

The influence of a sub-lithospheric decoupling layer on subduction zone surface motion, mantle circulation and slab dynamics

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Abstract

The lithosphere consists of a strong surface layer fragmented in a series of major and minor tectonic plates moving over a weaker and more buoyant layer of asthenospheric material. The boundary between the lithosphere and asthenosphere, the LAB, is a critical element of plate tectonics as it mechanically decouples the motion of strong tectonic plates from the flow in the underlying weaker mantle. Recent geophysical observations report an abrupt seismic velocity decrease and electrical conductivity increase beneath certain segments of the subducting Pacific plate, which may represent the channelised base of the subducting oceanic lithosphere. This inferred sub-lithospheric weakness is predicted to lubricate the base of subducting plates and enforce the decoupling at the LAB, with potential implications for our understanding of subduction geodynamics. However, little is known about the impact of this feature on subduction zone dynamics. This research work for the first-time compiles worldwide observations of a very thin and weak sub-lithospheric layer (SLL) embedded beneath the base of a descending oceanic lithosphere and computationally simulates SLL effects on subducting plate and mantle dynamics.

In the last few decades, the rapid progress in technology has contributed to increasing complexity of geological observations and capacity of computational infrastructures. With direct geological observations limited in both time and space scales, computational modelling offers the extraordinary opportunity to test different hypotheses for Earth's evolution through geologic time and ongoing processes, whilst promoting the development of multidisciplinary approaches. This study, therefore, employs subduction numerical models and addresses the SLL impact on slab deformation style including subduction surface motions, induced mantle flow and patterns of seismic anisotropy. Our results show that the physical properties of the sub-lithospheric layer fundamentally control the partitioning of surface motions between trench migration and horizontal plate velocities, but also the partitioning of the induced mantle flow as it splits into its poloidal and toroidal components. A regional SLL mechanism firstly provides a novel explanation for the high convergence and low trench migrations rates observed globally at natural subduction zones. Then, an SLL also delivers an alternative answer for plate margins characterised by some degree of toroidal flow and for the apparent lack of a direct

link between buoyantly driven convection (which drives only vertical and divergent motion) and the generation of toroidal energy (which corresponds to strike slip motion and plate spin). Additionally, a very thin and weak sublithospheric layer promotes the alignment of the maximum strain axes in the direction parallel to plate motion within the oceanic lithosphere and mantle directly beneath it, and in the direction parallel to the trench deeper in the sub-slab mantle. As a result, these findings contribute to the resolution of the slab anisotropy debate on whether an SLL can substantially influence three-dimensional flow around the subduction zone.

These outcomes reveal that a very weak and thin sub-layer embedded in the Earth's uppermost asthenosphere impacts the complexity of current natural subduction systems, for which little predictive and holistic modelling exists. By acting as a slippery base for the motion of the subducting lithosphere, this layer sharpens the decoupling between the strong tectonic plates and the weaker mantle underneath. This has significant consequences on slab subduction style and patterns of seismic anisotropy as observed in several subduction regions, such as Izu-Bonin-Mariana, Central America and Kamchatka-Aleutian among others, where the presence of this layer is confirmed by geophysical studies. This thesis work ultimately aims to provide a novel conceptual framework to reinterpret the geological record of complex plate-margins, with broader implications for better understanding the evolution of the Earth's dynamic system.

DECLARATION

This is to certify that:

- (i) The thesis comprises only my original work towards the PhD except where indicated in the preface;
- (ii) Due acknowledgement has been made in the text to all other material used, and
- (iii) The thesis is fewer than 100,000 words for a PhD or doctoral thesis exclusive of tables, maps, bibliographies and appendices.

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PREFACE

Sources of funding for the student’s candidature came from the Discovery Projects grant (DP150102887) from the Australian Research Council and the Melbourne Research Scholarship by the University of Melbourne. This thesis has been based around three manuscripts of original material that have been published or prepared for publication.

Roberta Carluccio was the primary researcher for all chapters in this thesis, including the planning and execution of the research and included articles. The inclusion of these publications into the thesis has been approved by the advisory committee. The primary supervisor, Rebecca Farrington, has signed the University of Melbourne Declaration for a thesis with publication. All co-authors have signed the University of Melbourne co-authors declarations form attesting that Roberta Carluccio contributed more than 50% of the work. Chapter 4 of this thesis consists of original published article material with Roberta Carluccio as the first author. Chapters 5 and 6 are written in a publication ready format, and Chapters 2 and 7 are to be finalised for literature review publication after thesis submission.

Chapter 4 is included as a published piece of work: **Carluccio, R.**, Kaus, B., Capitanio, F. A., & Moresi, L. N. (2019), “The impact of a very weak and thin upper asthenosphere on subduction motions”. *Geophysical Research Letters*, 46. <https://doi.org/10.1029/2019GL085212>. This paper was largely researched, compiled, written, and edited by Carluccio R. (i.e. contributing more than 50% of the material). The breakdown of author contributions is as follows: Kaus B., Capitanio F. A. and Moresi L. N. who assisted in designing the project and editing the manuscript with contributions from all co-authors.

Chapter 5 is based on a manuscript titled “On the Influence of a Sub-Lithospheric Layer Width on the Partitioning of the Subduction Induced Mantle Flow” that has been prepared for submission to *Earth and Planetary Science Letters* by **Carluccio R.**, Moresi, L. N., Farrington, R., Colli, L., & Capitanio F.A. . This paper was largely researched, compiled, written, and edited by Carluccio R. (i.e. contributing more than 50% of the material). The breakdown of author contributions is as follows: Moresi L. N., Farrington R., Colli L. and Capitanio who helped in designing the project and contributed with providing constructive feedback on the manuscript drafts and figures with contributions from all co-authors.

Chapter 6 has been written in a publication-ready format for *Geology* and titled “Asthenospheric flow around congested plate margins: the role of oceanic plateaux and sub- lithospheric layers” by the following authors **Carluccio R.**, Moresi, L. N., Farrington & R., Duffy B.. This paper was largely researched, compiled, written, and edited by Carluccio R. (i.e. contributing more than 50% of the material). This study contains numerical simulations, analysis and interpretations of original work performed during the PhD candidature. The breakdown of author contributions is as follows: Moresi L. N., Farrington R., B. Duffy B., who contributed to the project design and editing the manuscript with contributions from all co-authors.

Chapter 2 and **Chapter 7** represent the first draft of a literature review paper that will be wrap up after thesis submission for publication in *Geosphere* by **Carluccio R.**, Farrington, R., Moresi, L. N., and titled “Geodynamics and Seismic Anisotropies Constraints on Lithosphere and Asthenosphere Interactions at Subduction Zones: A Review”. This paper was largely researched, compiled, written, and edited by Carluccio R. (i.e. contributing more than 50% of the material). This study contains literature review and data analysis and interpretations of original work performed during the PhD candidature. The breakdown of author contributions is as follows: Moresi L. N. and Farrington R., who contributed to the project design and editing the draft with contributions from all co-authors.

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1. Chapter 1: Introduction

1.1 Overview

The Earth’s global geological circulation has driven the plate-tectonic cycles of continental drift over hundreds of millions of years and the changing pattern of continents profoundly influences the evolution of life on our planet. Underneath the Earth’s surface, the solid mantle migrates at a slow and steady rhythm as cold slabs descend and upwellings ascend, whilst Earth’s primordial heat is exploited over geologic time by mantle convective circulation. The Earth’s unique plate-tectonic style of mantle convection is unlike what is seen on any other planet. Plate tectonics relies on the concept of a rigid surface layer (the lithosphere) fragmented in a series of major and minor tectonic caps sliding over a weaker and more buoyant layer of asthenospheric material. The theory of plate tectonics has been fundamental to the study of Earth sciences for over 50 years, as it is so successful in explaining the broad picture of Earth’s evolution and has the ability to embrace many geological and geophysical observations. However, the plate tectonics model is purely a kinematic description of the manner in which events have happened and, as such, it lacks the thermal-mechanical explanation of why, as well as the capacity to predict when, particular geological events occur the way they do. The idea of the coupling between plate kinematics and mantle convection goes back more than a century ago together with the hypothesis that convective currents in the Earth’s lower mantle drive continental drift and deformation (Holmes, 1931).

The theory of mantle convection provides a physical explanation for both the direct energy supply and the mechanism responsible for plate motions, yet alone can neither explain the generation of plates nor predict their motion. The unification of the theories of plate tectonics and mantle convection remains one of the greatest scientific challenges of our century. Today, it is widely accepted that the global geological circulation is driven by negative buoyancy of the descending oceanic lithosphere that “pulls” the slab into the underlying mantle material at subduction zones (Forsyth et al., 1975). A corollary that followed the recognition of the so-called “slab pull” force as the most important of the plate forces is that plates are not only passive agents driven by mantle circulation, but they are a fundamental element of mantle circulation, and therefore mantle convection (Bercovici, 2003; Ricard, 2007; Turcotte and Schubert, 2014).

More recently, other possibilities in terms of major and minor regional force contributions have emerged. These works (e.g., Höink and Lenardic, 2010; Forte, 2011; Höink et al., 2011; Coltice et al., 2019) support the view that major drivers of plate tectonics could be slab-pull, as well as asthenosphere shear drag (or oceanic drag force) - the force acting underneath the base of an oceanic lithosphere - both being pivotal. Hence questioning the outcomes of the traditional force-balance calculations (e.g., Forsyth et al., 1975; Solomon et al., 1975; Chapple and Tullis, 1977) that identify slab pull as the main engine of plate tectonics.

The base of the lithosphere is defined as the boundary between the Earth’s strong lithospheric plates and the underlying weaker asthenospheric mantle, and as such, is a key element of plate tectonics and mantle convection that critically controls the mechanical coupling between the motion of the tectonic plates and the flow in the underlying mantle. The processes occurring at the lithosphere-asthenosphere boundary (LAB) are of fundamental importance for the better understanding of the lithosphere and mantle mutual interactions. However, it remains highly debated whether an oceanic basal drag force resists or drives plate motion. Forces associated with the basal shearing of the oceanic lithosphere in the proximity of subduction zones could propagate throughout the slab and in the mantle, thus influencing the force balance around the subducting plate, and therefore significantly contribute to the large diversity in slab deformation styles we currently observe at natural subduction zones.

1.1.1 Scientific context and relevance

A range of geophysical studies, such as Bagley et al. (2009); Kawakatsu et al. (2009); Naif et al. (2013); Stern et al. (2015); Hawley et al. (2016) among others, has recently brought considerable attention to the detailed structure of the asthenosphere along several segments of the Circum-Pacific Belt. Beneath the base of the subducting lithosphere, these works record an abrupt drop in seismic velocity and an increase in electrical resistivity across a boundary not more than ~ 30 km thick, which is compatible with a drop in viscosity of up to 4 orders of magnitude compared to the surrounding mantle material. Narrow and thin channels, that may be imperceptible to geophysical techniques, offer an overall more plausible explanation for some of the enigmatic features and elusive nature of the LAB (Rychert, 2015). Despite these observations indicating the presence of a thin and weak sub-lithospheric layer (SLL), to date, the geodynamic impact of this layer on subduction dynamics and large-scale mantle flow has remained largely unexplored.

In recent decades rapid progress in technology has offered new and exceptional possibilities for the development of computational approaches where numerical modelling has become an essential tool in the geosciences. With direct observations being limited across both time and space scales, numerical modelling provides the capacity to test different hypotheses and generate insights into the Earth’s evolution and present-day geological processes.

Geodynamic modelling of SLL’s impact on subduction dynamics offers a unique opportunity to shed light on its role in subduction dynamics, which has remained elusive to date. In the literature, structural and compositional heterogeneities have been invoked to explain the often complex tectonic evolution of convergent margins and inferred miscellaneous patterns of mantle flow – as such this thesis work represents a natural progression towards a better understanding into the dynamics of plate tectonics and the interaction between the lithosphere and mantle. Recent advances in computational power, as well as

the hard work of brilliant scientists, have made the modelling of a subduction zone possible and uncovered broad correlations between subduction zone observations and inferred parameters. While significant progress has been made, the modelling of a subduction zone has not yet resulted in a clear trend for slab and trench migration style, convergence velocity, and patterns of seismic anisotropy and mantle flow around the subduction zone – especially in the case of slow retreating or advancing trenches in natural settings (e.g., Garfunkel et al., 1986; Jarrard, 1986; King, 2001; Heuret and Lallemand, 2005; Lallemand et al., 2005; Sdrolias and Müller, 2006; Capitanio et al., 2007).

This thesis firstly aims to bridge a literature gap by providing insights into the evolution of natural convergent margins where trench migration rates are significantly slower than plate convergence rates. Secondly, it provides improved constraints on where SLL have been reported to exist. Thirdly, it highlights the implications of what a “slippery base” of a tectonic plate could mean for the decoupling of lithospheric plate motion from the mantle flow underneath, and its main driving forces. In particular, this work demonstrates through the use of subduction numerical modelling how very thin and weak layers embedded in the Earth’s deep interior can influence the complexity we currently observe in natural subduction systems. Outcomes presented herein show that the presence of this layer sharpens the decoupling between the tectonic plates and the mantle beneath; hence substantially contributing to plate motion and deformation, and patterns of mantle flow and seismic anisotropy around subduction zones. This research work, therefore, shows that an SLL has a significant wider range of effects than previously assumed (Becker, 2017) and reveals the importance of mantle tangential forces in promoting surface deformation. The work carried out throughout this thesis adopts the geological constraints provided by the aforementioned geophysical studies and well match natural observations of surface motions with relatively simpler numerical models. As such, these outcomes provide a novel conceptual framework for reinterpreting the geological record of complex plate-margins for which little predictive and holistic modelling exists. Ultimately, the models described herein provide an important step towards the development of a self-consistent characterisation of the lithosphere and mantle dynamic systems and their mutual relationships with broader implications in terms of understanding the secular evolution of the planet Earth.

1.1.2 Research aims

To address the impact of a sub-lithospheric layer on subduction dynamics, this thesis is divided into three interlinked objectives:

1. an assessment of the partitioning of the subduction zone surface motions between the motion of the plate and the migration of the trench;

2. an investigation of how the deformation flow induced in the mantle by the sinking slab is split between the vertical and rotational components of the three-dimensional velocity field; and
3. a final research module, which is further divided into two sub-aims:
 - (a) a study of the connection among slab and trench deformation style, patterns of mantle flow and sub-slab seismic anisotropy; and
 - (b) a compilation of subduction zone observations of sub-lithospheric channels, plate and trench kinematics, slab ages, dips and morphologies, and shear wave splitting measurements within the Pacific plate.

1.1.3 Structure of the thesis

The remainder of this thesis is structured to address the aforementioned research questions presented in §1.1.2, in the following way:

Chapter 2 provides general background information about mantle convection and plate kinematics, and a review of subduction dynamics and seismic anisotropy studies from a numerical modelling and mantle convection perspective, whilst detailing the gaps in the literature to be addressed throughout the thesis.

Chapter 3 describes the basic principle for the numerical modelling of a subduction zone. It starts by introducing the fluid dynamic conservative laws, material behaviour constitutive relationships, the boundary conditions tested and the computational framework used to perform the experiments presented in this thesis.

Chapter 4 uses 2D mechanical and thermo-mechanical subduction numerical models to address the impact of an SLL on the partitioning of subduction surface motions.

Chapter 5 employs 3D subduction numerical models to study the influence of an SLL width on the partition of subduction induced mantle flow, surface motions and the relationship between them.

Chapter 6 incorporates knowledge and models from the work accomplished in chapters 4 and 5 and builds upon them. For the first time this study: i) compiles a range of geophysical studies suggesting the presence of a very thin and weak decoupling layer at the base of the sinking oceanic plate; ii) reports a striking spatial coincidence featuring the presence of SLL combined with an anomalously buoyant layer of crustal material; and iii) develops large-scale three-dimensional numerical models to address the impact of the above factors, in isolation and joined together, on patterns of seismic anisotropy around the subduction zone.

Chapter 7 first summarises the results of each research module and then discusses in greater detail the scientific contribution of each of them. Subsequently, conclusions are drawn upon the relevance of the research findings of this thesis, whilst presenting their

limitations, future research directions and the broader implications for the modelling of slab and mantle dynamics during subduction.

Chapter 8 marks the end of the thesis and includes a summary that addresses the initial research aims and highlights the broader scientific relevance of this thesis work.

2. Chapter 2: Literature review

2.1 Background

Global geological circulation is driven by the negative buoyancy of oceanic plates subducting into lighter and weaker mantle material underneath them at convergent margins. Coupling between the descending slab and mantle, through the slab pull force and viscous drag, significantly affects the long-term chemical and thermal balance that self-sustains mantle circulation. At the surface, subduction of the oceanic lithosphere into the mantle results in the motion and deformation of tectonic plates and deep oceanic trenches. Patterns of trench migration influence lithospheric stress, mixing, heat flow, slab penetration, velocity and pressure fields partitioning with, therefore, important implications for surface tectonics as well as mantle convection. Most contemporary subducting slabs appear to retreat and stagnate in the upper mantle while a smaller number shows reduced trench retreat rates with either flatten or steepen slab morphologies, which may be accompanied by slab penetration in the upper lower-mantle transition zone (e.g., Schellart, 2008a; Goes et al., 2017).

Recent reviews of observational constraints and geodynamic models (e.g., Coltice et al., 2017; Goes et al., 2017) highlight that subduction zone modelling has not yet resolved clear trends in the observations for present-day subduction styles (i.e., slab morphology and dip), patterns of seismic anisotropy, toroidal to poloidal deformation field ratio, convergence and trench migration rates, particularly for slowly retreating or advancing trenches (e.g., Garfunkel et al., 1986; Jarrard, 1986; King, 2001; Heuret and Lallemand, 2005; Lallemand et al., 2005; Sdrolias and Müller, 2006; Capitanio et al., 2007). Although some correlations exist, such as between absolute plate convergence and age, and trench history and plate morphology (e.g., van der Hilst and Seno, 1993; Faccenna et al., 2001), it is possible that other forces could significantly contribute to the large diversity we currently observe in deformation mechanisms at subduction zones. The lithosphere-asthenosphere boundary (LAB) has long been known to critically control mechanical coupling between the motion of the strong tectonic plates and the flow in the weaker underlying mantle (Anderson, 1995). At this boundary, forces associated with the shearing of the oceanic lithosphere in the proximity of a subduction zone could propagate in the mantle, hence influencing the force balance around the subducting plate. Shear heating at the LAB due to high strain, caused by the discordant motion of the overlying oceanic plate relative to the mantle underneath and/or a plume-fed lubricated base of the lithosphere, could both induce significant weakening. Such weakening is a viscosity reduction that can be enhanced by the presence of melt, hydrated and/or volatiles phases and may result in the presence of one or multiple thin layers of entrained sub-lithospheric material (SLL) weaker than the surrounding mantle ponding at the base of the subducting lithosphere. This could play an important role in decoupling the motion of tectonic plates from the flow in the underlying mantle, promoting or opposing subduction motions with respect to

the mantle underneath. It has been debated (e.g., Long, 2013; Rychert, 2015; Stern et al., 2015) whether such sub-lithospheric inferred weaknesses could be significant enough to affect i) the subduction style and surface motions; ii) the partitioning of the background mantle flow; and iii) patterns of seismic anisotropy around the subduction zone, yet SLL impact has not been addressed.

Over the past 20 years, many review papers (e.g., King, 2007; Bercovici et al., 2015) have shed light on our understanding of the mantle-lithosphere dynamic leading to a substantial improvement in our knowledge. It is subsequently possible to identify a subset of lithosphere-aesthenosphere parameters which exert stronger control on subduction dynamics and, thus, substantially contribute to most of the complexity we observe in the geological record. For example in the field of computational geodynamics some of these include plate age, lithosphere and mantle viscosity contrasts and rheology (Bellahsen et al., 2005; Stegman et al., 2006; Garel et al., 2014), among others. In this literature review, I will show that when evaluated together, many fundamental questions into lithosphere and mantle interactions, can be answered. The consolidation of these insights will provide the literature base for the rationale of thesis work while further unravelling factors responsible for changes in surface motions, plate boundaries, patterns of mantle flow of current, and possibly ancient, tectonics margins.

In the first part of this literature review, section 2.2, I present the basic physical concepts of mantle convection, net rotation, plate kinematics and major plate forces resisting or enhancing surface motion adopted by the geodynamics community from a global mantle convection perspective. In section 2.3, I provide a subduction zone definition, followed by a description of observational constraints of slab inferred parameters from seismic tomography and slab seismicity, a present-day record of plate motion and trench mobility. I then summarise recent computational geodynamic literature of large-scale subduction system modelling. In section 2.4, concepts and definitions of seismic anisotropy and methods are covered, the link between seismic anisotropy and mineral physics constraints is explored, and patterns and models for the origin of seismic anisotropy in the wedge, slab and sub-slab are described. Part 2.5 concludes this literature review by providing an overview of the complex nature of the LAB and a 30-year summary of studies reporting a channelised base of a subducting plate.

2.2 Plate motions and dynamics

This first section provides an overview of plate motions and dynamics with a particular focus on modelling studies that improve our understanding of large-scale plate motions and subduction processes underpinning surface motions. I first look into the definition of mantle convection, then present surface kinematics as an initial basis for the better understanding of mantle and lithosphere mutual interactions. I subsequently analyse the

net rotation of the lithosphere, present-day plate and trench motions and dynamics, whilst discussing their strengths and weaknesses. Consequently, implications are drawn in terms of how to employ these observations to better understand lithosphere-mantle dynamics and other relevant forces. I then investigate the link among inferred slab parameters, convergence and trench migration rates. Finally, I present how subduction modelling can illustrate the role of subducting slabs and mantle structural heterogeneities in shaping Earth’s surface motions, slab dynamics and mantle flow.

2.2.1 Mantle convection

Global mantle circulation drives plate-tectonic cycles of continental drift over geologic time where the changing pattern of continents profoundly affects the evolution of our planet. The plate tectonics model has been fundamental in the study of surface deformation as it has the ability to encompass geological and geophysical observations, such as the distribution of both earthquakes and volcanic disasters. The theory that is so successful in explaining the broad picture of Earth’s evolution cannot although explain the details of why particular geological events occur as they do because it is fundamentally a descriptive model and lacks the thermal-mechanical explanation of why. This is because plate tectonics is purely a kinematical model that does not outline the essential force and energy source for plate motions. A dynamic theory of plate tectonics that describes the essential force balance and energy source/sink does not yet exist.

The theory of mantle convection was originally developed to understand the thermal history of the Earth and to provide a driving mechanism for Alfred Wegener’s theory of Continental Drift in the 1930s (Wegener, 1966; Schubert et al., 2001; Bercovici, 2007). While it was originally criticised and discredited, the subsequent discovery of sea-floor spreading (Hess, 1962; Vine and Matthews, 1963) ultimately led to the unifying theory of plate tectonics in the 1960s and revived interest in mantle convection as the driving agent of surface motion and deeper circulation (Runcorn, 1962a,b; Morgan, 1971). The term mantle convection refers to thermal convective circulation occurring within the Earth driven by the rising of hot (i.e., plumes) and sinking of cold (i.e., cold slabs) materials. Since the recognition of the importance of slab pull as the most important of the plate forces (Forsyth et al., 1975) and subducting slabs as basically cold downwellings (see also section 2.2.4), it has been generally accepted that plates are an integral part of mantle circulation, and not only passive agent simply driven by mantle convection (Bercovici, 2003; Ricard, 2007; Turcotte and Schubert, 2014).

Mantle flow, thus, governs plate tectonics, volcanic activity, as well as chemical segregation and cooling of the solid Earth (Bercovici, 2011). The theory of mantle convection provides a physical explanation for both the force balance and energy source responsible for deep Earth mantle dynamics, however simple fluid convection cannot yet explain the generation and self-organization of plates or predict their motion (Bercovici, 2011;

Coltice et al., 2017). To date, plate motion is described by the plate tectonic model while the dynamics happening in the Earth’s interior are physically explained by the theory of mantle convection. These theories, individually, fail to predict the specific time or location of an earthquake or a volcanic disaster. While discussion of the coupling between plate tectonics and mantle convection dates back more than a century (Holmes, 1931), a unification of the two in a comprehensive physical theory for the origin, motion and deformation of tectonics plates remains today one of the greatest challenges of this century (e.g., Bercovici, 2003; Ricard, 2007; Bercovici, 2011; Bercovici et al., 2015).

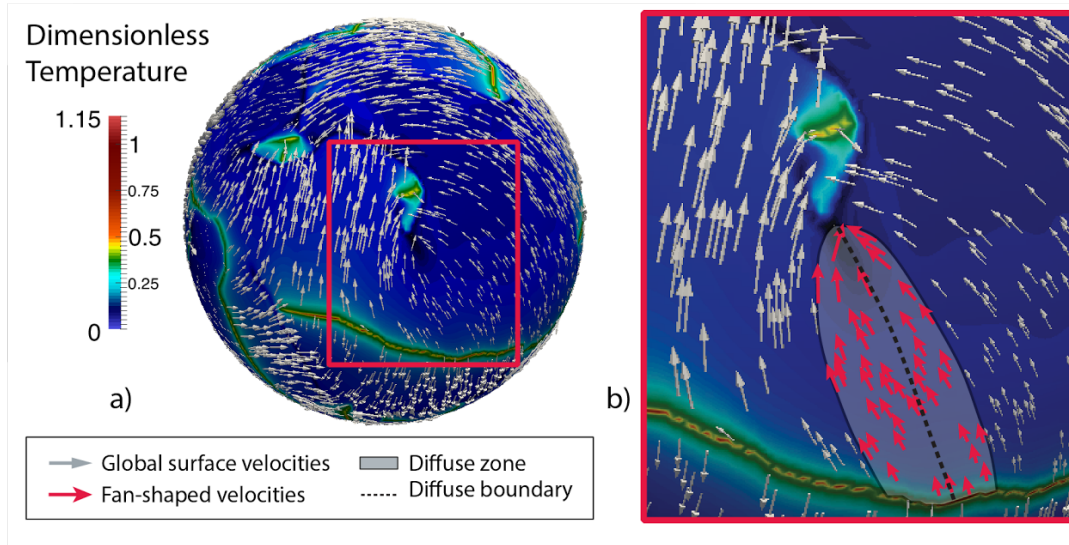


Figure 2.1: Geodynamic spherical model showing global temperature and surface velocities, taken from (Mallard et al., 2016). In a) subduction and mid-ocean-ridge regions emerging dynamically are represented by dark and light zones, respectively. In b) a zoom-in of a diffusive boundary, the red fan-shaped velocities indicate an intra-plate diffusive zone. Plate geometries in the model are built using the Gplate software (Williams et al., 2012) and through the use of synthetic sea-floor age and temperature data.

In recent decades numerical modelling has become an essential tool in geoscience with rapid progress in technology continuously offering new opportunities and multidisciplinary approaches. Geodynamicists have developed advanced models with numerical techniques and physics for simultaneously modelling surface tectonics and mantle convection featuring a self-organised evolving system (e.g., Mallard et al., 2016; Coltice et al., 2017); hence, improving our understanding of the feedback mechanisms between surface motion and convection. Mantle convection modelling has more recently progressed to the point where connections with surface motions can be made leading to the hypothesis that plate motions could be emergent features of the global dynamics of a unified single mantle-lithosphere system. Many authors (e.g., Cramer and Tackley, 2014; Mallard et al., 2016; Coltice et al., 2017) have now employed self-consistent spherical models showing a promising level of realism for the emergence of plate-like behaviours (e.g., figure 2.1). In these

calculations, tectonic plate fragmentation and mantle flow are enabled through the use of considerable lateral variations in viscosity, rheology and geometry. However, these models define the organisation and structure of either the lithosphere or the mantle as initial conditions, meaning that major challenges still remain in reproducing a global self-organised system able to emerge spontaneously from a fully dynamic modelling approach (Coltice et al., 2017).

In this context, the work presented in this thesis aims to show that regional and/or local viscosity reductions focused in very thin asthenospheric mantle layers embedded beneath the subducting lithosphere produce a better match to observations of natural global mantle flow and surface patterns of plate motions. Thus providing a new geodynamic framework for generating holistic models to help interpreting subduction zone processes.

