online at the repository:

https://drive.google.com/drive/folders/1cOvBFOf_iC6iaej_MioqIUM2P8YsF1A8?usp=sharing
Bibliography