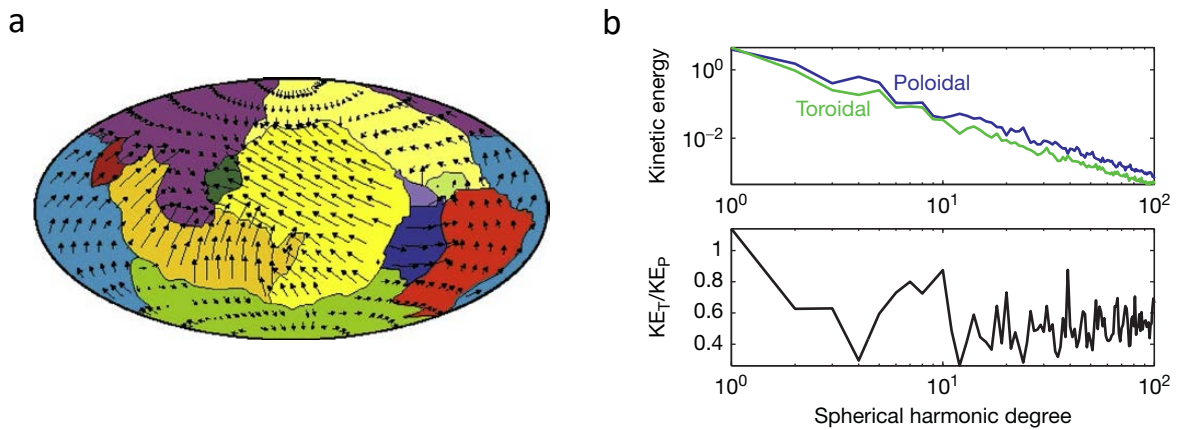


Figure 2.2: In a, an equal-area projection map of contemporary major tectonics plates and their direction of motion. The direction of motion of plates shows amounts of strike-slip motion (i.e., plate moving normal to each-other) similar to divergent and convergent deformations. In b, plots showing (top to bottom) power spectrum of present-day kinetic energy of poloidal and toroidal fields and their energy ratio versus spherical harmonic degree. Toroidal and poloidal potential appear comparable at most wavelengths (top plot), and their ratio, ≈ 0.7 , indicates that both flow fields are substantially active and representative of contemporary style of plate motion (bottom plot). Figure taken from Bercovici et al. (2015).

2.2.2 Plate kinematics

The solid-body rotation of a system of different sized spherical caps around an Euler pole well defines the relative surface motion of tectonic plates boundaries (i.e., whether they are divergent, convergent or in shear) and gives an accurate prediction of their deformation. The mathematical working model of plate tectonics relies on the above underlying principle of the Euler pole theory and continues to be used in present-day plate surface

models (Bercovici et al., 2015). A geodynamic conundrum of this description is that solid plate motions appears uniquely driven by the relative motion of plates along narrow boundaries, but also that observations report largely uncorrelated Euler poles (e.g., directions of motion (figure 2.2)). This results in a system of narrow plate boundaries where there is as much divergent and convergent motion as strike-slip deformations, which represents an important unresolved problem of the plate-tectonic style of mantle convection strictly related to the decomposition of the three-dimensional velocity field (Bercovici et al., 2015).

The contemporary velocity field of the solid Earth can be decomposed in terms of poloidal, $\vec{v}_p$, and toroidal components, $\vec{v}_t$ (Bercovici et al., 2015). A schematic representation of the poloidal and toroidal components of the three-dimensional velocity field can be observed in figure 2.3. Poloidal flow corresponds to the rise and fall of upwellings and downwellings and is simply the convective mantle circulation associated with vertical mass transport driven directly by thermal buoyancy ($\vec{\Delta} \times \vec{\Delta}_p = \vec{0}$). The horizontal divergent is associated with the poloidal flow and measures convergent and spreading surface motions (Dumoulin et al., 1998). Toroidal flow ($\vec{\Delta} \cdot \vec{v}_t = 0$) corresponds to vortex-like flow and rigid body rotation around a vertical axis. It expresses itself primarily as strike-slip shear motion and transform faults, and secondarily as plate spin. The vertical vorticity measures angular velocity and horizontal shear associated with toroidal flow.

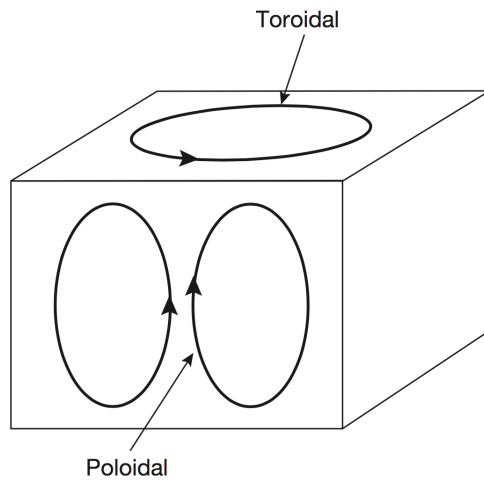


Figure 2.3: Schematic representation of flow lines illustrating toroidal and poloidal motions, taken from Bercovici (2003).

The present-day decomposition of the velocity field (figure 2.2 and 2.4) shows as much poloidal as toroidal kinetic energy comparable at most wavelengths (e.g., Hager and O’Connell, 1978; Forte and Peltier, 1987; O’Connell et al., 1991; Lithgow-Bertelloni et al., 1993; Dumoulin et al., 1998). Thus, toroidal motion is very significant for the Earth’s plate-tectonic style of convection. However, toroidal motion does not exist in convection

with constant or depth-dependent viscosity and is only weakly present with temperature-dependent viscosity (Christensen and Harder, 1991; Tackley, 2000). Hence, there is no direct link between the slab pull force driving buoyant convection (which drives only vertical and divergent motion) and the generation of toroidal energy. The generation of toroidal flow involves an additional flow field that would not otherwise exist and requires the coupling of convective motion with nonlinear rheological effects which only occur if the viscosity varies laterally (Bercovici et al., 2015). The challenge of how to generate buoyancy-driven poloidal motion to induce toroidal flow, either through horizontal viscosity variations or through boundary conditions (i.e., imposing plates, faults or strain weakening effects at plate margins) constitutes “the poloidal–toroidal coupling problem” (Bercovici et al., 2015).

Moreover, the average toroidal motion currently observed on Earth depends on the reference frame adopted and changes during the geologic time (Čadek and Ricard, 1992; Lithgow-Bertelloni et al., 1993; Tackley, 2000). Nonetheless, no plate margin is uniquely poloidal or toroidal, but it usually displays some combinations of the two where both flow fields are equally and very strongly localized with comparable peak values (figure 2.4). The present-day toroidal to poloidal ratio (i.e., the root-mean-square net vorticity to that of the divergence) is about 0.7 for most lithosphere reference frames (Bercovici et al., 2015). In a no-net rotation reference frame the toroidal component is generally smaller and the amplitude ratio is known to have varied between 0.25 and 0.5° over the last 120 million years (Olson and Bercovici, 1991; Lithgow-Bertelloni et al., 1993; Tackley, 2000). The full understanding of the mechanism capable of generating substantial laterally varying viscosity to trigger and self-sustained plate-like toroidal motion is perhaps the most critical issue in understanding the generation of Earth’s plate-tectonic style of mantle convection and is at the heart of the envisioned unified theory of mantle convection and plate tectonics. In the research chapter 5 of this thesis, I will be presenting a novel and alternative mechanism for the development of toroidal flow in a cartesian subduction model triggered by the subduction of a thin sub-lithospheric layer underlying the oceanic slab in the mantle beneath.

2.2.3 Net rotation

The net rotation of the lithosphere equates to the net spin that results from the differential motion between the lithosphere and the underlying mantle. The net rotation represents an integral part of the toroidal flow field and emerges from lateral viscosity variations within the lithosphere and mantle, such as heterogeneous cratonic keels, asthenosphere distribution and subducting slabs, as well as asymmetric arrangement of plates (Harper, 1986; Ricard et al., 1993; G  rault et al., 2012; Rudolph and Zhong, 2014). In a deep mantle reference frame, the toroidal to poloidal ratio approaches one indicating the significant

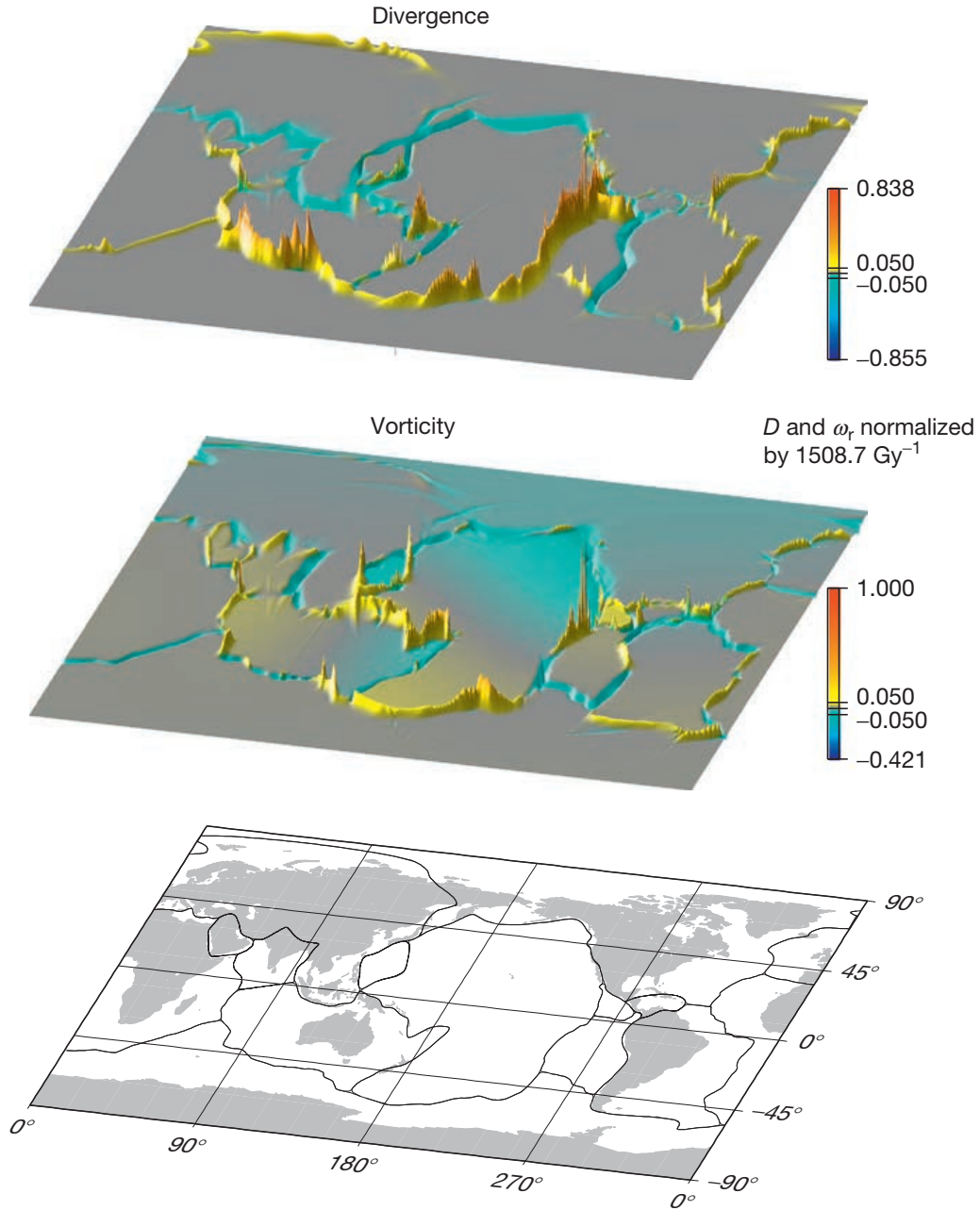


Figure 2.4: Estimation of surface divergence and vorticity of Earth's deformation field, and current plate boundaries (from top to bottom), taken from Bercovici et al. (2015) and calculated from the earlier work of Dumoulin et al. (1998).

role of lateral viscosity variations in the mantle on plate tectonics (Bercovici et al., 2000; Coltice et al., 2017).

Some authors (e.g., G  rault et al., 2012; Becker, 2017) claim that the controlling factors on the amplitude of net rotation and trench motions are slab properties, ridge position, amount of plate coupling and that, other lateral variations of viscosity, for instance, a locally weaker asthenosphere or a stiff continental keels have only limited effects on the net rotation, but are still able to influence trench migrations. G  rault et al.

(2012) results show that when the mantle is weaker underneath oceanic plates than in continents, it produces trench motion mechanisms and continental velocities more in line with observations.

In general, different approaches have been adopted to constrain the Earth’s reference frame and, thus, plate and trench motions; however, all derived “absolute” plate motion models show discrepancies in the amounts of lithospheric net rotation and trench migration across recent geological time. One approach to constrain Earth’s absolute reference frame is to constrain the position of the Earth’s spin axis using paleomagnetic data. Another is to consider collections of plumes as steady features relative to the whole mantle. Other common methods include the use as a reference frame of either a plate that would remain fixed over time, a large shear velocity province at the base of the lower mantle, identified slab remnants from seismic images, or azimuthal anisotropy in the upper mantle which would remain fixed over a long period of time.

The estimation and validity of these reference frames rely on different geodynamic assumptions that are debated, have a great number of uncertainties or come with the difficulty of going too far back in time (Coltice et al., 2017). For example, when using one of the hotspot reference frames one needs to correct for their motions relative to each other as these are indeed mobile. Moreover, beyond the Cenozoic, there is insufficient seafloor preserved to reconstruct plate motions and paleomagnetic data need to be combined with other techniques for old reconstructions. However, some amount of net rotation is predicted by all absolute reference frames. The hotspot reference frames locate the current pole in the southern Indian Ocean which has remained stable for the last 30 Myr (Coltice et al., 2017). Yet, differences arise among various absolute reference frames in terms of amplitude (Becker and Faccenna, 2009; Torsvik et al., 2010; Becker et al., 2015; Coltice et al., 2017). An alternative to these reference frames is the no-net reference frame. In this case, the total toroidal energy is only about 0.28 of the total poloidal kinetic energy. A recent trend of global plate motion studies in geodynamic modelling (e.g., Tackley, 2000; Torsvik et al., 2010; Alisic et al., 2012; Domeier and Torsvik, 2014; Torsvik et al., 2014; Williams et al., 2015; Coltice et al., 2017) is to adopt the lowest end member value for the amplitude of net rotation $\approx 0.13^\circ/\text{Myr}$ (Torsvik et al., 2010), which is based on the assumption that the net rotation should be as little as possible to fit with numerical modelling outcomes. The same assumption is also used to test the reliability of plate motions average values and rheological parameters (Alisic et al., 2012; Williams et al., 2015). However, larger net rotation amplitude values up to $0.2\text{--}0.3^\circ/\text{Myr}$ are suggested by azimuthal anisotropy measurements (e.g., Becker et al., 2008, 2014) and shear wave splitting studies (e.g., Kreemer, 2009; Conrad and Behn, 2010).

The estimation of a properly fixed reference to measure the net rotation of the lithosphere with respect to the mantle is another one of the most fundamental aspects of plate kinematics that still remains highly enigmatic. A reliable estimation of the net rotation is

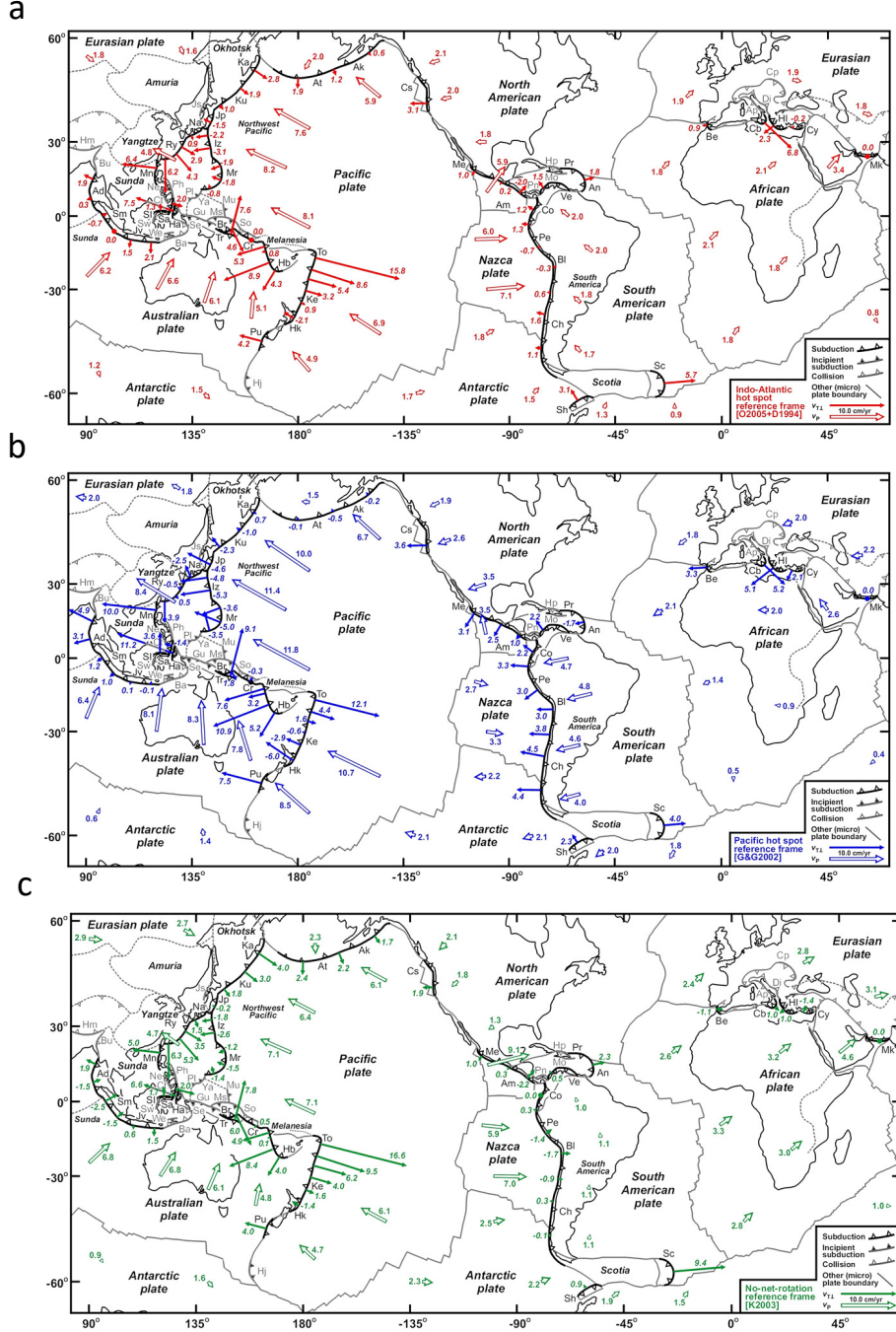


Figure 2.5: Subduction surface motions for several natural subduction regions taken from Schellart et al. (2008). Plate horizontal velocities and trench-normal migration rates are given for (in a) the Indo-Atlantic hotspot reference frame (O'Neill et al., 2005); (in b) the Pacific hotspot reference frame (Gripp and Gordon, 2002); and (in c) the no-net-rotation reference frame (Kreemer et al., 2003), a and b use the relative plate motion model of DeMets et al. (1994).

critical for determining a suitable absolute reference frame for surface motions, decomposition of the velocity field and for the full understanding of the interplay between mantle convection and plate tectonics (Coltice et al., 2017). However, a consistent outcome across

different reference frames is that contemporary absolute plate velocities “correlate negatively with the total continental area” and “positively with the fraction of plate boundary being subducted” (Minster et al., 1974; Forsyth et al., 1975), which is at the base of the recognition of slab pull as the most important force acting on a plate (figure 2.6).

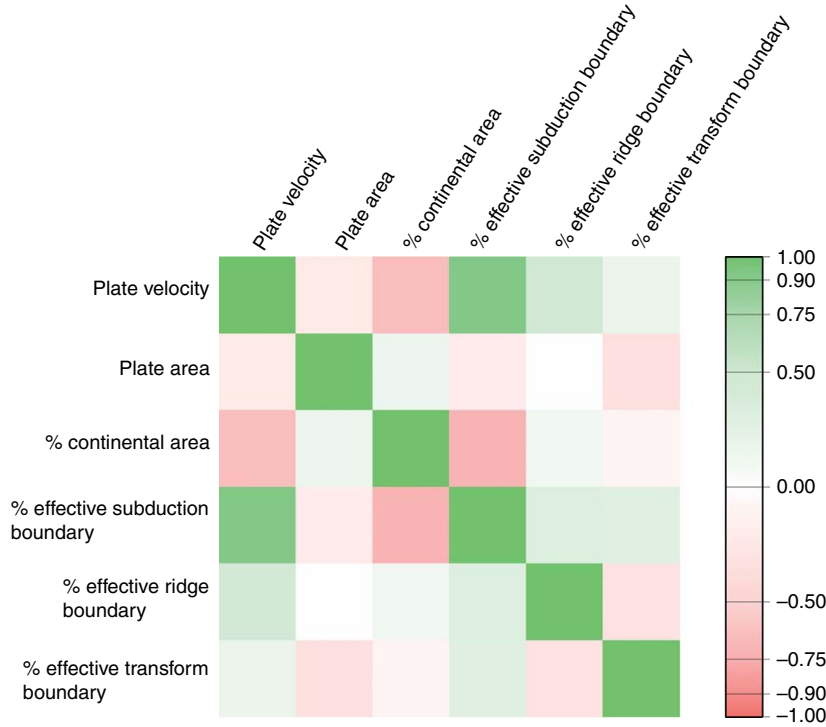


Figure 2.6: Covariance matrix highlighting the relationships between plates observations and characteristics, taken from King (2007) and originally from Forsyth et al. (1975). Covariance matrix in colour shading where green and red indicate a poor and strong correlation, respectively.

2.2.4 Plate dynamics

The Euler pole model of plate tectonics is a kinematic representation of motion, but not causative since it doesn’t describe the engine mechanisms responsible for plate motions, (Bercovici, 2003). An important challenge in geodynamics is to identify the physical mechanism by which vertical buoyancy forces in the mantle ultimately results in horizontal forces acting on and driving the plates, which represents a central assert in developing realistic plate models (as explained in §2.2.1). Several forces act on plates which can either promote or impede their relative motion. Combinations of these forces are responsible for surface plate movements. The velocity of present-day motion appears constant, indicating a state of dynamic equilibrium where a balance exists between driving and resisting forces. Classical investigations (e.g., Forsyth et al., 1975; Solomon et al., 1975; Chapple and Tullis,

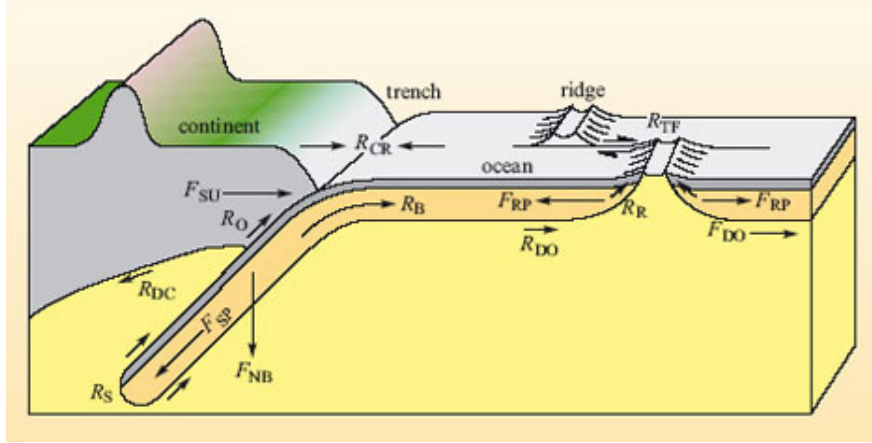


Figure 2.7: A schematic drawing illustrating the major forces acting at plate margins, taken from Rogers and Blake (2008).

1977) infer the Earth’s force budget by calculating the torque balance on each plate from a set of force parameters that when balanced results in appropriate plate velocities.

The outcome of these initial models revealed that the negative buoyancy of slab is the most efficient driver of natural surface motions (e.g., McKenzie, 1969; Elsasser, 1971; Richter, 1973). One of the most classical force-balance demonstrations was carried out in the seminal paper of Forsyth et al. (1975). Forsyth et al. observed that a plate velocity strongly correlates with the connectivity of a plate to a slab, which is the trench length, whilst it does not correlate with its ridge length (figure 2.6). From these observations followed the recognition of the slab pull force, which is the negative component of the buoyancy force, as the most important of the plate forces (Forsyth et al., 1975). From a dynamics perspective, plates then become not only driven by mantle convection but are an integral physical part of the geological circulation, and therefore mantle convection (Bercovici, 2003; Ricard, 2007; Turcotte and Schubert, 2014).

After the recognition of slab pull as the driver of plate motions followed a set of subsequent studies (e.g., Lithgow-Bertelloni et al., 1993; Lithgow-Bertelloni and Richards, 1998; Conrad and Hager, 1999; Becker and O’Connell, 2001) revealed the importance of mantle viscous dissipation in resisting plate motion. Viscous dissipation can occur at the base of and within the lithosphere, at each side of the subducting slab and along thrust faults at collision zones. Figure 2.7 illustrates a simplified overview of plates and mantle forces. The relative contribution of these forces to lithosphere motion can be explored by considering forces acting at i) plate margins and ii) the base of the lithosphere.

The most important plate forces are thought to act at their boundaries (Rogers and Blake, 2008). At destructive plate boundaries, the so-called “slab-pull” force pulls the whole oceanic plate down as a result of the negative buoyancy of the slab. The downward-acting component of the gravity-generated force is transmitted through plate bending of

the slab and triggers sinking of oceanic lithosphere into the mantle underneath. The magnitude of the slab-pull force is related to the dip angle at which the slab descends into the mantle ($F_{SP} = F_S \sin \theta$). The orthogonal complement to F_{SP} is the slab-normal force, F_{SN} ($F_{SN} = F_S \cos \theta$) originally not considered in Forsyth et al. (1975). As slabs descend into the mantle, they encounter resistance in the opposite direction due amounts of viscous shear stresses acting on the sides of the descending subducting slab and viscous mantle being displaced during subduction. This viscous resistance to slab motion is known as the slab resistance force, R_S . A further impediment to plate downward movement is given by the bending resistance, R_B . This force acts when plates need to flex, and then, bend at the trench before they begin to slide beneath the upper plate. Modelling studies (e.g., Elsasser, 1971; Funiciello et al., 2008; Schellart et al., 2008) have shown that the balance between this force and the opposing normal stress in the mantle contributes to trench retrograde motion and slab rollback.

At constructive boundaries (i.e., middle ocean ridge) acts the ridge-push force (F_{RP}). This corresponds to a divergent sub-horizontal force due to the upwelling of hot mantle material and consequent dynamic pressure field generating at mid-ocean ridges. The ridge resistance is the counter-force resting plate motion at these localities (R_R). An additional retarding force is the transform resistance force, R_{TF} . R_{TF} represents frictional stresses acting on a segment with significant tangential plate-motion component corresponding to strike-slip shear motion and transform faults deformation.

Other forces acting at plate boundaries are: i) the trench suction force F_{SU} which is an analogous of the ocean driving forces for the overriding plate. F_{SU} generates from convection induced mantle flow from above the subducting slab and can pull the upper plate toward the trench; ii) the overriding plate resistance, R_O opposing the pull of the slab against its walls; and iii) collisional resistance force R_{CR} acting on the upper surface of the slab as it is pulled down into the mantle. This force is usually prevalent in subduction zones with shallower dips, such as young oceanic plates with a no well-developed dip angle undergoing substantial viscous friction (Wilson, 1993; Rogers and Blake, 2008).

The oceanic drag is one of the forces acting in the mantle and particularly underneath the base of an oceanic plate F_{DO} . The convecting mantle drives plate motion when the underlying asthenosphere moves in the same direction of the above lithosphere. Conversely, when the asthenosphere moves slower or it moves in the opposite direction of an oceanic lithosphere then the oceanic drag becomes a resistant force, R_{DO} , and thus hampers plate motion. The analogue of the resistant force for the continental lithosphere is the continental resistance drag force, R_{DC} . R_{DC} , is due to the fact that continents generally have a “keel” and, therefore, larger resistance to movement than oceanic plates. The resulting bulk basal drag force is the sum of the oceanic drag and continental resistance drag forces, such as $F_{DO} + R_{DC}$ (Rogers and Blake, 2008).

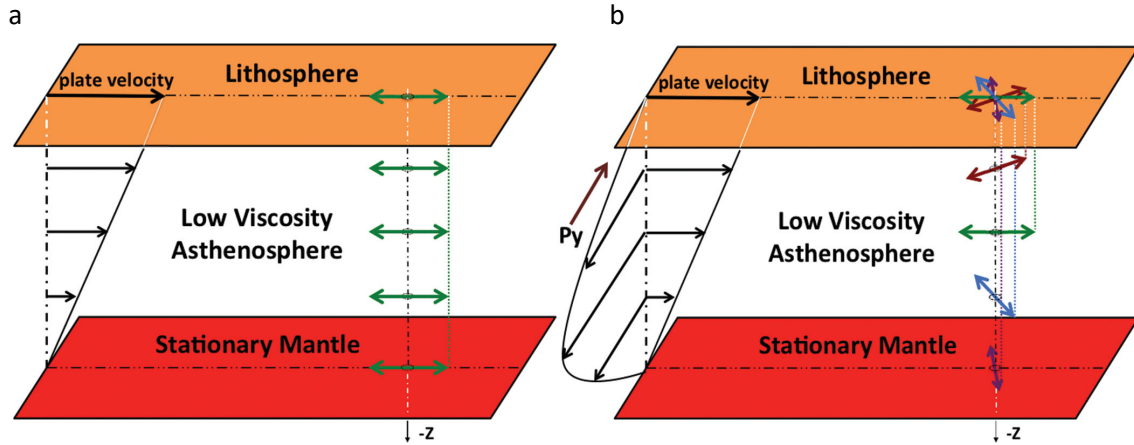


Figure 2.8: Cartoon illustrating types of asthenospheric flow and related lattice preferred orientation (LPO) from Natarov and Conrad (2012). In a, a Couette only flow that does not produce LPO rotation with depth, on the contrary in b, a combined Couette-Poiseuille flow leads to LPO variation in orientation with depth.

Observation and modelling of large-scale plate motions show that individual plates move at independent rates relative to each other, suggesting that contributions of driving and retarding forces likely vary among plates. Plate tectonics may, thus, be driven by a combination of several forces.

The early studies of the Earth's force balance revealed a major issue in discerning the contribution of the basal shear drag mantle force from the forces acting directly on the plate, which still remains today, with the concurrent question of whether plate motion is truly driven by forces within the lithosphere or from the mantle underneath (Forte, 2011).

In the famous series of parameterised forces studies (e.g., Forsyth et al., 1975; Solomon et al., 1975; Chapple and Tullis, 1977) the observed plate velocities were prescribed as an input and a balance of force would be inferred to match with plate motion through an inverse problem. Such an approach could not consider how other mantle forces (e.g., shear drag) could affect plate motions or their evolution over geologic time. Many geodynamic models assume that the mantle surrounding the subducting lithosphere is passive (i.e., isothermal or adiabatic) and that the flow below the lithosphere is entirely generated by the dynamics occurring within the slabs (e.g., McKenzie, 1969; Hager and O'Connell, 1981). For this class of models, the basal drag force is a resistant force as it emerges in the opposite direction to plate motion.

More recently other possibilities in terms of major and minor regional contribution of forces have emerged such as that the outcomes of the traditional force-balance modelling are interpreted as non-unique and dependent on which forces are assumed to be known (e.g., Höink and Lenardic, 2010; Forte, 2011; Höink et al., 2011; Coltice et al., 2019). Some of these studies support the possibility that plates can be driven by both slab-pull and asthenosphere shear drag forces. For example, Ghosh and Holt (2012) show that

basal shear tractions generated by buoyancy-driven mantle flow are fundamental drivers of surface motions and able to generate plate velocities that level more in line with the value currently observed on Earth, in agreement with other works (Rowley et al., 2016; Semple and Lenardic, 2018, 2020). Coltice et al. (2019) find that most of the sub-lithospheric mantle is passive and simply driven by the above lithosphere but it does not resist plate motion. Conversely, other recent studies (e.g., Natarov and Conrad, 2012; Lin et al., 2016) find that other scenarios are also possible where the asthenosphere anisotropy may not align with apparent plate motion in the chosen reference frame and thus imitates a resisting force (figure 2.8). For example, Lin et al. (2016) observations show the direction of asthenosphere fast azimuthal anisotropy coincidence with seafloor spreading, which magnitude can locally exceed that of the plate above but at no depth correlates with the apparent direction of plate motion. Additionally, their study indicates that within the low-velocity layer can coexist two distinct shear zones with a minimum in the anisotropy strength in middle. Natarov and Conrad (2012) explain the rotation of azimuthal seismic anisotropy with depth in the asthenosphere (i.e., LPO axis rotation) as the result of a combined Coulette-Poiseuille flow provided with the addition of a component transverse to plate motion for the horizontal pressure gradient (figure 2.8).

While many challenges remain ahead for the geodynamic community to disentangle, with the advent of large-scale three-dimensional models able to incorporate buoyancy forces derived from seismic studies (e.g., Vigny et al., 1991; Forte and Peltier, 1994; Becker and O’Connell, 2001; Ghosh and Holt, 2012; Forte et al., 2015), it has been determined that plate motions can be predicted on the basis of stresses generated in the mantle which represents a major step towards the development of a unified physical mechanism for the generation of the lithosphere and mantle as a single, internally coupled system (e.g., Lithgow-Bertelloni and Richards, 1998; Conrad and Lithgow-Bertelloni, 2004; Lowman et al., 2008; Forte et al., 2009; Forte, 2011; Rowley et al., 2016; Rolf et al., 2018).

2.3 Subduction zones

A subduction zone is a convergent plate boundary where an oceanic lithosphere descends underneath another tectonic plate (continental or oceanic) due to higher gravitational energy. At mid-ocean ridges new oceanic lithosphere is continuously created and, as it moves away from the spreading ridge it becomes denser due to cooling and thickening over geological time. A gravitational instability generates when the lithosphere is heavier than the surrounding material. As a consequence, it sinks into the underlying mantle for as long as it remains denser than the surrounding mantle rocks (Turcotte and Schubert, 2014). Both the pull of a slab into the mantle and the pull away from the ridge acting on a plate generate a horizontal pressure gradients such as that the surface cold boundary layer feeds the slab perpetually and leads to the appearance that the slab is pulling the plate.

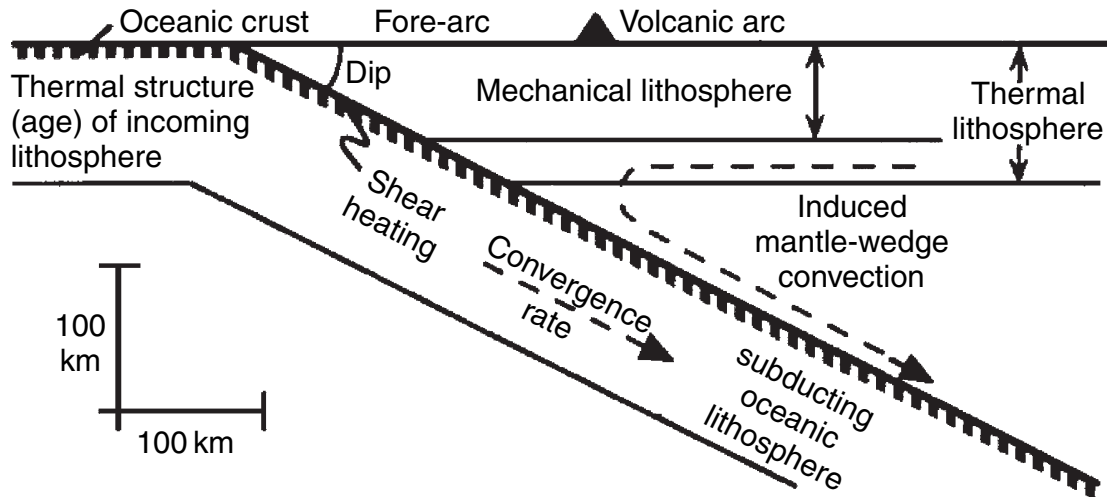


Figure 2.9: Schematic representation of a subduction region reproduced by King (2007) and originally from Peacock (1996).

From a mantle convection perspective subducting slabs can be described as downwelling of the Earth's cold upper boundary layer (Bercovici et al., 2015).

A generalisation of the boundary layer theory and scaling analysis carried out by Davies and Richards (1992) demonstrates that the values of slabs descendent velocity are comparable to the speeds of downwelling in simple convection, while also matching with surface motions and consistent with heat-flux calculations. This shows that the slab pull force is already well represented by employing cold downwelling convection in a simple fluid. This leads to the corollary that slabs not only are the main operators of plate motions and mantle flow but also cool off the Earth's interior.

Subduction zones are associated with substantial island arc and other types of volcanism, transport and recycling of sediments, water and volatile cycles, phase transformations and melt mixing processes (for more information see Oxburgh and Turcotte (1968); Ringwood (1975); Von Huene and Scholl (1991); Pawley and Holloway (1993); Kirby et al. (1996); Marton et al. (1999); Schmeling et al. (1999); Peacock (2001); Elliott (2003); Gaetani and Grove (2003); Rüpke et al. (2004); King (2007) and reference therein). Subduction zone processes are therefore of key importance to understand past and present evolution of the planet.

In an attempt to provide a succinct summary of current reviews and recent papers strictly relevant for the aim and topic treated in this work, the rest of this section will present the broad theme that emerges from recent slabs observations' and geodynamic numerical models from a global subduction dynamics perspective. I would direct the reader a number of recent reviews for more information about seismic observations, mantle wedge, phase change, subduction-transition zones, water in the mantle and subduction modelling, such as Christensen (1995); Bebout et al. (1996); King (2001); Eiler (2003); van

Keken (2003); van Keken and King (2005); Hirschmann (2006); Bercovici et al. (2015); King (2015); Goes et al. (2017).

2.3.1 Subduction zone observations and inferred parameters

The global geological circulation is driven by the negative buoyancy of the descending oceanic lithosphere into the underlying mantle and may result in complex three-dimensional geological structures as reported by a variety of observations across the geosciences (e.g., Isacks and Molnar, 1971; Van der Hilst et al., 1997; Müller et al., 2019). These include slab changes in dip angles along the arc length, tears and windows in the slab and other forms of slab and upper plate deformations. Most of our understanding of the evolution and characteristics of a whole subduction zone comes from indirect observations, such as seismic and anisotropy studies, gravity, topography, geochemistry, petrology, field geology and their integrations across space and time scales. In the last few decades, the advancement of technology magnified the rising in the complexity of the observations and the power of computational infrastructures. These contributed to the increase in numerical resolution and details of rheology formulations adopted in subduction modelling, which now attempts to reproduce specific subduction zones settings, as

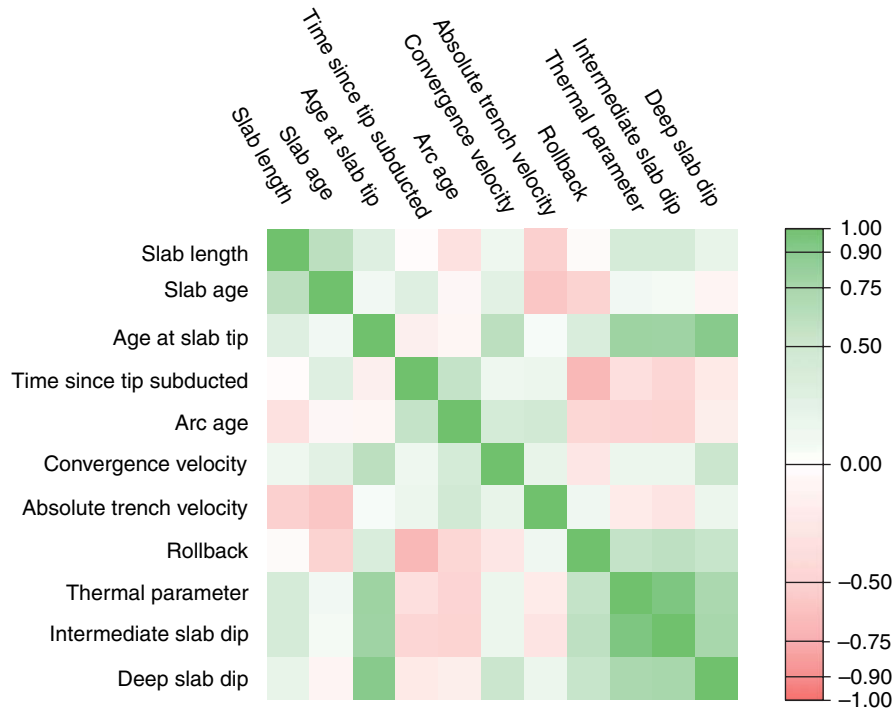


Figure 2.10: Multi-step analysis of 39 contemporary subduction zone parameters implemented through a covariance matrix taken from King (2007) who adapted it from the original work of Jarrard (1986). Green and red indicate a poor and strong correlation, respectively.

well as global deformation patterns, to test different hypotheses that were not possible to consider with previous generations of models (King, 2007).

Given the wide complexity of subduction settings, many studies have tentatively attempted to find a correlation between subduction observable properties and parameters, and especially, slab morphologies, ages and dips. Former classical work (e.g., Isacks et al., 1968; Isacks and Molnar, 1969, 1971) used earthquakes to infer slab morphology through the distribution of down-dip compressive stresses in a subduction zone. One of the most influential studies on this topic was carried out by Jarrard (1986) who employed a compilation of 26 parameters for 39 subduction zones to perform a multi-step correlation analysis. Jarrard (1986) finds that three independent variables correlate with observations: oceanic plate's age, convergence rate and intermediate slab dip (i.e., the average slab dip angle between the trench and the slab from the surface to 100 km depth). The correlation between the duration of subduction and moderate slab dips supports the assumption that subducted slabs are evolving, time-dependent features that are not at steady-state equilibrium. Additionally, another correlation emerged from a subsequent study between the slab thermal parameter (i.e., the product of the convergence rate and the plate age, Kirby et al. (1991, 1996)) and slab length. King (2007) uses a covariance matrix to summarise the outcome of Jarrard's work including the slab thermal parameter (see figure 2.10). This work shows a positive correlation between i) age of the plate calculated at the trench and at the slab tip, which implies that the amount of deformation in the deep slab is limited, and ii) the age of the arc and intermediate and deep slab dips (the average of the slab dip between 100 and 400 km). A very weak correlation is instead observed between slab dip and pattern of trench migration. Furthermore, Jarrard (1986)'s compilation classifies slab dip angles in two groups depending on the nature of the overriding plate. On average continental overriding plates are associated with shallower slab dips ($\sim 40^\circ$) than their oceanic counterpart ($\sim 60^\circ$); however, exceptions exist (e.g., Japan, Solomon and New Hebrides slabs). While slab dip seems to relate to the type of deformation in the upper plate (i.e., compression versus extension), it is not clear whether this is caused by the nature of the upper plate or trench migration.

The advent of tomography has fundamentally helped to shed light on the regional slab structure and geometry. Although it does not alone provide information about the stress field, these studies have imaged slab changes in dip and morphology along the length of many arcs. A summary of slab morphology cross-sections across Pacific subduction zones is presented in figure 2.11. One typical natural example of this complexity is offered by the Cocos slab. Along the length of this subduction region, the dip angle smoothness from north to south of at the least 20 degrees (Colombo et al., 1997) with observed variations in earthquakes' depth occurrence, plate age and speed, trench migrations, heat flow and seamount distributions. All of these variables are known to substantially affect

slab thermal structure, which in turn is known to influence slabs patterns of interactions with the upper-lower mantle transition zone (Loiselet et al., 2010; King, 2015).

Investigations of plate age and style of slab-mantle interactions have shown that stagnating slabs are more common in the western Pacific where the oceanic lithosphere is older on average than on the eastern counterpart whereas slabs are often observed to buckle and penetrate the transition zone (Goes et al., 2017). However, in natural settings, more complexity exists. For example, seismic tomography images show the old Mariana slab penetrating the 660 km discontinuity into the lower mantle, whilst the adjacent Izu–Bonin slab is being deflected at the transition zone (e.g., Zhou and Clayton, 1990; Fukao et al., 1992; van der Hilst and Seno, 1993; Widiyantoro et al., 1999; Miller et al., 2005). Miller et al. (2005) recorded a variation in seismic velocity at 300–450 km of the Izu-Bonin slab and interpreted the change in morphology as due to a tear in the slab. The South American slab also represents another classical example of slab complexity. The deep structure Nazca slab appears continuous both horizontally and with depth over most of the Andean trench, but it penetrates the transition zone in central South America (King, 2007). One more example of slab complexity is the Tonga subduction zone. Different explanations have been found to explain the complicate nature of the detached slab in Tonga. The Tonga slab appears to lie horizontally in the mid-mantle boundary for hundreds of kilometres before falling into the underlying mantle (van der Hilst, 1995). The fact that a seismicity cluster is found in a local correspondence with the location of the flat-lying fast seismic-velocity anomaly and far away from the rest of the observed seismicity in the ongoing subducting lithosphere has lead to the hypothesis that the seismically active flat-seismic anomalies could represent a slab remanent. On the same line of argument, other flat-laying cold anomalies recorded near the 660 km discontinuity located beneath the Japan, Java, and Izu–Bonin arcs, which are generally found aseismic and not considered slab remnants (van der Hilst et al., 1991; van der Hilst and Seno, 1993). Conversely, high-resolution 3D P-wave velocity images in Tonga shows a very narrow and continuous slab (Deal et al., 1999), whilst other of studies (e.g., Hamburger and Isacks, 1987; Fischer et al., 1991; Okal and Kirby, 1998; Gurnis et al., 2000; Chen and Brudzinski, 2001) invoke the contribute of the descending of allochthonous terrains, such the fossil Vitiaz trench and/or the interaction with a mantle plume in influencing the complicate nature of tonga subduction zone.

2.3.2 Trench mobility

The descendent motion an oceanic plate into the underlying mantle is accompanied by the movements of trenches at the surface over time. Trenches can migrate over geologic time and have either retreating, advancing, stationary or combined type of motion. Tectonic trenches are the surface tectonic expression of the direct interaction between a subducting slab and the underlying mantle. Therefore, patterns of trench migration have a substantial

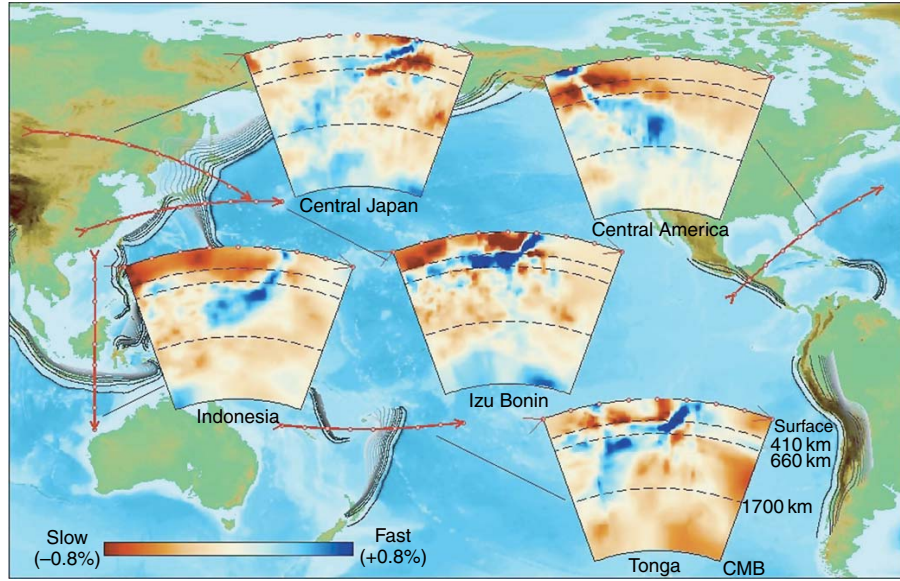


Figure 2.11: Cross-sections obtained through seismic tomography for 5 subduction zone in the Pacific, taken from King (2007) who adopted it from Albarède and van der Hilst (2002).

relevance for the better understanding of both subduction dynamics and surface tectonics. The way how a trench migrates can influence many processes, including the distribution of lithospheric stresses and heat flow, melt mixing, slab penetration into the lower mantle, and toroidal/poloidal mantle flow partitioning throughout a subduction zone. Trench migration mechanisms are also used to draw inferences about slab rheology and patterns of interaction with the upper-lower mantle transition zone and assess the realism of absolute plate convergence models (e.g., Garfunkel et al., 1986; Griffiths et al., 1995; Zhong and Gurnis, 1995; Christensen, 1996; Piromallo et al., 2006; Funiciello et al., 2008; Alisic et al., 2012; Gérard et al., 2012; Williams et al., 2015).

Jarrard (1986) locates as many advancing as retreating trenches currently present on Earth; however, predominant trench retreat was found by other studies (e.g., Garfunkel et al., 1986; Schellart et al., 2008). Subsequently, Heuret and Lallemand (2005) using an updated data sets find consistent outcomes to Jarrods on trench migrations. They further classify all trench motion in two groups one with almost no-trench motion and the other with moderate retreating velocity (up to 5 mm yr^{-1}). Their calculations show that back-arc extension correlates with trench retreat and back-arc compression with advance, also in accordance with Jarrard (1986). The observations of surface motion and plate age from Heuret and Lallemand (2005) exhibit no a clear link between present-day slab dip and age at the trench (Lallemand et al., 2005; Sdrolias and Müller, 2006).

The observed tendency for older and stronger slabs to advance and not to retreat is in disagreement with what is generally found in many geodynamic models (e.g., Gurnis and Hager, 1988; Griffiths et al., 1995; Scholz and Campos, 1995; Becker et al., 1999). Regional

gravitational potential investigations and geoid observations along with convection studies suggest that weak slabs provide a much better fit to the observations than stronger (high-viscosity) plates (e.g., Hager, 1984; Richards and Hager, 1989; Ritzert and Jacoby, 1992; Zhong and Gurnis, 1992; King and Hager, 1994; Moresi and Gurnis, 1996; Chen and King, 1998; Zhong and Davies, 1999). Observational (e.g., Moresi and Gurnis, 1996; Mitrovica and Forte, 1997; Zhong and Davies, 1999) and numerical (e.g., Schellart, 2004a; Bellahsen et al., 2005; Stegman et al., 2006; Di Giuseppe et al., 2008; Funiciello et al., 2008; Wu et al., 2008; Capitanio et al., 2009) studies constrain the viscosity between the oceanic lithosphere and the mantle to be within the range of 100-500. While I will argue more on this topic in the following §2.3.3.2, some authors have interpreted the observations of trench advance as due to a large variation in age of the subducting plate during the Cenozoic (e.g., Goes et al., 2008; Garel et al., 2014), non-linear rheology or interactions with the upper plate (Holt et al., 2015b) among other factors. Other studies invoke the effects of regional and local processes, such as small-scale flow, tears or plume-slab interaction, but also global mantle flow (Hager and O’Connell, 1981) to explain some of the ambiguities in trench kinematics (Gurnis et al., 2000; Cruciani et al., 2005). Cruciani et al. (2005) constructing regionalised upper mantle seismic model maps of slab geometry find more variability in the data. While this may mirror some of the real complexity of natural settings, it also offers displays a better fit between the slab thermal parameter and slab dip than either one taken individually, although the correlation is weak (King, 2007).

Absolute plate and trench motions depend on the choice of the reference frame and are connected to net rotation through their mutual dependence on slab rheology (Heuret and Lallemand, 2005; Funiciello et al., 2008; Schellart, 2008a; Schellart et al., 2008; Alisic et al., 2012; G  rault et al., 2012). The difficulty to find an unequivocal reference frame, and the specifics of the calculations, complicate the ability to draw inferences about subduction zone kinematics. The South America case provides a striking example that summarises these difficulties, as the difference in trench movement is so large that their trenches are predicted to retreat or advance depending on the choice of the reference frame (e.g., O’Neill et al., 2005; Becker et al., 2015; Coltice et al., 2017). It is not surprising then that after more than three decades of studies the relationship between trench migration, slab age and dips still present conflicting outcomes with no consensus on how to reconcile or dismiss any of them.

Although absolute trench motions, as well as plate motions, depend on the choice of the reference frame, some studies (e.g., Goscombe and Gray, 2008; Schellart et al., 2008; Becker and Faccenna, 2009) reveal that trench migration rates are 30-50% of the convergence rates, consistently across different plate motion reference frames. For example, this ratio corresponds to 0.48 for the hotspot reference frame and to 0.3 for the no-net rotation reference frame. Throughout this thesis, I will refer to the ratio between

trench migration and convergent plate velocity as both the subduction partitioning or V_{norm}^* . This is the main parameter discussed and studied in this Ph.D. work and more information about the research outcomes and discussion are available from chapter 4 to 8.

2.3.3 Geodynamics models of subduction and trench mobility

In this subsection, I will take a closer look at subduction numerical modelling and summarise the current literature on how trench and plate move, subducting and overriding plates deform, slab and mantle interact at the transition zone and how all of these factors are interconnected. For detailed information about each of these topic and surface tectonics, I would direct the reader to existing review papers (e.g., King, 2007; Billen, 2008; Becker and Faccenna, 2009; Wada and King, 2015; Coltice et al., 2017; Goes et al., 2017).

2.3.3.1 Dynamic and kinematic models

There are two major classes of subduction numerical and analogues models: fully dynamic and kinematic. In the fully dynamic type, the initial buoyancy of the slab self-sustains the subduction process and the larger-scale mantle flow, and no kinematic boundary conditions are applied. This group usually covers a greater region of the mantle and has a lower resolution. The numerical resolution and the difficulty to include more realistic subduction zone morphologies have hampered the ability of these models to reproduce detailed seismic and geological structures which require a more sophisticated thermal implementation; however, the increasing in computational power of the calculations (e.g., multigrid solvers, mesh refinement) and resources is continuously incrementing our ability to add more complex physical formulation in this category of models. Kinematic models are generally more suitable for the study of local processes, as they allow for the use of higher resolution and can adopt more realistic physical laws and parameters (Billen, 2008). While this class facilitates the study of a phenomenon in greater detail, slab speed is imposed and cannot address the dynamic impact on slab and mantle deformations because it is possible to define slab geometry and motion a priori. Many kinematic models often draw upon the assumption that slabs are rigid (i.e., have limited ability to deform internally) and are employed to investigate slab thermal structure and phase transition effects (e.g., Minear and Toksöz, 1970; Molnar et al., 1979; Peacock, 1991; Bina, 1996; Kirby et al., 1996; Peacock, 1996; Bina, 1997; Marton et al., 1999; Peacock, 2001; Marton et al., 2005).

2.3.3.2 The role of slab stiffness and buoyancy

The majority of the reference frames indicate that most of the present-day Earth's trench motions occur in conjunction with retreat (-VT), and advancing trench motion (+VT) can

be found along a limited number of segments of the Circum-Pacific Belt, accordingly with the analysis of Schellart et al. (2008), as shown in figure 2.5 and figure 2.5. Generations of 2D and 3D numerical and analogues self-consistent models have invoked the role of plate strength and buoyancy as the dominant factors of trench mobility (e.g., Bellahsen et al., 2005; Capitanio et al., 2007; Di Giuseppe et al., 2008; Schellart, 2008a; Ribe, 2010; Stegman et al., 2010; Capitanio and Faccenda, 2012; Fourel et al., 2014). These are “free” subduction experiments that at first did not include any interface at the base of the upper mantle or add an impenetrable lower mantle, all without considering an overriding plate. Outcomes derived from these models suggest that trenches retreat more easily than they advance and that buoyancy variation in the oceanic plate is the controlling parameter for trench mobility (e.g., Zhong and Gurnis, 1995; Lallemand et al., 2005; Schellart et al., 2008; Ribe, 2010; Stadler et al., 2010; Fourel et al., 2014). Accordingly, with these models, the subducting plate buoyancy and the bending resistance at the trench encourage retreat. In these generation of models, retreating trench motion appears associated with young and weak slabs, which being easier to deform can unbend at depth and arrive at the base of the 660 km transition zone with an angle lower than 90° (e.g., Garfunkel et al., 1986; Bellahsen et al., 2005; Di Giuseppe et al., 2008). Conversely, trench advance seems to correspond to high plate viscosity and buoyancy. When a highly viscous and dense slab happens to interact with the base of the upper mantle its stiffness can hamper its own ability to unbend easily at depth resulting in an overturned plate (dip $>90^\circ$) (e.g., Bellahsen et al., 2005; Di Giuseppe et al., 2008; Funiciello et al., 2008; Schellart, 2008b; Ribe, 2010; Stadler et al., 2010); however, roll-over slabs morphologies are not generally observed in seismic studies (e.g., Li and Van Der Hilst, 2010), but western Pacific slabs are on average older, steeper and appear to advance in the majority of the reference frames (Schellart et al., 2008).

In an attempt to provide a geodynamic explanation for the apparent present-day dichotomy among trench motions, slab dip angles, ages and morphologies occurring on the opposite sides of the pacific ocean, (Molnar and Atwater, 1978; Uyeda and Kanamori, 1979; Lallemand et al., 2005), some studies (e.g., Di Giuseppe et al., 2008; Funiciello et al., 2008; Lallemand et al., 2008; Doglioni and Anderson, 2015) appeal to a ‘westward drift’ of the lithosphere. These authors argue that the older age of the Western Pacific ocean may be responsible for the difficulty of the plate to unbend at depth, as it would make the lithosphere stiffer and thicker with time, resulting in trench advance and overturned slabs. Such an advancing motion would, in turn, promotes steeper slab dips reflecting in a more efficient slab pull and faster plate velocities that on the eastern Pacific counterpart (Di Giuseppe et al., 2008; Funiciello et al., 2008; Faccenna et al., 2009). However, westward Pacific trench motions are not observed in other reference frames such as the Indo-Atlantic hotspot or the no-net reference frames (O’Neill et al., 2005; Schellart et al., 2008) where

for example, slow trench advance (and not retreat) is predicted for the central Andes (Schellart et al., 2008).

Further issues regarding a ‘westward drift’ explanation for the motion of the lithosphere arise from what is reported in the same numerical models where advancing slabs move slowly due to their higher strength and large resistance to bending at the trench, whereas larger buoyancy results in a more efficient slab pull which corresponds to trench retreating motion at the surface, and not vice-versa (Bellahsen et al., 2005; Stegman et al., 2006; Schellart et al., 2007; Di Giuseppe et al., 2008; Schellart et al., 2011). From a thermal aging trend of the oceanic lithosphere, both trench advance and retreat are mechanically feasible. More recently thermomechanical calculations (Garel et al., 2014; Agrusta et al., 2017) have shown that old plates, which have larger plate buoyancy and viscosity, have also a higher propensity for trench retreat and slab rollback. Furthermore, plates with higher viscosity are expected to behave more elastically and unbend more easily at depth (Farrington et al., 2014; Fourel et al., 2014). This lead to the inference that slab pull is possibly not the driven mechanism for trench advance and other mechanisms have then been explored to explain trench advance. Some of these mechanisms adding kinematics conditions, such as a ridge push boundary condition enforced either at the trailing end of the downgoing plate (Capitanio et al., 2007; Capitanio, 2013), from the upper plate or at the arrival of buoyant features at the trench (Capitanio et al., 2010b; Magni et al., 2012; Fourel et al., 2014; Goes et al., 2014; Čížková and Bina, 2015), with the effect to promote trench advance. Additionally, in 3D, the toroidal flow may be associated with changes in trench migration mechanisms (Capitanio and Faccenda, 2012), as well as the curvature of oceanic arcs (Morra et al., 2006) and due to lateral changes in slab buoyancy and viscosity. Three-dimensional models generally show a larger propensity for trench advance near the slab-free edges where toroidal motion and extension dominate (e.g., Martinod et al., 2005; Funicello et al., 2006; Morra and Regenauer-Lieb, 2006; Morra et al., 2006; Schellart et al., 2007; Goes et al., 2008; Loiselet et al., 2009; Mason et al., 2010; Stegman et al., 2010; Li and Ribe, 2012; Duarte et al., 2013; Meyer and Schellart, 2013; Schellart and Moresi, 2013; Goes et al., 2014; Magni et al., 2014; Holt et al., 2015b). The observation common to all subduction zones that slabs should not be significantly stronger than the ambient mantle, further constrain the ability to reconcile these models with large-scale observations of plate and trench motions and should be taken in serious consideration, as it could be possible to dismiss any of these hypotheses taken by themselves.

In Chapter 4 of this thesis, we propose an alternative mechanism from which trench advance can easily emerge from numerical calculations and level more in line with values of subduction partitioning presently observed on Earth, consistent with many reference frames and seismic constraints. Our results suggest a mechanism that hampers trench retreat and promotes plate and trench advance motion, regardless of plate age and rheology (Carluccio et al., 2019b).

2.3.3.3 Influence of the overriding plate

Subduction margins are never entirely “free” to move as they are surrounded by neighbouring plates. While it is a challenge to discern the respective roles and feedback between descending and surrounding plates, intraplate coupling and upper-plate dynamics generally appear to provide resistance to trench motion and, in some extreme cases, even prevent subduction (e.g., De Franco et al., 2007; Běhouňková and Čížková, 2008; Androvičová et al., 2013). The amount of upper-subducting plate coupling controls the mobility of the upper plate (Goes et al., 2017). Additionally, thicker, lighter and more extended overriding plates (e.g., Zhong and Gurnis, 1997; Yamato et al., 2009; Capitanio et al., 2010b; Meyer and Schellart, 2013; Garel et al., 2014; Holt et al., 2015b) as well as, the addition of a mantle non-linear rheology (e.g., Holt and Becker, 2016; Billen and Arredondo, 2018) can further inhibit trench motion (e.g., van Dinther et al., 2010; Capitanio et al., 2011). Nonetheless, other studies suggest that the overriding plate can provide an additional driving force to subduction (e.g., van Hunen et al., 2002; Arcay et al., 2008; van Dinther et al., 2010; Čížková and Bina, 2015; Holt et al., 2015a,b). The role of an overriding plate on subduction dynamics is commonly studied employing kinematics models, from which contrasting slab stress patterns can emerge (Čížková et al., 2007), as this type of models may not promote the most energetically favourable approach (Han and Gurnis, 1999). These numerical experiments generally prescribe a fixed trench in contrast to a mobile one. A free to move trench at the surface seem to enhance penetration in the lower mantle, whilst a fixed trench appears to encourage retreat and slab flattening at the transition zone (e.g., Zhong and Gurnis, 1994, 1995; Capitanio et al., 2007; Arredondo and Billen, 2016).

Present-day observations report that slab dip angles are on average 20° steeper beneath oceanic than continental plates. These observations also record a greater propensity for the upper plate above the descending slab to undergo either extension or compression depending on the subducting dip angle (Jarrard, 1986; Lallemand et al., 2005). Since then, many have proposed links between seafloor age, slab width, subduction kinematics, lithospheric properties and type of deformation in the back-arc region (e.g., Barazangi and Isacks, 1976; Uyeda and Kanamori, 1979; Hager et al., 1983; Jarrard, 1986; Bird, 1988; Wdowinski et al., 1989; Lallemand et al., 2005; Lee et al., 2011). While a number of feedback mechanisms seem to occur between subducting and overriding plates, these appear not to be completely captured by geodynamic simulations according to the investigation of Schellart et al. (2008), bound to the Indo-Atlantic hotspots reference frame, from which no single subduction parameter fully correlates with the type of deformation in the upper plate. However, the issues associated with the choice of reference frame remain. Such as, that in the Pacific hotspot reference frame the upper-plate moving towards the trench is associated with compression and the upper-plate moving away from the trench correlates

to back-arc extension, whereas plate and trench motions in the Indo-Atlantic reference frame exhibit the opposite trend (Schellart et al., 2008).

Numerical models generally find that upper-plate extension and trench rollback motions are facilitated by a buoyant continental upper plate (Garel et al., 2014; Sharples et al., 2014; Holt et al., 2015b) whereas, a thicker and denser oceanic overriding plate develops compressive internal stresses and promotes shallower slab dips (e.g., Manea et al., 2012; Holt et al., 2015b; Taramón et al., 2015; Hu et al., 2016a; Rodríguez-González et al., 2016). Other studies attribute the style of overriding plate deformation to seafloor age-related variations (e.g., Molnar and Atwater, 1978; Capitanio et al., 2011). These show that the subduction of younger oceanic lithosphere in the Pacific corresponds to mountain building process, whereas older lithosphere relates to back-arc extensions. Series of regional studies have then been carried to provide explanations to the wide variability of the natural observations. For example, regional numerical calculations of the Nazca plate dynamics show that the older age of the slab beneath Bolivia than on either side, could be responsible for the change in morphology of the slab and induce trench curvature. However, the evolution of this subduction zone remains debated and other hypotheses have risen, such as those invoking the primary role of compressive stresses at the centre of wide slabs in generating a stagnating point and, a consequent change in trench and slab morphology (Russo and Silver, 1994; Schellart et al., 2008); however, neither of these models include an upper plate (Goes et al., 2017).

Finally, the subduction history is often invoked to explain episodic variations in slab dip angles and convergence rates and to provide a mechanism for successive phases of extension and compression in the overriding plate in many regions (e.g., Clark et al., 2008; Martinod et al., 2010; Lee and King, 2011; Capitanio, 2013). For example, in the Andes and Zagros where the formation of these orogenies caused buoyancy and thickness variations, which are thought to have significantly decreased plate and trench mobility over geological time (e.g., Iaffaldano et al., 2006; Husson et al., 2008; Meade and Conrad, 2008).

2.3.3.4 Slab-mantle transition zone interactions and phase transition

The early studies of Kincaid and Olson (1987) and van der Hilst and Seno (1993) of dynamic subduction first proposed the contribute of trench retreat to slab stagnation and flattening in the 660 km discontinuity observed in many segments of the Pacific corridor. This was confirmed by a subsequent generation of both large-scale and regional kinematic numerical experiments (e.g., Griffiths et al., 1995; Guillou-Frottier et al., 1995; Zhong and Gurnis, 1995; Christensen, 1996; Olbertz et al., 1997; Čížková et al., 2002; Torii and Yoshioka, 2007; Agrusta et al., 2017). Some of these authors (e.g., Christensen, 1996; Torii and Yoshioka, 2007) introduced a characteristic trench migration and dip length to predict the amount of slab flattening at the transition zone, but others (Karato et al., 2001; Torii

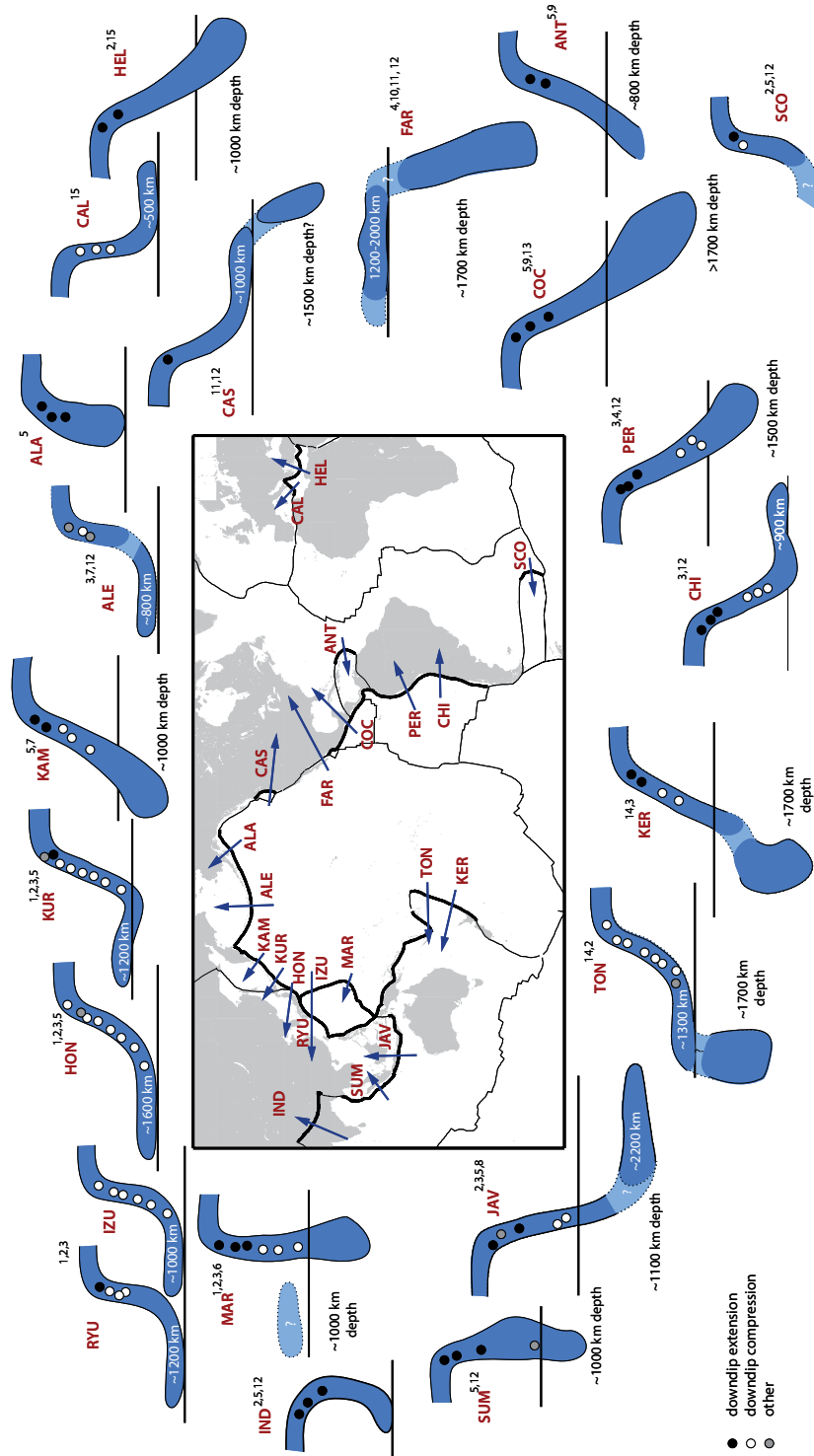


Figure 2.12: Schematic overview of several subducting slab morphologies approaching the upper-lower mantle transition zone as recorded by tomography studies, taken from Goes et al. (2017). Vectors on the maps indicate the approximated cross-sections, circles within the slab provide information about the predominant state of stress, texts within the slab indicate estimated depths for flat slabs in white, as well as for continuous slabs in black.

and Yoshioka, 2007; Agrusta et al., 2017) also stress the importance of transition-zone resistance and slab strength in affecting the deformation style in the discontinuity.

Models of free subduction confirmed the importance of slab internal and external forcing in influencing trench movements and modes of slab interaction in the transition zone (e.g., Ribe, 2010; Stegman et al., 2010; Čížková and Bina, 2013; Garel et al., 2014; Goes et al., 2017). From these investigations, trench retreat does not appear to be a consequence of the slab interaction with the upper mantle transition zone, but trench movements can be affected by the presence of a transition zone - a viscosity jump and/or phase change - (e.g., Becker et al., 1999; Bellahsen et al., 2005; Capitanio et al., 2010b; Lee and King, 2011; Čížková and Bina, 2013; Garel et al., 2014; Agrusta et al., 2017). For example, the buckling of the subducting lithosphere during its interaction with the transition zone can produce alternating phases of trench movements and speed plus variation in the slab dip angle (e.g., Capitanio et al., 2010b; Lee and King, 2011; Čížková and Bina, 2013; Garel et al., 2014). It's worth noticing that slab flattening is also a buckling response of the properties of the lithosphere (Houseman and Gubbins, 1997; Ribe et al., 2007) as models have revealed for cases of weak and/or fast subducting slabs (e.g., Lee and King, 2011; Čížková and Bina, 2013; Cerpa et al., 2014). Furthermore, modes of slabs interaction with the transition zone may influence the development of stress patterns in the overriding plate (e.g., Clark et al., 2008; Capitanio et al., 2010b; Lee and King, 2011), and a passive upper plate itself can contribute enhancing trench motion and slab buckling (Clark et al., 2008). Additionally, the inclusion of non-linear rheology show to affect both trench bending moment and pattern of interaction between the slab and the 660 km discontinuity (Billen and Arredondo, 2018).

The viscosity jumps and the downward deflection of the endothermic phase transition between upper and lower mantle are the two major factors in controlling the resistance of a slab to sink into the lower mantle. Secondly, limited changes in intrinsic density due to variations in chemistry between upper and lower mantle may also contribute (Goes et al., 2017). The endothermic phase transition from ringwoodite to bridgmanite + magnesiowüstite within the cold slab leads to local reduction in slab buoyancy and provides a strong stagnating force (e.g., Christensen and Yuen, 1985; Bina et al., 2001; Stixrude and Lithgow-Bertelloni, 2011; Faccenda and Dal Zilio, 2017). The phase buoyancy parameter in Christensen and Yuen (1985) describes the combined influence of viscosity contrast and phase transition at this discontinuity, which represent local buoyancy forces contribution in relation to the global thermal convective circulation.

Global-scale mantle convection calculations show that inferred upper- to lower-mantle viscosity contrasts from observations do not lead alone to slab buckling or layered convection, and that other factors could be at play or equally relevant for the large mantle circulation (e.g., van Keken and Zhong, 1999; Yanagisawa et al., 2010). A group of studies (King and Ita, 1995; Karato et al., 2001; Čížková et al., 2002) focused on the effect of

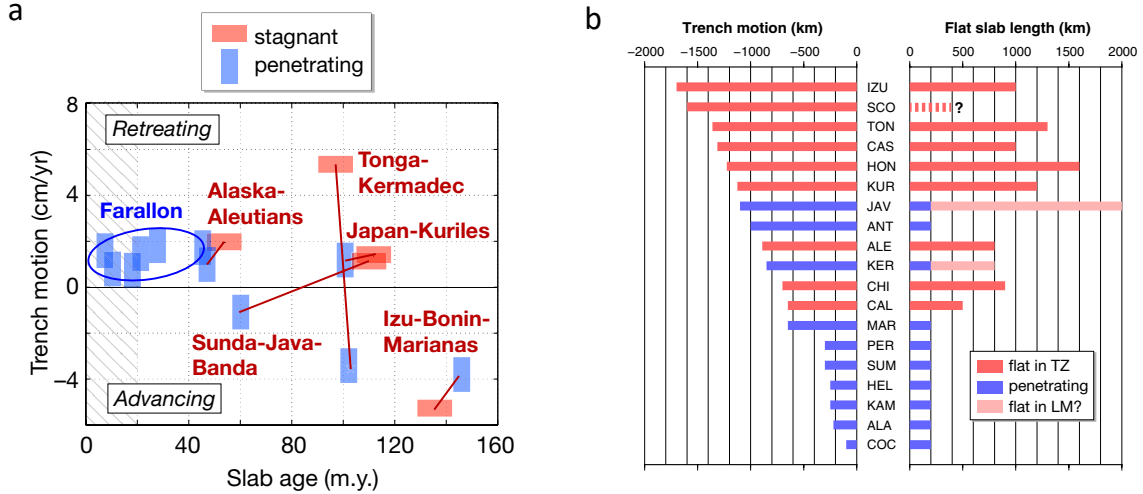


Figure 2.13: In a present-day relationship among trench migration, age of the oceanic lithosphere age and slab shape at the transition zone for some subduction regions. In b, total trench migrations, flat slab lengths and their relationship with mode of deformation at the transition zone approximated over the past ≈ 50 Myr. No simple correlations appear to emerge among all these factors; a and b are reproduced from Goes et al. (2017).

weakening within the slab and buckling in the 660 km transition zone. Their outcomes suggest that slab strength does not exert a major influence on whether slabs penetrate or not when the trench is free to move, but it could have local effects on slab deformation and morphology (Tagawa et al., 2007; Agrusta et al., 2017) promoting slab penetration through the transition zone. The prolonged weakening of the slab over time has also shown to destabilise stagnation with consequences on global mantle flow and plates motions (e.g., Pysklywec et al., 2003; Pysklywec and Ishii, 2005; Motoki and Ballmer, 2015). From which, emerges that plate strength could be a significant factor in controlling how a slab can interact with the transition zone (Goes et al., 2017).

The rheology structure of the slab influences the style of deformation at the 660 km (e.g., from sinking, stagnant to spreading slabs, or sinking or stagnant piles) when small changes in the parameters are imposed (Guillou-Frottier et al., 1995). From a physical point of view, older plates should descend easily in the lower mantle as some models' report (e.g., Ballmer et al., 2015). "Free" subduction experiments show a far more complex picture where older slabs, with a higher propensity for retreat, stagnate and flatten at the discontinuity and younger plates tend to penetrate at steeper dips for intermediate Clapeyron slopes and moderate viscosity jumps, with few exceptions (Goes et al., 2008; Garel et al., 2014; Agrusta et al., 2017). The age of the oceanic lithosphere, which controls its physical and rheological properties, can also substantially contribute to patterns and styles of slab interaction with the transition zone (Agrusta et al., 2017; Goes et al., 2017).

2.3.3.5 Other regional studies

The present-day observations of subduction zones provide only a snapshot of a time-dependent process where each subduction zone is in a different evolution stage (King, 2007). A variety of 3D models has been performed to address the dynamics of specific regions with substantial success in matching with geophysics constraints (van der Hilst and Seno, 1993; Moresi and Gurnis, 1996). These have shown that some of the difficulties relating present-day observations of trench migration and slab geometry can be overcome by taking into consideration that subduction is a time-dependent phenomenon and slabs evolve with time (King, 2007). Additionally, the history of trench retreat appears to correspond with the slab style of interaction in the transition zone.

For example, some areas of the Farallon and the Indian plate moved at a considerable fast speed in the early Cenozoic (e.g., Lee and Lawver, 1995; Müller et al., 2008b; Capitanio et al., 2010a). Such faster velocities have been attributed to slab avalanching into the lower mantle (Goes et al., 2008) or mode of mantle convection (Becker and Faccenna, 2011; Faccenna and Capitanio, 2013). Additional hypothesis including slab-plume interaction (e.g., Cande and Stegman, 2011; van Hinsbergen et al., 2011; Pusok and Stegman, 2020), double subduction system (e.g., Jagoutz et al., 2015; Holt et al., 2017; Faccenna et al., 2018a), removal of a continent keel or a team-play of multiple factors (e.g., Zahirovic et al., 2015) have been explored for India.

Over the Cenozoic, most subducting regions have been characterised by a net trench retreat motion which appears to promote slab flattening. Subducting slabs, such as Kermadec, Izu-Bonin-Marianas, Japan-Kurile- Kamchatka, Peru, with relatively stable over-time trench movement seem to more easily penetrate into the lower mantle (Goes et al., 2017). The only rollover slab is observed in India which is also the only one with a whole net trench advance across the length of the Cenozoic era. However, more recently similar concave shapes have been imaged locally also in the Anatolian, Bolivia and Izu-Bonin slabs (e.g., Replumaz et al., 2004; Sdrolias and Müller, 2006). Plate reconstructions for some other cases, such as the slabs under the Antilles, Java, and Kermadec arcs appear more complex, and, while substantial fluctuations in trench migrations and subduction mechanisms may have occurred over time, these remain highly debated (Ross and Scotese, 1988; Replumaz et al., 2004; Müller et al., 2008b; James, 2009).

Another set of numerical experiments (e.g., Billen and Gurnis, 2001, 2003) can match with a wide range of geophysical observations, for example geoid, topography, and the state of stress in the slab, by taking into account appropriate slab geometries for the Tonga– Kermadec and Aleutian slabs and varying phases rheology. Similar positive fitting to the observation was also found by Jadamec and Billen (2010) for the Alaska subduction zone when taking into account a more realistic morphology and geometry for the subduction zone. These models suggest that the trade-off that exists between slab

buoyancy and slab strength could help to explain the increase of slab buoyancy of $\approx 30\%$ necessary for their numerical results to match with observations of plate and trench motion, geoid and topographic profiles (e.g., Moresi et al., 1996). While these studies provide only an instantaneous snapshot of the flow field among other limitations, their ability to match with observations stress the necessity for more specific investigations of the same kind.

2.3.3.6 Flat subduction

Flat slabs occur beneath some of the segments of the Circum Pacific corridor and provide some of the most striking examples of morphological asymmetry in a subduction zone. These regions include the Pacific Plate underneath Alaska, the Juan de Fuca Plate beneath Washington and Oregon, and the Cocos and Nasca Plates under Central and South America (Bercovici, 2007). The evidence that most of this subducting lithosphere is young than in the western Pacific counterpart, means that the slab pull force is not the dominant mechanism and it may coexist with other relevant forces (van Hunen et al., 2000). Numerical calculations have shown that younger slabs present shallower dip angles than older ones, all other things being equal (e.g., van Hunen et al., 2000, 2004). Additionally, the inclusion of weak crustal rheology, mantle nonlinear rheology, oceanic plateau, viscous shear heating, “subduction fault” (e.g., van Hunen et al., 2000, 2005), weak mantle wedge above the slab (van Hunen et al., 2002; Manea and Gurnis, 2007; Bello et al., 2015), thick and strong overriding plate (e.g., Sharples et al., 2014), extended craton (e.g., O’Driscoll et al., 2009; Jones et al., 2011; Manea et al., 2012; O’Driscoll et al., 2012; Taramón et al., 2015; Hu et al., 2016a) and buoyancy anomalies in the subducting plate (e.g., van Hunen et al., 2002, 2004; Antonijevic et al., 2015; Hu et al., 2016b; Liu and Currie, 2016) all can contribute to crust and mantle weakening enabling shallow subduction. Furthermore, once under-plated, a flat subducting slab can influence the chemical and mechanical structure of the overriding plate producing large-scale continental uplift (e.g., Bird and Baumgardner, 1984; Spencer, 1996; James and Sacks, 1999; Humphreys et al., 2003; Jones et al., 2015). Very recently, Schellart (2020) explains flat subduction invoking the role of old ($> \approx 80\text{--}100$ Myr) and wide slabs (≈ 6000 km) through the subduction history. His numerical calculations show that large subduction zones can decrease their dipping angles at the centre of the plate leading to a stronger mantle suction force in the mantle wedge and shallower slab dip angle. With the use of plate reconstruction data, he argues that flat subduction in the Easter Pacific rim initiated in the geological past when the oceanic lithosphere was older and wide enough. Additionally, he also explains the lack of flat subduction in other areas, such as the Laramide flat-slab subduction in South China, where narrower subduction zone in the geological past may not have allowed the development of very shallow dips.

2.4 Constraints on mantle dynamics during subduction from seismic anisotropy

To provide a basis for the better understanding of the processes that exert first-order control over the global pattern of mantle flow around the subduction zone, especially in the wedge and sub-slab, this section of the literature review dives into the realm of seismic anisotropy. Shear wave splitting measurements offer some of the most relevant constraints available on dynamic processes in the Earth’s interior because of the direct link that exists between mantle deformation and macroscopic seismic anisotropy (Long and Becker, 2010). Seismic anisotropy studies, therefore, represent a very valuable tool for probing mantle dynamics. In subduction settings, the nature of the mantle flow field remains insufficiently understood, despite its relevance for the better understanding of slab dynamics and their evolution through time (Long and Silver, 2008). To deepen the understanding of subduction dynamics and explore the many processes that may contribute to natural patterns of complex three-dimensional mantle fields, this section of the literature review integrates constraints from local, regional and global shear wave splitting studies with insights from geodynamic modelling and mineral physics. This knowledge will, then, be integrated and compared with the outcome of my numerical subduction models presented in Chapter 7 in an attempt to provide a novel mechanism for the evolution of subduction anisotropy in the slab and sub-slab regions of a subduction zone. In the following subsections, I present a succinct summary about: 1) the physical concept behind seismic anisotropy; 2) the techniques used to probe the Earth’s interior; 3) worldwide observed trends; 4) mineral physics constraints; 5) insights on subduction modelling from a global mantle geodynamics perspective; and 6) models and hypotheses for mantle dynamics during subduction.

2.4.1 Physical concept

Seismic anisotropy is one of the tools used to probe the Earth’s interior. The term seismic anisotropy (SA) refers to the dependence of the seismic wave speed on the propagation direction or polarization of the wave in many regions of the Earth, for example, within slabs and in deeper discontinuities, such as the LAB (Long, 2013; Long and Wirth, 2013). When shear waves encounter an anisotropic medium, they split into fast and slow polarized waves and the polarization of these waves, along with the delay time between them, can be measured (Lynner and Long, 2014b). When mantle rocks are the deformed individual grains of intrinsically anisotropic minerals, for instance, olivine in the upper mantle, can develop a lattice-preferred orientation (LPO), or a shape-preferred orientation (SPO) for elastically distinct materials, such as of partial melt, resulting in the directional dependence of seismic wave speed (Long, 2013; Long and Wirth, 2013). As olivine crystals

fast seismic axes typically align with the direction of maximum extensional strain, seismic anisotropy can be used to infer patterns of mantle flow (Karato et al., 2008). Observations of seismic anisotropy can, therefore, provide relatively robust constraints on mantle dynamics given the link between deformation and resulting anisotropy, thus driving most of the scientific interest in characterising seismic anisotropy (e.g., Long and Wirth, 2013; Lynner and Long, 2014b).

2.4.2 Seismic anisotropy techniques to probe the Earth’s interior

The propagation of seismic waves can be affected in many ways by the presence of seismic anisotropy in the Earth’s interior. Some of these ways include splitting or birefringence of shear waves, the directional dependence of P wave travel times, and the difference between the propagation of Love and Rayleigh surface waves, among others (Long, 2013). In subduction and other geological settings, a widely used technique for characterizing upper mantle anisotropy is shear wave splitting (e.g., Vinnik et al., 1989; Silver and Chan, 1991; Long, 2013; Lynner and Long, 2014b). The shear wave splitting technique relies upon the aforementioned physical principle of shear waves splitting into two normalised polarized fast and slow components (quasi-S waves) when travelling through an anisotropic medium. The fast splitting direction (ϕ) provides information about the orientation of the fast wave and, thus, the strain geometry. The delay time (δt) between the time of arrival of the two polarised waves contains details about the strength and/or the thickness of the anisotropic layer (Lynner and Long, 2014b). Then, constraints provided by laboratory experiments on mineral physics are used to interpret the geometry of mantle from observations of shear wave splitting (Long, 2013).

SKS phases are typically used to probe subduction systems, as they can provide an accurate measure of anisotropy in the sub-slab mantle, slab, and mantle wedge (e.g., Andersson et al., 2004; Baccheschi et al., 2007; Christensen and Abers, 2010; Hanna and Long, 2012; Long, 2013). The shear wave splitting technique can be applied to SKS phases (Long, 2013). This converts SKS phases from a P wave in the outer core to an S wave at the core-mantle boundary and, therefore, the polarization of the wave before it undergoes splitting is known (Long and Becker, 2010). From a physics perspective, SKS splitting may record anisotropy in any place along the ray path from the core-mantle boundary to the surface, but they are generally employed to survey the upper mantle, with a few exceptions (e.g., Long, 2009). Other common techniques include teleseismic S originating from deep earthquakes (e.g., Marson-Pidgeon and Savage, 1997; Long and van der Hilst, 2005), ScS triggered at regional distances (e.g., Tono et al., 2009), local S phases derived from slab earthquake (e.g., Smith et al., 2001; Pozgay et al., 2007; Abt et al., 2009; Wirth and Long, 2010), or combinations of the last two methods (e.g., Fouch and Fischer, 1998; Soto et al., 2009; Abt et al., 2010).

While the theory underneath seismic anisotropy appears straightforward, seismic anisotropy measurements can be significantly noisy making it difficult to isolate a particular source of anisotropy. Therefore a variety of techniques has been developed to simplify this process. For example, SK(K)S phases have been commonly used to isolate contributions in the sub-slab volume from seismic stations located above a subduction zone (Lynner and Long, 2014b). This technique can apply a correction for the effect of anisotropy in the mantle wedge (e.g., Long and Silver, 2009; Abt et al., 2010) in order to isolate the sub-slab; however, some of these corrections for wedge anisotropy may often be inexact, as anisotropy in the mantle wedge are usually intricate (e.g., Wirth and Long, 2012; Long and Wirth, 2013; McCormack et al., 2013). While the SK(K)S phases approach can provide a first-order overview of the anisotropy beneath slabs, it may become more problematic when trying to place detailed constraints on sub-slab anisotropy. On the other hand, the source-side shear wave splitting technique has recently been shown to be more effective in its ability to isolate the contribution of the sub-slab alone. In light of this, a range of sub-slab anisotropy patterns along some of the subducting segments of the Pacific plate has been recently reclassified (see Lynner and Long (2014b,a)).

The source-side shear wave splitting method uses teleseismic S phases generated from earthquakes in the subducting slab and recorded at distant stations (e.g., Vinnik and Kind, 1993; Russo and Silver, 1994; Müller et al., 2008a; Russo, 2009; Russo et al., 2010; Foley and Long, 2011; Di Leo et al., 2012; Eakin and Long, 2013; Lynner and Long, 2013). Provided that the anisotropy signal beneath the receiver is taken into account, this technique is able to isolate the signal from anisotropy underneath the slab. This approach appears to provide better constraints on anisotropy beneath subducting slabs while circumventing the complexities associated with the mantle wedge. Additionally, it can provide more elaborate information about the variation of anisotropy both at various depths and along strike. For more details about reviews of these technologies and their applications, I would direct the reader to the existing literature (e.g., Silver, 1996; Savage, 1999; Vecsey et al., 2008; Long and Becker, 2010; Long, 2013) including global and regional databases of splitting measurements (e.g., Liu, 2009; Wüstefeld et al., 2009).

2.4.3 Mineral physics constraints

Seismic anisotropy are due to the lattice crystallographic-preferred orientation (LPO) of intrinsically anisotropic minerals and shape-preferred orientation (SPO) that result in the alignment of material with distinct isotropic elastic properties (in the relation to the neighbouring material), such as melt pockets, or cracks in the shallowest crust (Long and Becker, 2010; Long, 2013). An LPO develops when mantle materials deform under the dislocation creep regime. Olivine is the primary constituent of the upper mantle rocks and has a large single-crystal shear wave anisotropy ($\approx 18\%$). In the upper mantle, olivine LPO crystals are predicted to make the largest contribution to anisotropy, with

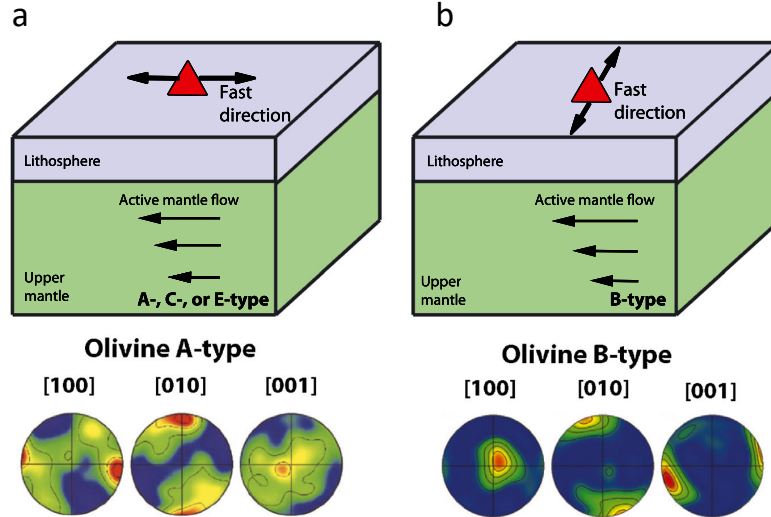


Figure 2.14: The influence of A and B-type olivine fabric on the interpretation seismic anisotropy through a schematic cartoon, reproduced from Long and Becker (2010). In a, fast seismic axes tend to align parallel to horizontal flow direction for A-, C-, or E-type types of olivine fabric; in b, perpendicular fast flow direction generally result from B-type olivine fabric. Typical examples of olivine aggregates distribution in both conditions are shown at the base of the cartoon and were reproduced from Jung et al. (2006); Karato et al. (2008).

orthopyroxene contributing for a smaller amount. In the lower mantle, other minerals can also affect anisotropy, such as perovskite, post-perovskite, and ferropericlase (Long and Becker, 2010). In certain regions of the upper mantle, for example in the mantle wedge of subduction zones, the contribution from the shape preferred orientation of partial melt to seismic anisotropy can become relevant (e.g., Ando et al., 1980; Zimmerman et al., 1999; Vauchez et al., 2000), and may result in complex patterns of mantle flow (e.g., Karato et al., 2008; Mainprice, 2010; Long, 2013).

In the subduction system, there are up to 5, from A to E, types of olivine fabric significant to the study of mantle anisotropy (figure 2.14). The A-type of olivine LPO develops under high strain simple shear deformation and results in the tendency of the fast axes of individual olivine crystals to align in the direction of shear. This type of deformation prevails at relatively low stresses, high temperatures, and low water contents (Karato et al., 2008). The general principle for which fast shear wave splitting directions generally follows the orientation of horizontal mantle flow beneath a seismic station is based upon the simple shear experiments on aggregates of olivine of Karato et al. (2008). Supported by the outcomes of these experimental results, concurred the assumption that vertical velocity gradients in the mantle result in similar patterns of developed shear strain (Long, 2013). The A-type olivine LPO rule of thumb remains today commonly used in subduction frameworks that interpret measurements of anisotropy in terms of patterns of

mantle flow. However, even in the simplest two-dimensional flow model, patterns of shear wave splitting and mantle flow are expected to be significantly more complex than in the case of horizontal simple shear (e.g., Chastel et al., 1993; Levin et al., 2007). Patterns that can be further complicated by certain mantle rocks that carry out their own strain history and the uncertainties related to how to quantify the strength of an anisotropy in deformed mantle aggregates (e.g., Kaminski and Ribe, 2002; Becker et al., 2006).

Many recent works have shown that olivine LPO geometries may be affected by different factors such as water content, stress, temperature, pressure and melt, all being relevant for the characterization of seismic anisotropy in the subduction system (Long, 2013). The type-B olivine fabric dominates at high water content, low temperature and high stresses as confirmed by experimental work (e.g., Jung and Karato, 2001; Jung et al., 2006; Karato et al., 2008) and found in natural peridotite rocks (e.g., Mizukami et al., 2004; Skemer et al., 2006; Tasaka et al., 2008). Such LPO fabric results in the tendency for olivine fast axes to align in the shear plane, changing by 90° the relationship between the fast splitting direction and the inferred flow direction (Jung and Karato, 2001).

Other type olivine fabric include C-, D-, and E, with the dominant fabric-type among these being determined by stress and temperature conditions, as well as water content during deformation. These fabrics have been identified by a series of experimental studies (e.g., Katayama et al., 2004; Jung et al., 2006; Katayama and Karato, 2006), as well as, documented in natural rock samples (e.g., Mehl et al., 2003; Michibayashi et al., 2006; Tommasi et al., 2006; Jung, 2009; Mainprice, 2010). Many studies (e.g., Couvy et al., 2004; Mainprice et al., 2005; Raterron et al., 2007; Jung, 2009) highlight the importance of pressure in controlling the fabric geometry. Some of these (Couvy et al., 2004; Mainprice et al., 2005; Jung, 2009) suggest a pressure-induced transition from B to C, and vice versa, being particularly important for the interpretation of sub-slab anisotropy (Long, 2013). While some uncertainties remain, including those regarding the effect of partial melt on patterns of mantle flow, melt appears to develop type-B olivine LPO (Holtzman et al., 2003); yet the applicability of some of these experiments to mantle rocks is still debated (Karato et al., 2008). The C- or E- types of olivine fabric are good candidates to dominate in parts of the wedge away from the low- temperature, high-stress fore-arc mantle (Kneller et al., 2005) and in the sub-slab mantle (Karato et al., 2008). The differences between C, E with A-type fabric can be more elusive. These types show no significant difference in the expected fast direction for horizontal mantle flow. However, the vertical radial flow case shows that the C-type fabric tends to align normal to the shear plane and, type A and E fabric display only small splitting. For intermediate cases of dipping layers (e.g., above or beneath the dipping sides of the descending slab) differences in splitting signatures are more difficult to discern and may depend on the exact geometry and deformation history of the dipping layer (e.g., Skemer et al., 2012).

The identification of a certain type of olivine fabric has, thus, a consequence on the interpretation of seismic anisotropy. The presence of water in the subduction system can further complicate the assessment of an olivine LPO. Petrological processes such as lithospheric formation at mid-ocean ridges, hydrothermal circulation in oceanic crust, bending-induced faulting and sediments melting during subduction can destabilise several hydrous minerals in the subducting slab at large pressures and temperatures resulting in water being released (e.g., Cagnioncle et al., 2007), provoking melting (e.g., Grove et al., 2009) and other crustal-mixing processes in the wedge mantle (e.g., Bostock et al., 2002; Hyndman and Peacock, 2003; Reynard et al., 2007; Hilaireret and Reynard, 2009).

Some petrological studies (e.g., Katayama et al., 2009; Mainprice and Ildefonse, 2009; Bezacier et al., 2010; Van de Moortèle et al., 2010; Jung, 2011; Nishii et al., 2011) through experimental work suggest that also hydrous minerals mineral could strongly contribute to seismic anisotropy. Such minerals include antigorite which phases show significant single-crystal shear wave anisotropy, AVs, (e.g., talc (AVs = 68%), chlorite (AVs = 76%), hornblende (AVs = 31%), brucite (AVs = 31%), from Mainprice and Ildefonse (2009)), and are all significant for the characterisation of the shallow part of the mantle wedge from seismic anisotropy (e.g., Kern, 1993; Mookherjee and Capitani, 2011). From which follow that even if serpentinite is only present in relatively thin layers in the mantle, it can still make a significant contribution to the observed anisotropy. In fact, in subsequent studies (e.g., Katayama et al., 2009) has been shown that hydrous minerals promote alignment of waves fast splitting in the direction parallel to the trench above the subducting slab; however different outcomes were also found in other studies (see Jung, 2011). Such conflicting findings highlight the complexity of the work involved whilst trying to constrain mantle anisotropy, especially for the case of the mantle wedge. The current literature shows that more work is still necessary to clarify the relationships between deformation geometry and the resulting anisotropy around a subduction zone, including a more in-depth multidisciplinary effort for the understanding of why deformation experiments and natural samples sometimes yield to different outcomes (Long, 2013).

2.4.4 Anisotropy in the mantle wedge, the subducting slab and sub-slab

Around a subduction zone, seismic waves splitting corresponds to regional anomalies in the direction of maximum finite strain and provides useful information about the geometry of the mantle flow and strength or thickness of the anisotropy (Long, 2013). This is particularly relevant for our understanding of how slabs sink from surface to bottom and help to shed lights on many processes, including surface motions, overriding and subducting plates deformations, back-arc formation and melt and volatiles transport in the wedge and behind the subducting slab, among others. While a variety of ray path combinations can be utilised to assess anisotropy in various parts of the subduction system,

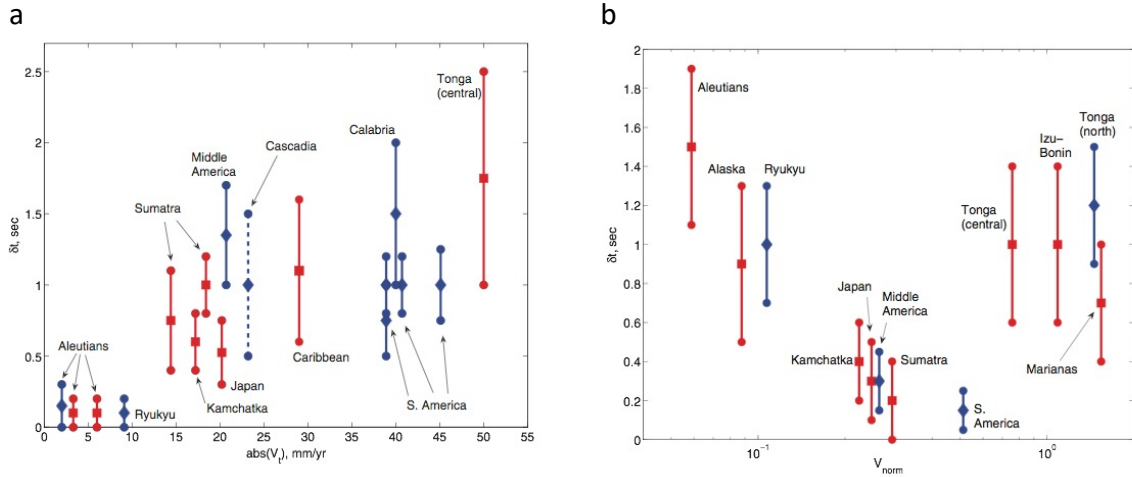


Figure 2.15: In a, average sub-slab delay time, δt , versus absolute trench migration rate, $|V_t|$; in b, average wedge delay time, δt , versus the ratio between absolute trench migration and convergence rates for individual subduction segments, V_{norm} . Red lines refer to advancing trenches and blue to retreating trenches. The figure is reproduced from Long and Silver (2008).

splitting studies in subduction zones commonly focus to anisotropy in the sub-slab mantle, slab, and mantle wedge.

2.4.4.1 The mantle wedge

The mantle wedge above the subducting slab is important for a variety of processes and especially for the transport and storage of melt and volatile phases. Figure 2.16 shows Long and Silver (2008) compilation of average fast direction of shear waves splitting for many subduction zones in the Pacific. The documented patterns of mantle flow in the wedge include whole-trench-parallel, whole-trench-perpendicular, or trench-parallel to trench-perpendicular transition and vice versa from the backarc to the fore-arc. The overall recorded average delay times range from 0.1 up to 1.5 s (Long, 2013). There is currently no single model able to explain such large results variability, however, some factors have shown to be more important than others in controlling the wedge mantle flow. A global compilation of seismic anisotropy studies (Long and Silver, 2008) has shown that mantle wedge delay times of anisotropy correlates positively with the ratio between convergence and trench migration rates, which is V_{norm} (figure 2.15). V_{norm} , in turn, could be sensitive to a broader spectrum of subduction parameter related to kinematics and mantle rheology, ambient mantle flow, small-scale convection, proximity to slab edges, and slab morphology, and the contribution of the local temperature structure can promote the development of type-b olivine and/or serpentinite fabric and volatile distribution in the wedge (Long and Wirth, 2013). While previous work has elucidated some mantle dynamics behind certain modes of wedge anisotropy and subduction-related parameters,

our understanding of the mantle wedge dynamics and the origin of seismic anisotropy signatures is far from complete (Long, 2013; Long and Wirth, 2013).

2.4.4.2 The subducting slab

Patterns of seismic anisotropy in the slab are sparser and less constrained than within the sub-slab or wedge mantle. From a global comparison study (Long, 2009) emerge no clear relationship between the mostly trench-parallel fast SKS splitting directions and fossil spreading direction in the oceanic lithosphere. This suggests that the primary contribution to SKS anisotropy derives from heterogeneities structure within the slab instead that frozen structures inherited from the time of lithospheric formation. Some studies (e.g., Brudzinski et al., 2007; Wang and Zhao, 2008; Huang et al., 2011a,b) have invoked the dominant contribution of the shallow part of the slab in inferring total SKS splitting direction. For example, some argue for the role of aligned serpentized cracks (Faccenda et al., 2008), clay minerals such as talc in the uppermost part of the slab (Song and Kim, 2012a) or frozen lithospheric anisotropy beneath the dipping part of the slab (Wang and Zhao, 2008; Song and Kim, 2012b; Wang and Zhao, 2012; Audet, 2013) all primarily reflecting strong anisotropy in the shallow (40-50 km) dipping part of the slab, that positively correlate with the fossil spreading directions for the slab, but with no obvious counter-evidence in SKS splitting measurements (Long, 2013).

2.4.4.3 The sub-slab

Global trends of sub-slab anisotropy appeared largely dominated by trench-parallel fast splitting directions (Long and Silver, 2008; Long, 2009, 2013); however, more recently this view has changed. Measurements of shear wave splitting of SK(K)S phases can suffer the sporadic locations of seismic instrumentation, which may generate an inadequate lateral and depth representation of the volume beneath the sub-slab mantle, as explained in previous §2.4.2. Source-side shear wave splitting (e.g., Lynner and Long, 2014b,a) studies provide further constraints on sub-slab anisotropy. Lynner and Long (2014b) show that both trench and plate-motion-parallel flow trends are more equally distributed across the globe. To date, only two parameters seem to correlate positively with anisotropy strength and average fast direction: trench migration (in the Pacific hotspot reference frame) (Long and Silver, 2008) and slab ages (Lynner and Long, 2014b). Based on their observations Lynner and Long (2014b) propose that the sub-slab mantle is characterized by two distinct anisotropic and mantle flow regimes. Following this view, the sub-mantle beneath younger lithosphere (<95 Myr) would be dominated by 2D entrained flow causing the entrainment of thick mantle layer, conversely beneath older lithosphere (>95 Myr), would result in the entrainment of a thin sub-lithospheric decoupling layer (figure 2.17).

2.4.5 Geodynamics modelling and seismic anisotropy

A variety of numerical and laboratory modelling tools (e.g., McKenzie, 1979; Fischer et al., 2000; Hedjazian and Kaminski, 2014) have been developed to simulate and gain insights into the process of subduction geodynamics. These models have helped to provide knowledge into the study of seismic anisotropy and how it relates to the dynamics of three-dimensional (3D) mantle flow during subduction. Seismic anisotropy develops when mantle materials are deformed in the dislocation creep regime, which corresponds to non-Newtonian rheology (Long, 2013). The treatment of rheology in numerical models is therefore important. Many published geodynamic studies (e.g., Piromallo et al., 2006; Paczkowski et al., 2014b) employ Newtonian rheology to model the development of seismic anisotropy due to mantle flow, which is easier to implement; however, it represents a significant simplification for anisotropy studies and may not be able to fully predict some important aspect of subduction dynamics (e.g., Billen, 2008).

A series of additional methods have been developed to relate subduction models and observations of seismic anisotropy. Some of these include visualisation tools such as calculating the velocity field streamlines (e.g., McKenzie, 1979; Paczkowski et al., 2014b; Király et al., 2017), locally applying appropriate elastic tensor (e.g., Fischer et al., 2000; Hall et al., 2000; Long et al., 2007), or involving a statistical tool for simulating the development of LPO, such as DRex (e.g., Lassak et al., 2006; Conder and Wiens, 2007; Morishige and Honda, 2011; Faccenda and Capitanio, 2012, 2013; Hedjazian and Kaminski, 2014).

In the literature previous 3D numerical and laboratory subduction experiments have investigated the development of toroidal flow associated with trench retreat (e.g., Funicello et al., 2003; Piromallo et al., 2006; Becker and Faccenna, 2009; Capitanio et al., 2010b; Stegman et al., 2010; Capitanio and Faccenda, 2012; Faccenda and Capitanio, 2012, 2013; Király et al., 2017). A smaller subset of these models has focused on the relationship between mantle-lithosphere properties and anisotropic signals that may result from the development of toroidal flow into the wedge and sub-slab mantle. Numerical models show that trench retreat (e.g., Becker and Faccenna, 2009; Becker and Kaus, 2010), toroidal flow at the slab free edge (e.g., Becker et al., 2003; Faccenda and Capitanio, 2012), and buoyancy variation within a slab (Capitanio and Faccenda, 2012) trigger mantle flow parallel to the trench in the sub-slab, which seems to require little or no coupling with the underlying mantle (Paczkowski et al., 2014b) and is promoted by slab penetration in the lower mantle (Paczkowski et al., 2014a). For the fore-arc (Faccenda et al., 2008; Faccenda and Capitanio, 2012, 2013) show that trench-perpendicular seismic anisotropy prevails in the shallow mantle during plate advance, whereas trench-parallel anisotropy dominates in the lower mantle during rollback.

Paczkowski et al. (2014a,b) perform 3D instantaneous kinematic calculations and report that both trench-parallel and trench-perpendicular signatures beneath the subduct-

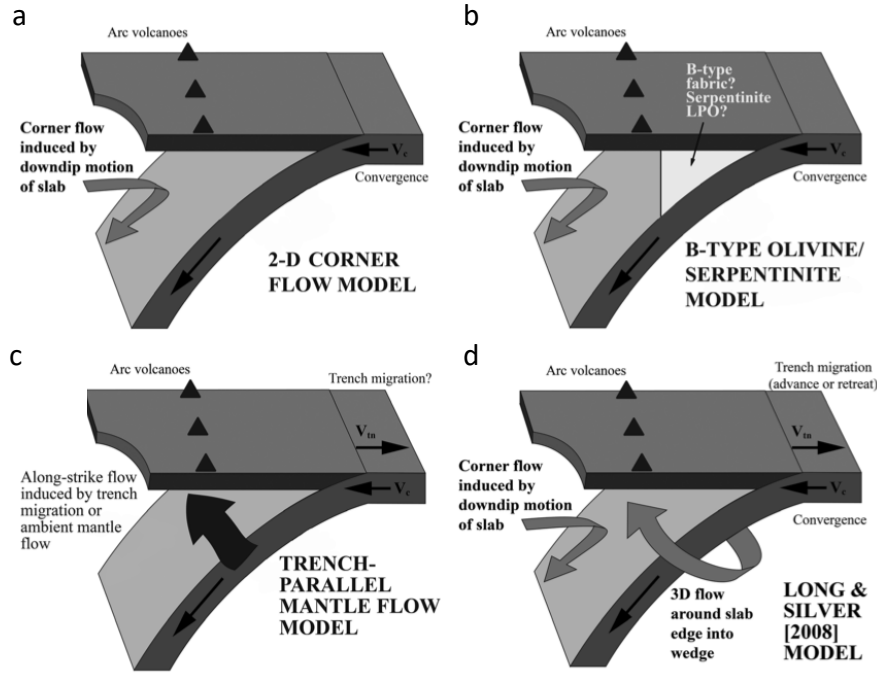


Figure 2.16: Models of shear wave splitting regimes, taken from Long and Wirth (2013). From (a) to (d) are shown the following: the two-dimensional corner flow, the B-type olivine, the along-strike flow and the entrained mantle flow model.

ing can be due to the slab dip angle and maximum depth, the amount of mechanical decoupling between the slab and the subjacent mantle and the orientation and magnitude of the regional mantle flow. Her result suggests that trench-parallel signature predominates in long slabs that penetrate in the deeper mantle, whereas trench-perpendicular anisotropy prevails in shorter subducting slabs with larger coupling between slab and sub-lithospheric mantle.

Kneller and Van Keken (2007, 2008) employ 3-D kinematic models and implement more realistic slab geometries. Their findings show that complex plate morphologies and steep slabs promote the development of trench-parallel anisotropy in the mantle wedge successfully merging their outcomes with observations for the Mariana and southern Andean subduction systems. They also show that mantle flow induces a strong toroidal flow component and trench-parallel flow in the wedge up to 100 km away from the slab edge. A subsequent series of fully dynamics and instantaneous models by Jadamec and Billen (2010) simulate the Easter Alaska subduction system using realist slabs shapes. Their study displays how the employment of non-Newtonian rheology promotes toroidal flow in the wedge decoupling the slab from the flow in the underlying mantle. Other studies also suggest the contribution of small-scale convection originated from lower crust in favouring trench-parallel flow in the mantle wedge (Behn et al., 2007).

2.4.6 Dynamics of the mantle during subduction

A puzzling trend emerges globally in the observations of seismic anisotropy for the wedge and sub-slab mantle. This trend is not often compatible with the simplest model for mantle dynamics, which invokes two-dimensional entrained mantle flow caused by overriding plate motion and simple (A-, C-, or E-type) olivine LPO fabric, as it can not adequately predict splitting in the observations of many natural subduction settings (Lynner and Long, 2014b). Several sub-slab and wedge mantle flow models have been proposed to aid the interpretation of any possible correlation between seismological indicators and subduction parameters (see figure 2.16 for a summary of mantle wedge flow regimes). Among the many subduction parameters, the convergence velocity is possibly the most recurrent one, although, it is the amount of strain rather than the plate velocity itself which controls the strength and geometry of a LPO of a mantle material (e.g., Zhang and Karato, 1995; Faccenda and Capitanio, 2012). Plate velocities are often used as an imperfect proxy for finite strain and anisotropy strength because they can be directly observed (in contrast to mantle strain) and indirectly related to mantle strain (i.e., faster-moving plates are more likely to induce larger finite strains in the mantle). Below, are described the most popular models that have been proposed to predict global dynamics of the mantle during subduction.

2.4.6.1 Wedge mantle: hypothesis and models

The observations of both trench-parallel and trench-perpendicular flows have led to several hypotheses having been proposed to predict global dynamics of the mantle wedge during subduction (Long and Wirth, 2013; Lynner and Long, 2014b). Here I summarise eight of the most popular models:

- (i) The 2D corner flow is the simplest model for characterising mantle wedge anisotropy (figure 2.16a). This model implies viscous coupling between the slab and the underlying mantle resulting in plate-motion-parallel flow above the slab and beneath the overriding plate. For the 2D corner flow model, an analytical solution exists (e.g., McKenzie, 1979; Turcotte and Schubert, 1982) which has been used to model the thermal structure of the wedge (e.g., Kelemen et al., 2003), as well as the location of arc volcanoes (e.g., Grove et al., 2009). If the corner flow model combined with A-type or similar olivine fabric was the prevailing situation in the wedge, both a predominantly trench-perpendicular fast directions and an increase in anisotropic strength with increasing convergence velocity would be anticipated. However, this is not what is always observed in nature (e.g., Long and Silver, 2008) and a trench-parallel signature tends to dominate in the wedge. Alternative models for wedge anisotropy have then been proposed.

- (ii) The olivine B-type fabric and other hydrous phases can contribute to anisotropy pattern by changing of 90° degrees the relationship between strain and resulting fast splitting direction (figure 2.16b). The first-order prediction that emerge from the B-type fabric model suggests a transition from trench-parallel in the fore-arc region to trench-perpendicular in the back-arc region. Experimental and numerical studies (e.g., Kneller et al., 2005, 2007; Kneller and Van Keken, 2007, 2008) show that such fabric may dominate under the conditions present in a) the shallow part of the mantle wedge for low temperatures, high stresses, and the presence of a significant amount of water (e.g., Jung and Karato, 2001; Jung et al., 2006); ii) in the fore-arc mantle wedge (e.g., Kneller et al., 2005, 2007) explaining the transition from B-type olivine in the fore-arc to A-, C-, or E-type in the remaining part mantle wedge registered in several subduction zones (e.g., Nakajima and Hasegawa, 2004; Lassak et al., 2006; Long and van der Hilst, 2006; Kneller and Van Keken, 2008); or iii) in a thin layer directly above the slab (e.g., Tasaka et al., 2008; Katayama et al., 2009). Other contributions to splitting can come from others anisotropic minerals, such as serpentinite rocks, volatile elements, or other hydrated phases released in the mantle wedge from the subducting slab (e.g., Kneller and Van Keken, 2008; Hilairet and Reynard, 2009). Some line of evidence (e.g., Dewandel et al., 2003; Katayama et al., 2009; Bezacier et al., 2010; Van de Moortèle et al., 2010; Jung, 2011; Nishii et al., 2011) reports that serpentinite LPO in a thin layer above the subducting slab can produce trench-parallel fast direction and large delay times (≈ 1 s). While many factors can influence the processes that contribute to oceanic lithosphere hydration, if serpentine LPO was to represent the dominant mechanism for anisotropy in most mantle wedges, one might expect a correlation between (1) the transition from trench-parallel to trench-perpendicular (as for the B-type model) and the location corresponding to the breakdown in antigorite stability; and (2) the delay time and plate age (i.e., temperature) or dips, as these parameters influence where serpentinization of the oceanic lithosphere is likely to occur (e.g., Ranero et al., 2003); however no clear correlation emerge between these subduction zone parameters and patterns of wedge mantle anisotropy.
- (iii) The hypothesis that a mantle wedge trench-parallel flow can be driven by trench migration and a combination of A, C or E olivine fabric (e.g., Turner and Hawkesworth, 1998; Smith et al., 2001; Pozgay et al., 2007; Hoernle et al., 2008; Abt et al., 2009; Heyworth et al., 2011) is supported by several studies (e.g., Conder and Wiens, 2007; Jadamec and Billen, 2010; Faccenda and Capitanio, 2012) in individual subduction systems (figure 2.16c). These report along strike pressure gradients and rapid mantle flow around a slab free edge triggered by trench-parallel mantle flow in the wedge, and possibly due to the migration of the trench or background mantle flow. If trench

migration was the primary cause for mantle wedge anisotropy, one might expect to see a relationship between delay times in shear waves arrival (as a proxy for the strength/coherence of trench-parallel flow and the resulting anisotropy) and trench migration rate (V_t or $|V_t|$). However, a strong correlation does not emerge between these two parameters, as well as, there is no apparent relation between δt and the along-strike component of the ambient mantle (Long, 2013). Furthermore, if the along-strike component of the background mantle flow was the primary factor for trench-parallel flow, then downgoing slabs would need to be decoupled from the wedge above them and do not themselves represent the major driver of wedge flow, which is highly improbable at the global scale.

- (iv) Subduction zone obliquity and transpression in the shallow part of the mantle wedge may resulting in trench-parallel fast directions in the wedge (e.g., Mehl et al., 2003). A corollary of this model predicts little or none 2D corner flow compatible anisotropy for systems with small or no obliquity. While subduction curvature could exert a second-order control on mantle wedge anisotropy, it is unlikely that it would be the major driver of flow in the wedge or one would expect to see a correlation between anisotropy strength and the amount of trench-parallel slab motion, expressed as either the subduction obliquity angle or the trench-parallel component of the convergence velocity, V_c , which has yet to be found.
- (v) Geodynamic numerical models by (Kneller and Van Keken, 2007, 2008) suggest that trench-parallel anisotropy in the wedge mantle can be induced by complex local or regional slab morphology. They successfully apply their models to the Marianas and South American subduction systems and propose that the geographic location of slab complexity correspond to trench-parallel flow in the wedge. Yet, this model applicability to global natural cases requires testing against other subduction systems; however, in the numerical models presented in §6, we also observe that regional slab morphology has an effect on the development of sub-slab mantle flow and geometry of finite strain and these results could be of global applicability once applied to the wedge mantle.
- (vi) This model hypothesises that a sub-lithospheric decoupling “hidden” layer entrained beneath the subducting slab (figure 2.16d) triggers mantle flow in the wedge (Long and Silver, 2008). In this model, flow in the wedge is controlled by a competition between 2D flow induced by the plate convergence (parameterised as V_c) and trench migration (parameterised as V_t) rates. This model (figure 2.15b) predicts that subduction systems dominated either by convergence (low values of $V_{norm} = |V_t|/V_c$) or trench migration (high V_{norm}) record large delay times, whereas systems with intermediate values of V_{norm} exhibit weak splitting, which appears well supported by observations (Long and Silver, 2008).

- (vii) Numerical models (e.g., Honda and Yoshida, 2005; Behn et al., 2007; Honda, 2011; Morishige and Honda, 2011; Wirth and Korenaga, 2012) have tested the hypothesis of small scale convection affecting the wedge mantle flow. The outcomes show that wedge viscosity exerts the strongest control on the development of small-scale convection, in which gravitationally unstable crustal material beneath the arc is likely to induce trench-parallel mantle flow. Yet, mantle wedge viscosity is among the least well-constrained parameters, and more studies are needed to explore more in detail the role of small-scale convection in subduction settings (Long and Silver, 2008).
- (viii) The melt SPO model, namely the shape preferred orientation due to melt/hydrous phases or volatile, predicts a sharp transition in splitting behaviour from the fore-arc to the arc to the back-arc; as presumably, the region directly beneath the arc will be the richest in partial melt and thus show a trench-parallel signature. Experimental evidence exists and shows that when small amounts of partial melt are present they affect the geometry of olivine LPO in the surrounding matrix, in addition to providing an SPO effect (e.g., Zimmerman et al., 1999; Vauchez et al., 2000; Holtzman et al., 2003; Holtzman and Kendall, 2010). The major predictions this model makes are that the larger delay time should be recorded directly underneath the arc, and secondly, the recorded average delay time should correlate with the amount of partial melt and the volcanic production in each wedge. Nonetheless, the former is seldom been observed in natural settings and the latter is based on geodynamics interpretations of mechanisms which remain debated (Long, 2013).

2.4.6.2 Sub-slab mantle: hypothesis and models

The models used to characterise the mantle wedge are usually dependent from a larger set of parameters and can be more difficult to disentangle compared to the sub-slab case. However, global sub-slab splitting datasets demonstrate a combination of trench-parallel, trench oblique, and plate motion-parallel (i.e., the motion of the lower plate relative to the upper plate) trends in fast splitting directions globally (e.g., Peyton et al., 2001; Long and Silver, 2008; Long, 2009; Foley and Long, 2011; Di Leo et al., 2012; Lynner and Long, 2014b,a). Shear wave splitting behaviour beneath oceanic plates far away from trenches is relatively well understood because simple shear in the asthenosphere. In this case, the geometry of the seismic fast directions can be calculated as the vectorial difference between sub-lithosphere and plate movements (e.g., Behn et al., 2004; Conrad et al., 2007). However, the rich variety of sub-slab observations necessitate models for subduction-related anisotropy to explain the apparent lack of entrained flow beneath slabs.

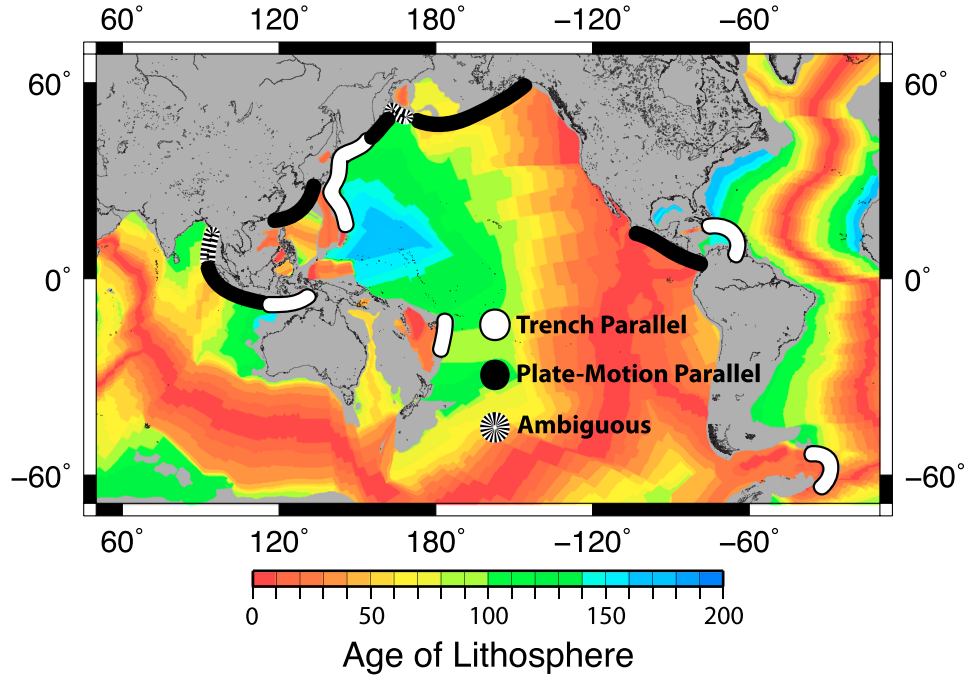


Figure 2.17: Relationship between plate age, implemented from the model of Müller et al. (2008b), and sub-slab fast directions from source-side side shear wave splitting measurements, taken from Lynner and Long (2013). Plate motion-parallel fast directions seem to correspond to younger plate ages.

- (i) The perfectly entrained 2D mantle flow to explain trench-parallel fast directions beneath the slab. In this scenario mantle material is entrained with the subducting slab, along with a small component of trench-parallel strain that gives rise to the observed trench-parallel fast directions; however, this model implies that mantle material beneath the slab is entrained without significant deformation and, thus, with no anisotropy in many subduction zones, which does not seem physically feasible (Long, 2013).
- (ii) The trench-parallel flow model was initially introduced by Russo and Silver (1994) and, then, further documented by Long and Silver (2008). This model aimed to provide an explanation to observed global patterns of trench-parallel flow in the sub-slab. From a global compilation of seismic anisotropy studies and plate kinematics (Long and Silver, 2008) a correlation emerged relating absolute trench retreat migration rate and shear wave splitting delay time. Trench-parallel flow in the sub-slab was then interpreted as due to large trench retreat generating substantial toroidal motion and efficient plate-mantle decoupling, consistent with numerical model calculations (Paczkowski et al., 2014a,b). On the other hand, the correlation between absolute trench retreat migration rate and shear wave splitting delay time could be the result of asthenosphere and slab properties affecting the flow in the

underlying mantle and not necessarily be related to high trench migration rates, as I will discuss later in chapter 5, 6 and 7.

- (iii) The age-dependent anisotropy model (Lynner and Long, 2014b) explains fast splitting directions and mantle flow regimes as determined by slab age and provides a relatively good match with natural observations (figure 2.17). This model predicts that trench-perpendicular flow beneath the subducting slab prevails underneath younger lithosphere (<95 Myr) where 2D entrained flow dominates and results in the entrainment of a thick layer beneath the sub-slab; whereas trench-parallel flow occurs in the sub-slab mantle when 3D return flow prevails producing a thin entrained decoupling layer beneath the base of an old subducting lithosphere (>95 Myr). While this observation suggests the strong influence of mantle flow on patterns of seismic anisotropy, which is very likely, if age would correlate strongly with the property of the sub-lithospheric layer entrained beneath the slab, then observations of a sharp LAB boundary would also correlate positively with oceanic lithospheric age; however, such inferred asthenospheric weaknesses are observed under a wide range of age for the oceanic lithosphere, and regardless of the tectonic setting (convergent vs divergent), see section 6. Secondly, in most subduction settings (e.g., Tonga, New Hebrides or Ryukyu) there is no positive correlation between surface motions, slab age and pattern of deformation in the upper-plate, and more uncertainties in models can arise depending on the choice of the reference frame.
- (iv) The “hidden” asthenospheric anisotropy model associated with the entrainment of either a thick layer with a moderately low viscosity or an extremely thin material with a larger viscosity reduction beneath the slab relative to the underlying mantle (figure 2.18a and b) (Long and Silver, 2009). In this case, the strain associated with the entrained flow does not lead to observable anisotropy and is, thus, “hidden”. This would require either a) a mechanism for which a “thick” layer of entrained asthenospheric flow directly beneath the slab is accommodated by diffusion creep, which does not develop lattice preferred orientation (LPO) and thus seismic anisotropy (e.g., Karato and Wu, 1993); or b) a very thin layer (≈ 10 km) of asthenospheric material being subducted with the slab so that any anisotropy is invisible to shear waves passing through it (e.g., Long and Silver, 2008). Both mechanisms require the advection of an asthenosphere that has distinct characteristics from the neighbouring mantle material (Long and Silver, 2009; Long, 2016). While it would be very difficult to transport and advect a whole thick layer of asthenosphere material, as the asthenosphere solely represents the depth range over which the temperature is close enough to the mantle solidus to cause a significant reduction in viscosity (Long, 2013). On the other end, a thin layer of entrained asthenosphere (figure

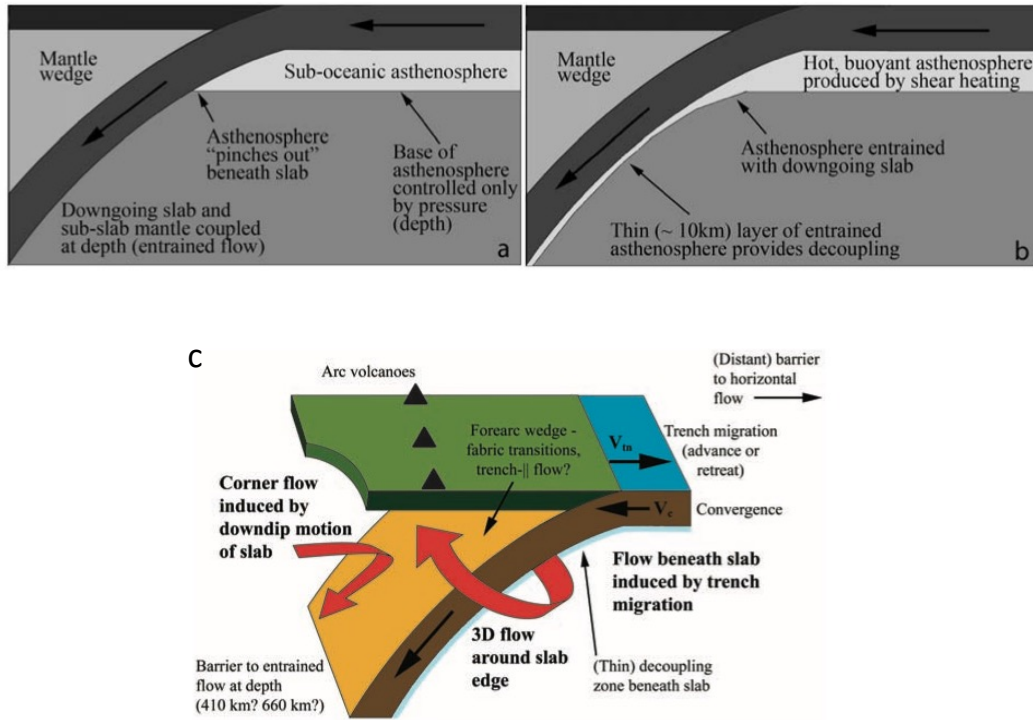


Figure 2.18: An illustration of the - hidden - asthenosphere anisotropy model for the cases of either a thickened or thinned asthenospheric layer, in a and b respectively, including possible forming mechanisms for the two layers, from Long and Silver (2009). In c, a cartoon illustrating the hypothesis of three-dimensional mantle flow being affected by the presence of a thin layer of entrained asthenospheric mantle material, reproduced from Long and Silver (2008). A thinner, more buoyant and less viscous sub-lithospheric layer (SLL) could allow for sub-slab decoupling affecting resultant patterns of seismic anisotropy.

2.18c) acting as a decoupling zone between the oceanic lithosphere and the sub-slab mantle, could favour the motion of the subducting slab to induce plate-motion parallel flow (Long and Silver, 2009). The viability of the “hidden” asthenosphere model necessitates a mechanism able to explain physical (and possibly compositional) different properties of the asthenosphere compared to the underlying mantle (Long and Silver, 2008). Different hypotheses currently exist that could support this model.

- (a) by invoking a plume-fed, chemically depleted asthenosphere being introduced in the sub-lithospheric mantle (Phipps Morgan et al., 1995). Numerical and laboratory experiments from Phipps Morgan et al. (2007) show that a thin ($\approx 10\text{--}30$ km) layer of hot, buoyant asthenosphere can be entrained beneath downgoing slabs during subduction.
- (b) numerical models (e.g., Turcotte and Schubert, 1982; Bevis, 1988; Nippres et al., 2007) have revealed that shear heating due to high strain at the LAB induced by the discordant motion of the oceanic lithosphere from the flow in the underlying mantle can provide substantially large amounts of strain in the sub-lithospheric mantle than in the mantle material directly underneath.

2.5 The lithosphere-asthenosphere boundary

The section aims to provide a brief summary of the current literature about 1) the nature of the lithosphere-asthenosphere boundary; 2) details about the structure of the asthenosphere and other inferred weaknesses; 3) collection of the geophysical pieces of evidence from previously published studies that report the presence of strong viscosity reduction in a thin layer at the base of the descending lithosphere beneath some of the segments of the Pacific basin; and 4) implications of a low-velocity zone on the evaluation of the global mantle flow.

2.5.1 The nature of the lithosphere-asthenosphere boundary

The lithosphere-asthenosphere boundary (LAB) is the boundary between the Earth’s strong lithosphere and the underlying asthenosphere. From a mechanical point of view, the lithosphere implies a concept of strength and relative permanence whereas the asthenosphere entails the notion of the “low viscosity layer” of the upper mantle (Anderson, 1995). Since the model of plate tectonics relies on the principle of a relatively coherent rigid lid translating over a weaker and more buoyant layer of asthenospheric material, the lithosphere-asthenosphere boundary becomes a first-order structural discontinuity that accommodates differential motions between the strong tectonic plates and the underlying weaker mantle. The LAB critically controls the degree of mechanical coupling between

the lithospheric plates and the flow in the mantle beneath and, therefore, exerts a primary role on plate tectonics. At this boundary corresponds a strong vertical variation in seismic waves speed or impedance, including changes in the strength and orientation of both radial and azimuthal anisotropy. This is known as the Gutenberg seismic discontinuity (G) and it is often used to define the depth of the LAB. The LAB is also defined as a rheological boundary between the Earth's conductive outer thermal boundary layer and the remaining convective mantle. In geodynamics, this boundary is generally equivalent to an isotherm of 1300° (Artemieva and Mooney, 2002).

Further interest in the study of the LAB was driven by the recognition of the LAB as the most extensive of plate boundaries on the planet. Such that, a body of geophysical studies carried out over a decade of seismic surveys (e.g., Li et al., 2000; Collins et al., 2002; Li et al., 2004; Kumar et al., 2005; Wölbern et al., 2006; Cammarano and Romanowicz, 2007; Heit et al., 2007; Kumar et al., 2007; Kustowski et al., 2008; Lebedev and Van Der Hilst, 2008; Nettles and Dziewoński, 2008; Dalton et al., 2009; Kawakatsu et al., 2009; Romanowicz, 2009; Rychert and Shearer, 2009; Fischer et al., 2010; Rychert et al., 2010; Schmerr, 2012; Sakamaki et al., 2013) revealed systematic variations in apparent lithospheric thickness and depth of the LAB correlating with different tectonic environments at global and regional scales (figure 2.19). Some of the correlations emerged from these studies and relevant for this thesis work include:

1. the evidence of a global low-velocity zone with a transition from a fast-seismic lid

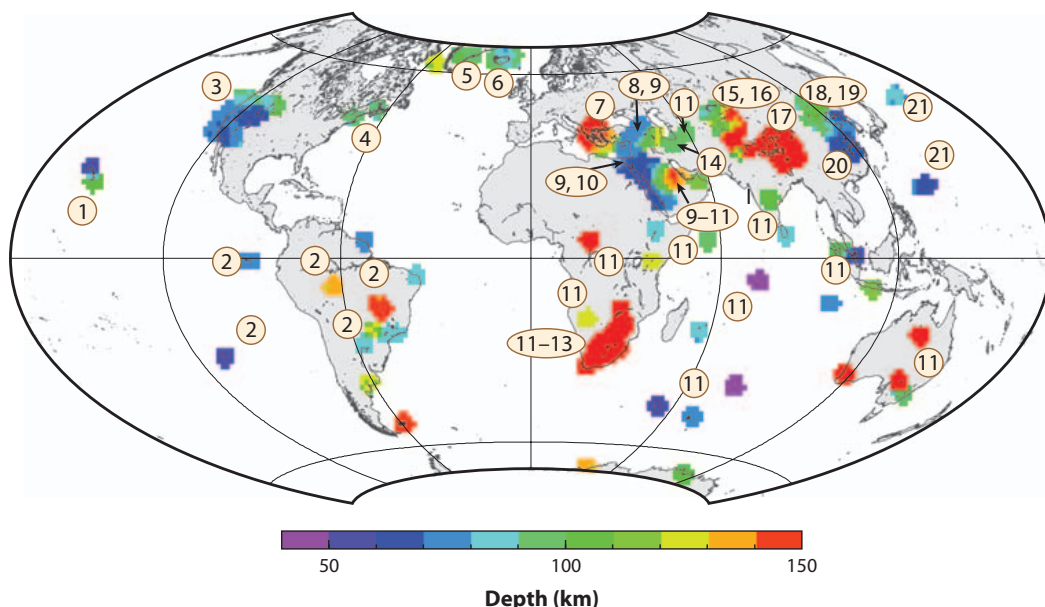


Figure 2.19: A global collection of measurements reporting a negative velocity drop in correspondence with the lithosphere-asthenosphere boundary (LAB), reproduced from Fitton (1987). LAB depths tend to relate to tectonic settings, although even if the colour scale ends at 150 km some phase arrivals were recorded up to 300 km beneath the surface.

to a slower deeper layer occurring in a depth interval of 50 to 220 km;

2. a possible depth of the interface that correlates with the tectonic setting and, in particular, shear velocity drops are recorded at mean depths of i) 94 ± 4 Km beneath Precambrian shells and platforms; ii) 81 ± 2 beneath Phanerozoic formation and orogenic belts; and iii) 70 ± 4 km beneath ocean basins;
3. a recorded regional seismic velocity drop of 5-10% across a sharp boundary no more than 20 km thick recorded in the depth range of 35 to 120 km beneath the ocean basins;

While the observed global velocity reduction could likely represent a widespread lithosphere-asthenosphere boundary under oceans and other tectonically active regions, many works (e.g., Jordan, 1978, 1988; Boyd, 1989; Pearson et al., 1995; Carlson et al., 1999; Griffin et al., 1999; Cammarano and Romanowicz, 2007; Rychert and Shearer, 2009; Abt et al., 2010) interpret this as an intra-lithosphere interface in cratonic regions and locate the LAB at a greater depth (figure 2.19).

The nature, depth and defining mechanisms of the seismic discontinuity associated with this boundary are still not completely understood and the LAB remains today a hotly debated topic in the academic world; for broader insights about this topic, I would therefore direct the reader toward existing excellent reviews and casework (e.g., Rychert et al., 2005; Thybo, 2006; Eaton et al., 2009; Fischer et al., 2010; Rychert et al., 2010; Schmerr, 2012; Sakamaki et al., 2013). Here, I would like to present hypotheses proposed over the last few decades to explain the origin of this boundary and associated discontinuity.

As mentioned earlier, when a shear wave propagates through a discontinuity, it may record a drop in velocity in response to a variation in the physical properties, such as density, bulk modulus, shear modulus, grain size, viscosity, temperature, as well as water, melt $\pm$ volatiles availability. At the LAB, post-glacial rebound investigations have confirmed a one to two orders of magnitude viscosity jump throughout the interface, (Mitrovica, 1996), consistent with geoid observations Moresi and Gurnis (1996).

Many studies invoke the primary role of some specific factors such as temperature, anisotropy and melt signatures over others to explain the nature of the lithosphere-asthenosphere discontinuity and associated low-velocity zone (e.g., Revenaugh and Jordan, 1991; Fischer et al., 2010; Schmerr, 2012; Beghein et al., 2014; Rychert, 2015; Clark and Leshner, 2017). However, if the temperature alone was the controlling factor for the origin of the LAB, then it would be expected a global gradual transition between the two layers, conversely many other studies, including high-resolution tomography images, report a sharp regional transition between the two blocks (e.g., Stern et al., 2015; Hawley et al., 2016). To the same end, a purely anisotropic interpretation for the discontinuity

would also not necessarily define the LAB because anisotropy could be frozen within the lithosphere from previous episodes of deformation (Rychert, 2015).

These observations led to the subsequent hypothesis that water or melt present in one or multiple channels could weaken the mantle and identify the base of the plate. The shallow dehydration that occurs concurrently with lithosphere formation at middle ocean ridge could relate with a rise in hydration with depth of the rocks moving away from the ridge. Additionally, the shearing due to the relative motion between plates and mantle could cause the concentration of melts in layers or lens and volatile storage at the base of the plate (Kawakatsu et al., 2009; Sakamaki et al., 2013). On the other end, the oceanic lithosphere is commonly predicted to be cold and there should not necessarily be melt at its base. Therefore, further investigations and integrated approaches are needed to elucidate the nature of this boundary, as it remains today a poorly constraint topic in the broad spectrum of the geosciences.

2.5.2 Constraints on the structure of the seismic low-velocity zone

The presence of a weak and extended layer of asthenospheric material lubricates the undersides of the tectonic plates allowing them to move freely over the mantle and eventually descend underneath. Radial viscosity variations are a key element of plate tectonics strongly influencing solid Earth’s dynamics (Mitrovica and Forte, 1997). Large-scale bulk viscosity values for the asthenosphere usually range between $\sim 0.5 \times 10^{18} - 10^{21}$ Pa·s (Mitrovica, 1996; Hu et al., 2016b; Freed et al., 2017), which correspond to 1-2 order of magnitude viscosity jump relative to the underlying mantle and are confined within a 100-300 km thick depth interval (Becker, 2017; Richards and Lenardic, 2018). These large-scale values are generally inferred from global numerical models, observations and laboratory-derived creep laws, and explained by invoking the role of temperature and pressure; however, many uncertainties remain on the choice of appropriate rheologies and parameters within. Furthermore, the ability to distinguish between a thin layer of very low viscosity and a thick layer of greater viscosity can be hindered by the no uniqueness relating viscosity and thickness of an LVZ (Cathles, 1975; Richards and Lenardic, 2018; Carluccio et al., 2019b), such as that the resulting inferred values may be conjectured mean values averaged over the entire thickness of the asthenosphere and do not reflect the detailed structure of the low-velocity zone. Other difficulties in imagining the structure of the asthenosphere also arise from the trade-off that exists between the resolving power and lateral extent of seismic and other modelling studies. This can hinder the ability to distinguish thin layers embedded within the asthenosphere, as well as their lateral extension. To the same end, laboratory experiments suggest that the viscosity profile of the asthenosphere could be less homogeneous than what previously envisaged (Van der

Meijde et al., 2003). More recently, a range of geophysical studies has brought considerable attention to the detailed structure of the asthenosphere at subduction zones (e.g., Revenaugh and Jordan, 1991; Gaherty et al., 1996, 1999; Bagley and Revenaugh, 2008; Bagley et al., 2009; Kawakatsu et al., 2009; Schmerr, 2012; Naif et al., 2013; Stern et al., 2015; Hawley et al., 2016; Rychert et al., 2018). These studies reveal a significant and sharp drop in seismic wave velocity and an increase in electrical conductivity underneath the base of a number of descending segments of the Circum Pacific lithosphere (figure 2.21). These studies suggest that local and regional asthenospheric configurations may lead to up to ≈ 4 orders of magnitude viscosity reduction due to enhanced melt content or hydrated $\pm$ volatile phases occurring within a no more than 50 km depth interval which is thought to have lubricated base of a subducting plate. Such asthenospheric configurations would be consistent with large-scale and seismic anisotropy models as long as their lateral extent remains limited compared to the rest of the plate (Becker, 2017).

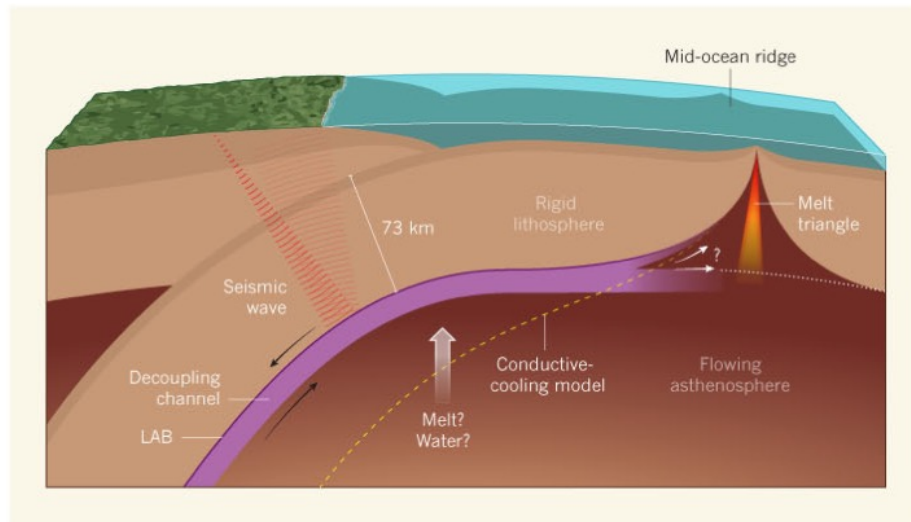


Figure 2.20: A schematic cartoon representing the Pacific plate subducting beneath New Zealand taken from Rychert (2015). At this locality, the Pacific plate is thinner than what predicted by the half-space cooling method. Recent research (e.g., Stern et al., 2015) suggests a sharp LAB underlined by 10 km thick channel enriched with hydrated, melt and/or volatile phases that decouples plate movement from mantle flow underneath.

In the asthenosphere, the peridotite with minor carbon and hydrogen (lherzolite-[C+H⁺O] system) at a temperature of ≈ 1430 °C can be partially molten (Kamenetsky et al., 1998; Green et al., 2001; Schubert et al., 2001; Scoppola et al., 2006). Deformation experiments on partially molten olivine aggregates and lherzolite rocks show that both water and melt contents of an asthenospheric rock can significantly diminish its viscosity, as well as induce large shear deformation, which promotes the ability of melt to segregate into melt-rich regions (Kohlstedt et al., 1995; Holtzman et al., 2003; Zimmerman and Kohlstedt, 2004). Melt can segregate into lenses and channels whilst undergoes large

strain deformation. Thus, inside the asthenosphere, there may be one or more melt-rich layers, with a combined vertical thickness of 5–30 km, able to contain ≈ 10 times the amount melt of adjacent rocks and therefore expected to be significantly weaker than the neighbouring rocks. The capacity of melt to segregate may allow such heterogeneities to become shear zones. Some of these shear zones are preserved in the geological record and observed in exhumed mantle rocks of ophiolite complexes (Kelemen and Dick, 1995; Dijkstra et al., 2002) or mantle xenoliths that record a shear achieved at the lithosphere–asthenosphere interface (Kennedy et al., 2002). There is, therefore, geological evidence in support of flow in the upper mantle undergoing decoupling at the base of the lithosphere due to the relative motion of the two blocks and resulting in melt ponding at the interface (Russo and Silver, 1996; Doglioni et al., 1999; Bokelmann and Silver, 2000; Kawakatsu et al., 2009).

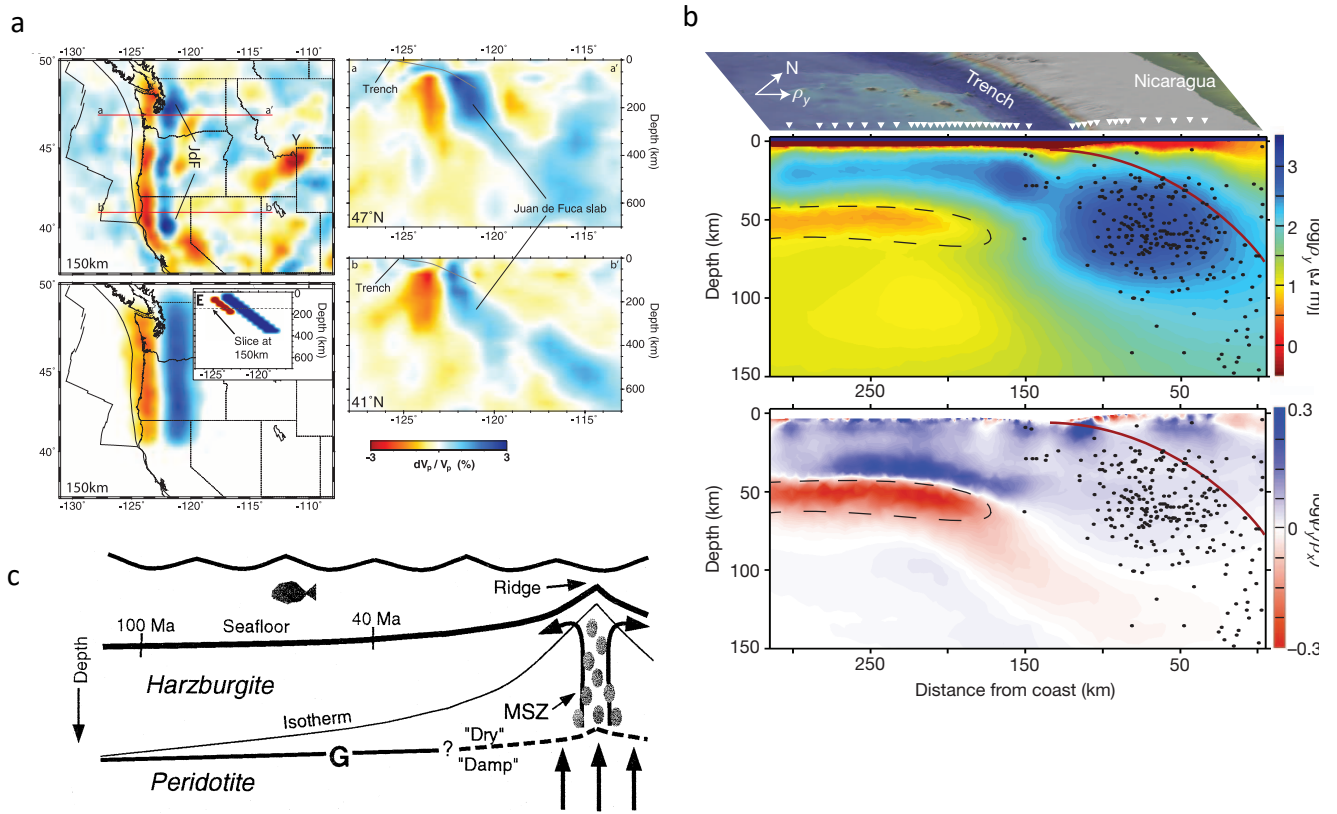


Figure 2.21: Selected geophysical observations reporting an abrupt seismic velocity drop or electric conductivity increase underneath the base of the subducting Juan the Fuca and Cocos plates, reproduced from a Hawley et al. (2016) and b Naif et al. (2013), respectively. These observations indicate a channelised base of the subducting oceanic lithosphere. In c, a summarising cartoon displaying a petrological mechanism that could trigger viscous layering at the LAB, taken from Gaherty et al. (1999).

2.5.2.1 Geophysical observations for a channelised base of a subducting plate

The particular structure of the lithosphere has decade after decade continued to attract academic interest because of the fundamental role it plays in the plate tectonic style of mantle convection we observe on our planet. More recently, advances in computation resources, resolution and power of geophysical methods have helped to provide further geophysical evidence for the regional occurrence of thin and weak sub-lithospheric channels ponding at the lithosphere-asthenosphere interface of many subduction zones in the Pacific basin (figures 2.21 and 6.1). I'll below compile and describe a collection of 11 geophysical studies. These report the presence of a channelised LAB in many regions, including:

- the North Island of New Zealand where the Pacific tectonic plate descends underneath the Australian plate. Here the analysis of seismic reflection images reports a V_p drop of 10 ± 4 %, within a “low-seismic-wave-channel” of 8-12 km thick at a depth of 100 km corresponding to the LAB (Stern et al., 2015);
- several zones in the Pacific beneath Tonga, New Hebrides, Philippine sea, Kamchatka, Cascadia and Cocos ridge. These were detected by seismic surface waves that identify a reflection coefficient drop of about 3.5% with values up to 10.3% within one or multiple melt-rich layers (Schmerr, 2012);
- underneath the base of the Pacific lithosphere where it subducts underneath Tonga, Sanighe and Izu-Bonin. Revenaugh and Jordan (1991) employ ScS reverberation techniques and record a sharp negative SS precursor ($\delta v_s/v_s \sim 6\%$) across the G discontinuity. For the author, such a velocity drop requires the presence of partial melt, possibly catalysed by a flux of hydrated and volatile phases which are thought to have preferentially decreased viscosity over density. Revenaugh and Jordan (1991) suggest that the G discontinuity could be best explained as the seismic expression of the spinel→garnet facies boundary (figure 2.21c). Additionally, strong corner flow in the asthenosphere could contribute to the upward migration of melt and anisotropy from compositional layering (Ribe, 1989) and further accentuate the strength of the reflector;
- a Tonga-Hawaii channelized corridor bounded at 68 km by a strong G discontinuity. The analysis by Gaherty et al. (1996) of S, SS, and SSS body waves and R1 and G1 surface waves, as well as reflectivity profile of ScS reverberations from the work of Revenaugh and Jordan (1991) lead to a $\sim 6\%$ velocity drop occurring over an interval of less than 30 km thick, consistent with what also reported by (Revenaugh and Jordan, 1991). The discontinuity was, then, interpreted as the fossilised lower boundary of a melt separation zone resulted from a chemical transition caused by decompression melting at spreading centers (see Gaherty et al., 1996; Hirth and Kohlstedt, 1996);

- across eight ray paths across the circum-Pacific corridor (Bagley and Revenaugh, 2008). Bagley and Revenaugh (2008) findings reveal a shear wave impedance drop at the LAB of approx. 8-10% focused in a 30 km interval at the base of the Pacific plate;
- Gaherty et al. (1999) explores the characteristics of the G discontinuity underneath the Philippine sea combining two data sets. The first dataset includes the travel times and impedance contrasts across mantle discontinuities using ScS-reflectivity profiles. The second compiles observations of frequency-dependent travel times of three-component (S, sS, SS, sSS, SSS, Sa) and surface (R1, G1) waves, which constrain the anisotropic velocity structure within discontinuities. Both Bagley and Revenaugh (2008) and Gaherty et al. (1999) works describe the G as a sharp discontinuity significantly deeper beneath the western younger Philippine Sea plate than in the older counterpart, located at 89 and 68 km, respectively, implying that thermal cooling model alone does not control the thickness of the lid as it would require a thinner lid with decreasing age and this is not observed;
- an entrained thin layer of asthenospheric material is observed all the way down to the upper-lower mantle transition zone in northeast China. Wang et al. (2020) employ the P-wave receiver function technique and suggest a melt and/or volatiles rich layer following the slab all the way to the 660 km discontinuity;
- the Philippine sea and Pacific plate close to Japan by Kawakatsu et al. (2009). They perform receiver function analysis of borehole broadband ocean base seismic data and record 7-8% shear waves velocity drop that translates into a melt fraction of ~ 3.5 to 4.0% , which appear consistent with a viscosity drop at the LAB even greater than 10^3 . The author best-fit model to explain this large viscosity drop employs a layered model for the structure of the asthenosphere;
- Bagley et al. (2009) re-examine the regions underlying the ocean-ward base of the Honshu slab sampled by Obayashi et al. (2006) with multiple ScS reverberations and SH-polarized shear phases. They report a velocity drop in an interval up to 50 km of partial melt. They further attribute the melt layer to the combined effects of water and temperature and estimate a maximum temperature anomaly of $\sim 150^\circ\text{C}$ resulting in a maximum olivine water content after melting of 0.100 wt.% and lower viscosity near the slab;
- beneath newly formed lithosphere at the Juan de Fuca and Gorda Ridges. Rychert et al. (2018) use tomography and scattered wave imaging (S-to-P receiver functions) to study the lithosphere-asthenosphere boundary and report the presence of a sharp base of the oceanic plate;

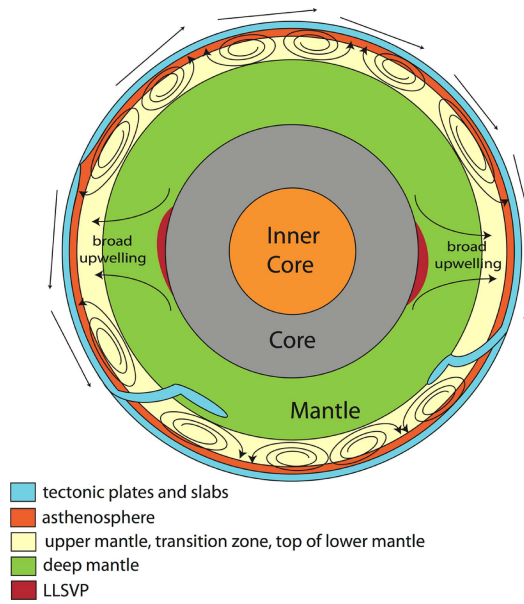


Figure 2.22: A schematic cartoon of mantle flow being arranged into two large updrafts underneath Africa and the Pacific plate. Asthenosphere properties appear responsible for the contemporary long-wavelength expression of the mantle circulation, taken from King (2016).

- underneath the Juan de Fuca plate high-quality tomography images display buoyant asthenosphere accumulating beneath the subducting slab (Hawley et al., 2016). These images appear compatible with a ~ 10 km thick layer experiencing up to 4 order of magnitude viscosity drop relative to the surrounding mantle material; and
- the Cocos plate in the Middle America trench offshore of Nicaragua where a magnetotelluric study reveals a high conductivity layer up to 30 km thick and confined to depths of 45 to 70 km (Naif et al., 2013). Because partial melt is stable at these depths in a warm damp mantle, they interpret the conductor to be a partially molten layer capped by an impermeable frozen lid that is the base of the lithosphere.

For more detailed information on each of these particular observations and techniques employed, I would direct the reader to the individual papers cited above. Additionally, the final research chapter of this thesis, §6 and figure 6.1, compile and plot the geographic location of these observations, and provides particular information about the nature of each of these subduction zones.

2.5.3 Possible implications for global mantle flow

The presence of an extended low-velocity asthenosphere in the mantle stabilizes large-scale horizontal flow by allowing the lithospheric tectonic plates to move freely and limiting the amount of viscous dissipation over geologic time (King, 2016). King (2016) shows the results of two end members scenario for the modelling of the Earth’s radial viscosity structure with and without an asthenosphere. The former case leads to a large-scale updrafts convection pattern (degree 1-2 long wavelength) consistently with the recent tomographic images of the “underworld” and the current idea of the organization of the deep mantle flow (Zhang et al., 2010; French and Romanowicz, 2015). The latter case exhibits the development of multiple (i.e., ~ 20) buoyant upwellings. The calculation of long-wavelength pattern of convection with one or two large upwelling(s) simulates the development two low shear velocity provinces (LLSVPs) which are (currently) seen to dominate the lower mantle flow and located at the base of the mantle beneath Africa and the South Pacific (figure 2.22). The LLSVPs are thought to originate from the remnants of compositionally distinct subducted slab material or material left over from a primordial magma ocean at the base of the mantle (Labrosse et al., 2007) piled up into two large-scale upwellings. Large-scale modelling studies (e.g., King and Hager, 1990; Labrosse et al., 2007; Zhang et al., 2010) have provided a confirmation of the primary role of asthenosphere structure and rheology on the development of the plate tectonic regime and the organisation of the global mantle flow as we know it today on Earth (Höink et al., 2012; Anderson and King, 2014; Anderson and Natland, 2014; King, 2015). Despite the large-scale implications that the low viscosity layer of the upper mantle has proven to exert on the reorganisation of mantle flow, the role of a thin and even weaker boundary layer in the top asthenospheric layer has remained unexplored to date. Such thin channels could serve as a decoupling layer between the motion of the strong tectonic plate and the underlying weaker mantle by facilitating plate motion and enhancing vigorous convection into the mantle; hence significantly contributing to characterise mantle and lithosphere modes of interactions over geologic time. The very existence of thin boundary layers at the LAB and, possibly at other discontinuities, could provide the perturbation necessary for the emergence of the plate tectonic regime (Bellahsen et al., 2005). It is, then, only natural to question what insights into the Earth’s interior dynamics could be achieved by incorporating such thin layers in future 3D spherical convection models and test boundary layer prediction in driving radial mantle circulation.

3. Chapter 3: Methodology

3.1 Introduction

Computational modelling of geophysical processes consists of at the least three principal stages. These are: i) the definition of an idealisation of the physical problem of interest, ii) the choice of an appropriate mathematical model to reproduce the physical problem; and iii) the computational experiments (Peiró and Sherwin, 2005).

This research project therefore commences with the identification the problem under investigation, as such “What is the role of a weak and thin decoupling layer on subduction surface motion, induced mantle flow and sub-slab anisotropy?”. The subsequent stage involves the choice of an appropriate set of governing equations to represent a simplification of the natural process. In geodynamics, geophysical and geological processes are widely described by a typical continuum mechanics approximation. This approximation is defined by a set of balanced equations - the conservative and constitutive equations - which together represent the governing equations of the problem of interest. The conservative equations or Navier–Stokes equations provide an accurate description of fluid dynamics. These are solved in combination with the constitutive relationships which define the behaviour of a geological medium under deformation. Then, it is necessary to select an appropriate model setup to represent an idealisation of the physical problem with a suitable set of initial and boundary conditions. The final step involves computational calculations, as well as, code sensitivity and optimization tests, necessary to proceed to the solution of the problem, perform the computer simulations and outputs analysis. This stage includes also the choice of variables to be tested, measured and analysed to address the problem under investigation.

In this thesis work, I have focused mainly on the study of viscosity, thickness and width of the weak layer (the full scheme is available in each of the research chapters). It is worth noting that it is often necessary to introduce simplified assumptions to reduce the complexity of the computations and costs associated with cluster computing systems. It might also not always be possible to maintain the fidelity of the idealisation, since some geological processes occurring in the Earth’s interior are not completely understood, such as those involving complicate pressure and temperature conditions where direct measurements’ are limited in both time and space scales. However, extraordinary efforts have been made by the computational geosciences community to bridge the gap between observations and numerical results, and provide a robust computational framework to answer many unresolved questions in the geosciences.

This chapter offers a brief overview to the reader on the concepts of fluid dynamics that geodynamicists apply to model the evolution of the Earth’s interior over a geological time scale. Here, I will be introducing the concepts of conservation and constitutive laws, as well as associated boundary conditions (§3.2.1, 3.2.2 and §3.3, respectively). This is followed by a description of the numerical framework, Underworld2, used in this work

to derive such equations and perform the computer simulations (§3.4). Underworld2 is a 2-D and 3-D parallel thermo-mechanical finite element code, combined with a marker and cell approach to simulate large-scale geodynamic processes. This chapter continues by describing the non-dimensionalisation rule implemented throughout this thesis work. While a throughout explanation of the model set-up is provided in each research chapters, the chapter concludes with the illustration of the numerical results of the terminal velocity of a sphere falling into a less viscous fluid. This is used to further prove the reliability of thesis calculations with basic physics principles and previous work.

3.2 Governing equations

In computational geodynamics, geophysical processes are generally described by a set of conservation and constitutive balanced equations. The conservation laws are based on a continuum mechanics approximation that considers Earth’s materials as continuous geological media. Such approximation implies, on both macroscopic and microscopic scales, that rocks do not contain voids, or if gaps exist, then these are filled with some substances. As rocks characteristics may vary, then a continuous implementation is necessary at every geometrical point in space and time. A continuum mechanics description is widely used to describe fluid dynamics, as well as mantle rocks behaviour, since over a geological timescale they act as a viscous fluid (see §3.2.2), despite being known that mantle compositional and structural heterogeneous exist at various scales (Ricard, 2007). The conservation laws are below shortly summarised. Further details can be found in the existing literature, particularly in Ricard (2007); Ismail-Zadeh and Tackley (2010) and Turcotte and Schubert (2014), and reference therein.

3.2.1 Conservation equations

The mass, momentum and energy conservation equations generally adopted by the geodynamics community are described in this subsection.

3.2.1.1 Mass conservation

The Eulerian form of the mass conservation relationship can be expressed as:

$$\frac{\partial \rho}{\partial t} + \nabla \cdot (\rho \mathbf{v}) = \frac{\partial \rho}{\partial t} + \frac{\partial (\rho v_i)}{\partial x_i} = 0 \quad (3.1)$$

Equation (3.1) defines the mass balance within an elementary volume during the displacement of a continuous media where ρ , $\mathbf{v}$ and $\rho \mathbf{v}$ are the material density, the velocity vector and mass flux, respectively. Eulerian and Lagrangian formulations can be applied

to the conservation laws depending by the nature of the geometrical point of observation from which the equation is written.

The incompressible form of the mass conservation equation assumes that the Lagrangian particles have constant density and no density variation are observed by the geometrical observation point $D\rho/Dt = 0$. If also pressure and temperature variations are limited, and no other processes that may lead to volume changes occur in the geological media, the mass balance equation assumes the form of the continuity equation:

$$\nabla \cdot \mathbf{v} = 0 \quad \text{or} \quad \rho \frac{\partial v_i}{\partial x_i} = \sum_{i=1}^3 \frac{\partial(\rho v_i)}{\partial x_i} = 0 \quad (3.2)$$

The continuity equation describes “the conservation of mass during the displacement of a continuous medium” and is widely used in computational geodynamics; however, it is worth nothing that it can represent a significant numerical simplification, for example in the case of modelling whole mantle convection processes (Tackley, 2008)).

3.2.1.2 Momentum conservation

The equation of the conservation of momentum is a differential equivalent to the traditional Newton’s second law of motion (i.e., $\mathbf{f} = m\mathbf{a}$). Momentum variations are identified by “balancing momentum changes with body forces acting in the volume Ω and to the surface forces acting on its surface Σ ” (Ricard, 2007). The total momentum is generally expressed as:

$$\int_{\Omega} \rho \mathbf{v} dV \quad (3.3)$$

Momentum changes are caused by: i) advective transport of momentum through the surface Σ ; ii) forces acting on the surface; and iii) internal body forces.

Continuous body mode of deformation depends on the balance between internal and external forces (e.g. Karato et al., 2001; Ribe, 2001; Turcotte and Schubert, 2014). For a continuous medium in the gravity field, the Eulerian relationship between these forces and deformation can be expressed as:

$$\rho \frac{D\mathbf{v}}{Dt} = \nabla \cdot \boldsymbol{\sigma} + \rho \mathbf{g} \quad (3.4)$$

or

$$\rho \frac{Dv_i}{Dt} = \rho \left(\frac{\partial v_i}{\partial t} + v_j \frac{\partial v_i}{\partial x_j} \right) = \frac{\partial \sigma_{ij}}{\partial x_j} + \rho g_i, \quad (3.5)$$

being i, j , the spatial directions following the Einstein summation convention, ρ the fluid density, g_i the gravitational acceleration, v_i the velocity, x_j the spatial coordinate and σ_{ij} is the total stress tensor.

The relationship between pressure and total stress tensor, $\sigma_{ij} = -P\delta_{ij} + \tau_{ij}$ introduces pressure into the momentum equation (Gerya, 2009). The total stress field σ_{ij} can be decomposed into a static part, $P = -\sigma_{kk}/3$, “the hydrostatic pressure”, which exists even if the fluid is at rest; a deviatoric stress term τ_{ij} that indicates a variation of stress from the hydrostatic pressure stage; and the Kronecker’s delta, δ_{ij} , which equals 1, if $i = j$ and 0, if $i \neq j$.

The Navier-Stokes equation can be obtained by combining the moment equation (3.4) and the relationship of the total stress tensor, as follow:

$$\frac{\partial \tau_{ij}}{\partial x_j} - \frac{\partial P}{\partial x_i} + \rho g_i = \rho \frac{Dv_i}{Dt}. \quad (3.6)$$

where $\frac{Dv_i}{Dt}$ refers to the velocity vector substantive time derivative.

The right-hand-side term in equation (3.6), can be neglected since geological rocks behave as viscous fluids over geologic time. Thus, for an incompressible, constant-viscosity highly viscous fluid, equation (3.6) simplifies to:

$$\rho \frac{\partial v_i}{\partial t} + \rho v_j \frac{\partial v_i}{\partial x_j} = \eta \frac{\partial^2 v_i}{\partial x_j \partial x_j} - \frac{\partial P}{\partial x_i} + \rho g_i = 0. \quad (3.7)$$

Equation (3.7) represents the original form of the Navier–Stokes equations. This equation is solved for pressure and velocity. For further details about the derivations of the conservation’s laws in a geodynamics perspective, I will direct the reader to the works of Landau and Lifshits (1959); Ismail-Zadeh and Tackley (2010), Ricard (2007) and reference therein.

3.2.1.3 Energy conservation

The Energy conservation equation counts the gain and lose in energy occurring in a volume per unit of time. If all the parameters are constant (homogeneous medium), we can write the conservation of energy in a general form as:

$$\frac{DT}{Dt} = \frac{\partial T}{\partial t} + \mathbf{v} \cdot \nabla T = k \nabla^2 T + H \quad (3.8)$$

$$\text{or} \quad (3.9)$$

$$\frac{\partial T}{\partial t} + v_j \frac{\partial T}{\partial x_j} = \frac{\partial}{\partial x_i} k \frac{\partial}{\partial x_i} T + H = k \sum_i \frac{\partial^2 T}{\partial x_i^2} + H \quad (3.10)$$

Where T is temperature, H refers to external sources of heat (e.g., radioactive, adiabatic, shear and latent) and the thermal diffusivity $\mathbf{k}$ is:

$$\mathbf{k} = \frac{k}{\rho c_p}, \quad (3.11)$$

where k is the thermal conductivity, which is a tensor, but it is usually approximated as a scalar in geodynamics, and c_p is the heat capacity. From the Fick's law the heat flux $\mathbf{q}$ can be expressed as:

$$q_i = -k \frac{\partial T}{\partial x_i}, \quad (3.12)$$

3.2.2 Constitutive equations

Rheology is the study of how matter flows and deforms. The constitutive laws describe the intrinsic properties of matter determining its behaviour and relate kinematic quantities (e.g., displacement, velocity, deformation and fluxes) to dynamic quantities (e.g., forces and stresses), (Carluccio, 2016; Malkin and Isayev, 2017). Rocks can deform either elastically, viscously or plastically or in a combination of them (figure 3.1).

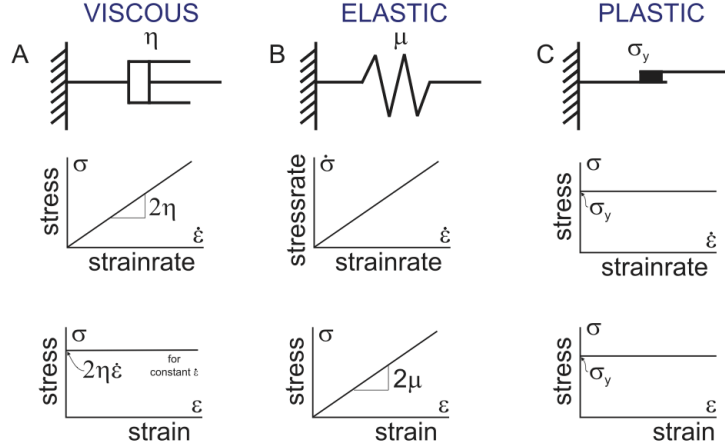


Figure 3.1: Schematic sketch representing rocks' (A) viscous, (B) elastic or (C) plastic behaviours under deformation, taken from by van Hunen et al. (2011) and originally from Ranalli (1995). From top to bottom are illustrations of A, B, C mechanical equivalents, stress:strain rate and stress:strain rate responses.

The ratio between viscosity, η , and shear modulus, μ , is the Maxwell time, τ_M ($\tau_M = \frac{\eta}{\mu}$). The Maxwell time defines whether a material deforms elastically or viscously; however, both behaviours can occur simultaneously. For instance, in the case of mantle convection viscous deformation prevails for time larger than the Maxwell time, conversely elastic behaviour dominates for lower values, for example during earthquakes.

Mantle typical values for viscosity and shear modules correspond to $\eta \approx 10^{21}$ [Pa...] and $\mu \approx 10^{10}$, respectively, which reflects into a Maxwell relaxation time of approximately 10^3 yrs. The mantle has a relatively larger τ_M than the lithosphere and thus, it behaves mainly viscously over a geologically large time scale (from millions to billions of years). Viscous constitutive laws are, then, often used to model mantle behaviour. On the contrary, the lithosphere has a smaller characteristic time, and an elastic rheology approximation should be used to model lithosphere deformation, especially in case of a short time scale (days to seconds) (Ricard, 2007). Ultimately, a plastic approximation describes rocks having finite strain (e.g., faults or shear zones). As such, plastic yielding occurs when differential stresses overcome the rock stress threshold value. Temperature, pressure and geochemical composition substantially affect the strength of a body of rock under deformation, and all these factors can largely vary through the inner regions of a planet. In spite of all efforts, more detailed rheology approximations are still needed for the modelling of specific Earth's regions and complex geological settings.

3.2.2.1 Newtonian rheology

Newtonian rheology is defined by the linear relationship relating the stress and strain rate second invariants. This relationship can be expressed by taking into account the total stress and linearly relating the deviatoric stress to the rate of deformation, or strain rate, $\dot{\epsilon}_{ij}$ (Gerya, 2009). As shown in the following steps:

$$\tau_{ij} = 2\eta\dot{\epsilon}_{ij} + \delta_{ij}\eta_{bulk}\dot{\epsilon}_{kk} \quad (3.13)$$

or

$$\tau_{ij} = 2\eta\left(\dot{\epsilon}_{ij} - \frac{1}{3}\delta_{ij}\dot{\epsilon}_{kk}\right) + \delta_{ij}\eta_{bulk}\dot{\epsilon}_{kk} \quad (3.14)$$

where η_{bulk} and $\dot{\epsilon}_{kk}$ are the bulk viscosity and strain rate, respectively, and η is the shear viscosity. The strain rate is defined as:

$$\dot{\epsilon}_{ij} = \frac{1}{2}\left(\frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i}\right) \quad (3.15)$$

where v_i is the fluid velocity vector. The strain rate $\dot{\epsilon}_{ij}$ is usually split into a deviatoric ($\dot{\epsilon}'_{ij}$) and an isotropic ($\frac{1}{3}\dot{\epsilon}_{kk}\delta_{ij}$) parts, as follow:

$$\dot{\epsilon}'_{ij} = \dot{\epsilon}_{ij} - \frac{1}{3}\dot{\epsilon}_{kk}\delta_{ij} \quad (3.16)$$

For the case of an incompressible fluid with linear Newtonian rheology, the stress and strain rate can be related as follow:

$$\tau_{ij} = 2\eta\dot{\epsilon}_{ij}^{vis} \quad (3.17)$$

or

$$\dot{\epsilon}_{ij}^{vis} = \frac{1}{2\eta}\tau_{ij}, \quad (3.18)$$

from which, for any i and j , the definition of viscosity follows:

$$\eta = \frac{\tau_{ij}}{2\dot{\epsilon}_{ij}^{vis}} \quad (3.19)$$

In this case, we can compute the second invariant of the deviatoric stress and the strain rate as:

$$\eta = \frac{\tau_{II}}{\dot{\epsilon}_{II}}, \quad (3.20)$$

where

$$\tau_{II} = \left(\frac{1}{2}\tau_{ij}\tau_{ij} \right)^{(1/2)} \quad \text{and} \quad \dot{\epsilon}_{II} = \left(\frac{1}{2}\dot{\epsilon}_{ij}\dot{\epsilon}_{ij} \right)^{(1/2)}. \quad (3.21)$$

While in geodynamics, it is typically assumed that the material viscosity is isotropic and scalar, η is a fourth-order viscosity tensor relating stress to strain-rate as follow:

$$\tau_{ij} = 2\eta_{ijkl}\dot{\epsilon}_{kl}^{vis} \quad (3.22)$$

In a Newtonian viscous rheology, η is independent of the state of stress; however, it should be noted that stress can be strongly dependent on other factors, such as temperature, grain size, confining pressure and fluid content (Gerya, 2009), among others. For a more detailed description of the equations of motion for a viscous fluid, I will direct the reader to the traditional fluid mechanics textbooks (e.g., Landau and Lifshits, 1959).

3.2.2.2 Viscous Creep rheology

Over geologic time the mantle behaves as a highly viscous fluid (see §3.2.2) and its behaviour depends largely on temperature and pressure. Specifically, at high temperature (i.e., many rocks' melt temperatures) and high-stress conditions materials behave as non-Newtonian fluids with dislocations and atoms becoming fairly mobile and resulting in creep. Conversely, at a very low-stress and low temperature conditions, the diffusion process dominates and materials behave as Newtonian fluids with a viscosity that depends

exponentially on pressure and inversely from absolute temperature. Hence, the general viscous creep form of the equation relating the viscous stress and the strain is described as follow:

$$\tau_{ij} = C^{\frac{1}{n}} \dot{\epsilon}_{ij} \dot{\epsilon}_{II}^{\frac{1-n}{n}} \quad (3.23)$$

where C is a proportionality factor defined from the thermodynamic conditions and n is a power-law exponent. When $n = 1$, equation (3.23) describes the behaviour of a Newtonian fluid with $C/2$ being the fluid's viscosity dependent only from temperature and pressure, whereas when $n > 1$, equation (3.23) indicates a non-Newtonian fluid behaviour.

The proportionality factor, C , in equation (3.23) can be represented as:

$$C(T, P) = C^* d^m \exp \left(\frac{E + PV}{RT} \right) \quad (3.24)$$

where T is the absolute temperature, P is pressure, C^* is the proportionality factor that does not depend on temperature and pressure, E is the activation energy, V is the activation volume, R is the universal gas constant, and d is the grain size. Grain size is inconsequential for dislocation creep regime (i.e., $m = 0$); however, for diffusion creep may be important and m usually ranges between 2 and 3. The non-Newtonian behaviour is on average characterized by $n=3.5$ (Ismail-Zadeh and Tackley, 2010).

The temperature-pressure dependence is particularly relevant for the better understanding of convection in transporting heat within the mantle and throughout the lithosphere. Both dislocation and diffusion creep mechanisms are thermally activated processes and play a control over thermal convection in the mantle and some lithospheric processes. In addition to these mechanisms, when stresses are high enough the "low-temperature (Peierls) plasticity" can become a significant process to bypass the diffusion-controlled climb of edge dislocation - which is the limiting process of the dislocation creep regime - and force dislocations through obstacles (Ismail-Zadeh and Tackley, 2010). In this case, the dependence of the strain on the stress can be expressed as:

$$\dot{\epsilon} = A \exp \left[-\frac{H_0}{RT} \left(1 - \frac{\tau_{II}}{\tau_P} \right) \right] \quad (3.25)$$

with τ_P being the Peierls stress usually ranging between 2–9 GPa. Further details about concepts of computational geodynamics and rock rheology can be found in Ranalli (1995); Kameyama et al. (1999); Ismail-Zadeh and Tackley (2010), and Karate and Marathe (2008).

3.2.2.3 Visco-elastic rheology

A visco-elastic geological medium behaves elastically on a short time scale and viscously on a long time scales. This type of rheology can be modelled with a combined viscous and elastic law, hence being particularly relevant as viscous creep mechanisms can also relax elastic stresses in the lower lithosphere. The visco-elastic (Maxwell) rheology is widely used computationally to reproduce lithospheric processes. In this case, the elastic strain ε_{ij} and the deviatoric stress τ_{ij} , consistent with Hooke law of elasticity, can be related as:

$$\tau_{ij} = \mu \varepsilon_{ij}, \quad (3.26)$$

Assuming Newtonian formulation, the relationship between the fluid strain rate and the stress becomes:

$$\tau_{ij} = 2\eta \frac{\partial \varepsilon_{ij}}{\partial t}, \quad (3.27)$$

The Maxwell model for a viscoelastic geological material assumes that its strain rate is a superposition of the elastic and viscous strain rates (Ismail-Zadeh and Tackley, 2010), namely:

$$\frac{\partial \varepsilon_{ij}}{\partial t} = \frac{\tau_{ij}}{2\eta} + \frac{1}{\mu} \frac{\partial \tau_{ij}}{\partial t} \quad \text{or} \quad \left(1 + 2tr \frac{\partial}{\partial t}\right) \tau_{ij} = 2\eta \frac{\partial \varepsilon_{ij}}{\partial t} \quad (3.28)$$

where $tr = \eta/\mu$ is the aforementioned viscoelastic relaxation (or Maxwell relaxation, τ_M) time.

3.2.2.4 Plasticity

Plastic rheology describes the relationship between stress and deformation leading to permanent deformation of geological media, which is also known as a non-recoverable strain. This occurs in rocks when they are subjected to differential stresses larger than their yield-stress and fail. As such, plastic behaviour is generally found in the form of faults or shear-zones. Given that, the effective viscosity of the shallow lithosphere is larger compared to layers underneath, its deformation is no longer controlled by viscous creep. It is, instead, predominantly determined by the movement of lithospheric blocks along pre-existing faults near the surface and by rocks' distributed micro-cracking at lower depths. The scalar visco-plasticity representing one form of weakening mechanisms is the easiest to implement numerically. The basic assumption consists in finding an effective plastic viscosity, η_p , such as that the deviatoric stress is bonded to some sort of yield stress parameter.

Following the von Mises relationships (Hodge and Prager, 1951), this can be expressed as:

$$\boldsymbol{\tau} = \eta_p \dot{\boldsymbol{\epsilon}}, \quad (3.29)$$

Assuming η_p is isotropic and considering τ_{II} bounded by some yield criterion, (i.e. $\tau_{II} = Y$), then we can compute:

$$\eta_p = \frac{Y}{\dot{\epsilon}_{II}}, \quad (3.30)$$

The so-called “Byerlee’s law” outlines crustal rocks’ yield stress dependence from pressure (depth), in accordance with a conspicuous amount of laboratory experiments on various rocks types (Byerlee, 1978). Brittle rocks behaviour is often represented by the “Mohr-Coulomb failure criterion”, which generally an approximation of the Drucker–Prager criterion in geodynamic modelling (Ismail-Zadeh and Tackley, 2010). The Drucker–Prager criterion linearly relate shear and normal stresses and can be written as:

$$\tau_f = \sigma_f \tan(\phi) + c \quad (3.31)$$

with τ_f and σ_f the shear and normal stresses respectively; c a constant cohesion independent pressure term (typically 20-50 MPa) and (ϕ) internal friction angle (generally 30-50°). This equation can be more conveniently expressed in terms of the maximum shear stress, $\tau_{\max}$, and the differential stress, $\bar{\sigma}$, as:

$$\tau_{\max} = \bar{\sigma} \sin \phi + c \cos \phi \quad (3.32)$$

In computational geodynamics, rocks failure dependence from pressure can be approximated by a series of yield criteria, wherein visco-plastic approximations make different assumptions about the functional form of Y (Spiegelman et al., 2016).

When using a “Von Mises” plasticity model, VM, (see equation 3.30) it can be assumed:

$$Y^{VM} = c, \quad (3.33)$$

A further development of VM which limits the second stress invariant by introducing the dependence to a friction angle, and a confining pressure, p , is the “Drucker-Prager” (DP) criterion. This is generally implemented as:

$$Y^{DP} = \sin(\phi)p - c \cos(\phi) \quad (3.34)$$

The pressure-dependence of Y^{DP} reflects in the material having larger strength at higher pressure (Kaus, 2010). Additionally, the total confining pressure term can be decomposed into a lithostatic pressure, $p_{\text{lith}} = \rho_0 g z$, and a dynamic pressure term, p' , due to viscous deformation and residual buoyancy (Spiegelman et al., 2016), as follow:

$$p = p_{\text{lith}} + p' \quad (3.35)$$

When pressure is assumed to be only the lithostatic term, instead of the total pressure, this excludes the normal stress dependence from the dynamic pressure. In this case, the effective yield criterion is called “Depth-Dependent Von Mises”, DDM, (Moresi and Solomatov, 1998; Tackley, 2000; Stein et al., 2004; Spiegelman et al., 2016) and is given by:

$$Y^{DDM} = \sin(\phi)p_{\text{lith}} - c \cos(\phi) \quad (3.36)$$

3.2.2.5 Visco-elasto-plastic rheology

The natural behaviour of rocks is altogether visco-elasto-plastic. This can be computed by decomposing the total deviatoric strain rate $\dot{\epsilon}_{ij}$ into the three components:

$$\dot{\epsilon}_{ij} = \dot{\epsilon}_{ij}^{\text{vs}} + \dot{\epsilon}_{ij}^{\text{pl}} + \dot{\epsilon}_{ij}^{\text{el}} = \frac{1}{2\eta_{\text{vis}}} \tau_{ij} + \dot{\lambda} \frac{\partial Q}{\partial \sigma_{ij}} + \frac{1}{2\mu} \frac{D\tau_{ij}}{Dt} \quad (3.37)$$

$\dot{\epsilon}_{ij}^{\text{vs}}$, $\dot{\epsilon}_{ij}^{\text{pl}}$ and $\dot{\epsilon}_{ij}^{\text{el}}$ are the viscous, plastic and elastic strain rate components respectively. $\dot{\lambda}$ is a scalar plastic multiplier and $Q = \tau_{II}$ is the potential plastic flow parameter. See other works for further details on this topic (e.g. Moresi et al., 2007; Ismail-Zadeh and Tackley, 2010) or §4.9 for the numerical visco-elasto-plastic rheology implementation adopted in this thesis work.

3.2.3 Other equations

The conservation equations together with the rheology laws are needed to describe the fundamental behaviour of a geological body of rock; however, additional sets of relationships are also necessary to further prescribe the fluid intrinsic mode of deformation, such as the equation of phase transformations or the incompressibility/compressibility assumptions.

3.2.3.1 Boussinesq approximation

The Boussinesq approximation commonly used in numerical geodynamics assumes materials to be incompressible (Gerya, 2009; Ismail-Zadeh and Tackley, 2010, e.g.). For the major part of lithosphere and mantle rocks, in absence of phase transformation, incompressibility represent an appropriate approximation, since they have very small thermal expansion and compressibility terms. However, incompressibility could be an over simplification resulting in contradiction with taking into account external sources of heat, for example when modelling the evolution of the early Earth (Ismail-Zadeh and Tackley, 2010). The Boussinesq approximation is one of the equations of states equation (3.40) below, in addition to the equation of state (1) for the conservation of energy and the equation of state, equation (3.38), and the equation of state (2) on the relationship for the isotropic parts of the stress/strain tensor, equation (3.39), as follow from Becker and Kaus (2010):

$$E = c_p \rho T, \quad (3.38)$$

$$\rho = f(T, P), \quad (3.39)$$

$$\Delta \rho = \alpha \rho_0 \Delta T \quad (3.40)$$

with E being a heat energy term of the “caloric” equation, $\rho = \rho_0 \varepsilon_{kk}$, $P = -\sigma_{kk}/3$, α the thermal expansion coefficient, $\Delta \rho$ the density difference from the reference state ρ_0 . When a Boussinesq approximation is implemented, the momentum equation takes density anomalies to be temperature-dependent. An example of a widely used equation to describe density-linear dependence from temperature is $\rho = \rho_0[1 - \alpha(T - T_0)]$, where T_0 is the reference temperature (Ismail-Zadeh and Tackley, 2010).

3.2.3.2 Thermal structure of the oceanic lithosphere

The half-space cooling method provides a reasonable approximation for the modelling of the conductive cooling of the oceanic lithosphere as it moves away from spreading ridges. In this case, the heat conduction equation can be simplified as follow:

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2} \quad (3.41)$$

given the following initial conditions:

$$\begin{cases} T(z, 0) = T_m \\ T(0, t) = 0 \end{cases} \quad (3.42)$$

with T_m being the temperature of the upwelling mantle, the solution of this equation is given by:

$$T(z, t) = T_m \operatorname{erf} \frac{z}{2\sqrt{\kappa t}} \quad (3.43)$$

where the error function expressed as follow:

$$\operatorname{erf} x = \frac{2}{\sqrt{\pi}} \int_0^x \exp(-x^2) dx \quad (3.44)$$

where z is the depth below the surface, κ is the thermal diffusivity (assumed to be constant) and t is the age of the oceanic plate. Assuming isostasy, the increase in oceanic lithosphere buoyancy must be compensated by an increase in thickness of the above sea-water layer. The total weight of a column of lithosphere is thus obtained by integrating its density in equation (3.40), where $\Delta T = T - T_m$ and T is substituted by equation (3.43).

The surface heat flow hence varies inversely with the square root of lithospheric age (i.e., $1/\sqrt{t}$), and sinking of the oceanic lithosphere with $\sqrt{t}$. This is consistent with observations of sea-floor bathymetry up to ages of 70 to 80 Myr (e.g. Schubert et al., 2001; Turcotte and Schubert, 2014). At older ages, the oceanic plates are thinner than what is predicted by this method and other sources of heat may be of major relevance than the half-space cooling method illustrated here, such as small scale convection or mantle plumes, among others (Turcotte and Schubert, 2014).

3.2.3.3 Thermal structure of the continental lithosphere

Radiogenic heat is a relevant heat source for the continental lithosphere and its recycling cycle through the Earth's interior. Radiogenic elements are of fundamental importance in the long term budget of the Earth. Indeed, the continental crust contains more than 100 times the amount of heat-producing elements of the oceanic crust and more than 1000 times the amount in the upper mantle (e.g. Schubert et al., 2001; Turcotte and Schubert, 2014). As most of the continental lithosphere is old (>100 Myr), its thermal structure can be described as being in a steady-state ($\frac{\partial T}{\partial t} = 0$). Assuming constant material properties with depth, a surface temperature, T_s , and a surface heat flux, q_s , the solution of the additional term of the heat conduction equation (H due to radiogenic heat production) becomes:

$$T(z) = T_s \frac{q_s z}{k} - \frac{az^2}{2k} q(z) = q_s - Az \rightarrow q_s + q_m + AD \quad (3.45)$$

where the subscripts s and m refer to surface and mantle, and D indicates the thickness of the layer containing heat producing elements. Beside the initial concentration

of radiogenic elements, A_0 , their concentration, A , decreases linearly with depth of the crustal layer. This relationship is generally expressed as follow:

$$A(z) = A_0 \exp(-z/D) \quad (3.46)$$

In the continental lithosphere A_0 is usually $\approx 2.5 \times 10^{-6} \text{ W/m}^3$ and $D \approx 10 \text{ km}$, which indicates that about the 50% of the continental surface heat flow is produced in the upper crust, and only the remaining 50% derives from underneath. Both A and q_s in continents decrease with lithospheric age; however, this trend is less clear for continents than oceans (Turcotte and Schubert, 2014).

For more information about the thermal and compositional structure of the lithosphere and mantle, I would direct the reader to classical textbooks and famous research articles (e.g. O’Connell and Hager, 1980; Ranalli, 1995; White and Lowrie, 1998; Davies, 1999; Schubert et al., 2001; Fowler, 2004; Bercovici, 2007; Turcotte and Schubert, 2014)

3.3 Boundary conditions

Boundary conditions (BDs) are needed to solve the governing equations, determine the intrinsic behaviour of the modelled phenomenon and perform the computer simulations. In geodynamics, BDs are generally mechanical or thermal, or both depending on the nature of the process studied. Here, a brief summary of only some of these conditions, which could be relevant for this thesis work, is provided following the description of Gerya (2009).

Mechanical boundary conditions can include free slip, no-slip, free surface, fast erosion and infinity-like prescribed velocity, periodic and combined conditions.

A **free slip** condition entails that the orthogonal velocity to the boundary is equal to zero, and the gradients of the other two components are also zero. This implies zero shear strain rate and shear stress along the boundary. The free slip boundary condition requires to the boundary normal to the x -axis:

$$v_x = 0, \quad (3.47)$$

$$\frac{\partial v_y}{\partial x} = \frac{\partial v_z}{\partial x} = 0 \quad (3.48)$$

A **no-slip** condition indicates that all the velocities on the boundary vanish.

$$v_x = v_y = v_z = 0 \quad (3.49)$$

A **free surface** condition requires both the shear and normal stresses at the boundary are equal to zero. This can be explicated as:

$$\sigma_{ij} = 0 \quad (3.50)$$

To implement a free surface boundary condition, one can either include a deformable “free surface” grid or a low viscosity “sticky air” layer above the free surface (see Kaus et al., 2010; Crameri et al., 2012; Rose et al., 2017).

A **prescribed velocity** condition is a moving boundary condition that involves an imposed non-zero velocity at a model boundary. If this condition is applied to two opposite boundaries in order to maintain mass conservation in the model, this allows the boundaries to be displaced over time in response to flow dynamics.

A **periodic** condition is usually established for a pair of parallel lateral boundaries. It prescribes that all the material variables, including mesh variables (e.g., pressure and velocity) are identical at both sides of each boundary. This condition is typically used in mantle convection modelling to mimic a spherical/cylindrical shell in the Cartesian coordinate system.

Combined conditions represent a composite of different types of boundary conditions.

Thermal boundary conditions are constant temperature, zero heat flux, constant heat flux, periodic boundary and combined boundary conditions.

A **constant temperature** condition can be expressed as follow:

$$T = \text{const}(x, y, z, t) \quad (3.51)$$

it usually applied at the top and bottom boundaries of the modelled box.

A **zero heat flux** (insulating boundary, symmetry condition):

$$q_x = -k \frac{\partial T}{\partial x} = \frac{\partial T}{\partial x} = 0 \quad (3.52)$$

This boundary condition is usually applied to the left and right sides of a model. Given that no temperature gradient is applied to this boundary, the heat can not flux through a it.

A **constant heat flux** does not prescribe a temperature limit at the boundary, however, it requires a heat flux across the boundary which can be time and space dependent.

$$q_x = -k \frac{\partial T}{\partial x} = \text{const}(x, y, z, t) \quad (3.53)$$

Periodic boundary and **combined boundary conditions** are thermal analogies of the aforementioned mechanical cases.

3.4 The numerical code: Underworld2

The computational codes Underworld2, Uw2, was used to perform the computational simulations shown in this thesis. The code numerically solves the governing equations defined in §3.2.1 and is based on the finite element method (FEM). Different numerical techniques exist to approximate partial differential equations, PDEs, and for major details on each of these techniques including the FEM and how to obtain a numerical solution, I would direct the reader to the in-detail descriptions of Strang and Fix (1973); Zienkiewicz and Taylor (2000); Peiró and Sherwin (2005).

Underworld2 is a 2-D and 3-D parallel, Eulerian thermo-mechanical finite element code, combined with a marker and cell approach to simulate large-scale geodynamic processes. In Uw2, Lagrangian particles carry material information and are advected using the velocity solution. Particle information is, then, simply mapped to the Gauss points (which are used for the numerical integration) using a nearest neighbour scheme (Mansour et al., 2020). The finite element mesh is based on a hybrid approach that well represents complex fluid behaviours, for example the motion of the Earth’s mantle over geologic time. Underworld2 allows the mesh to be defined as static or deformable, in addition to the tracking of stress deformation history associated with fluid motion, such as in the case of elastic deformation. The details of the numerical method, software implementation and relevant numerical benchmarks are described in a series of papers including Moresi and Solomatov (1995); Moresi et al. (2003); Stegman et al. (2006); Moresi et al. (2007) and Mansour et al. (2020).

In comparison to the previous version of the code, Underworld2 has a python-friendly interface which provides a programmable and flexible interface. This interface allows a limited set of accessible core functionality and an endless number of possibilities to users for implementing individual work-flows independently. In fact, Underworld2 has been designed to work well with Jupyter notebooks, Binder and Docker systems to facilitate learning activities, as well as enhance replicability and reproducibility of research findings across the scientific community (Mansour et al., 2020). Ultimately, the code can perform simulations in high resolution on multiple high-performance supercomputers clusters.

Shape functions are needed to generate a system of equations which equals the number of unknowns (e.g., classic algebraic system). To determine the unknowns the integral of the weighted residual is required to be zero on an element basis. This can be obtained by weighing the residual of the approximate solution by a set of weighting functions, each of which is, in turn, integrate over the element volume and equate to zero. Many methods can be used to achieve this outcome, the difference between which depends on the choice of the weighting functions. In Underworld2 the weak form of the problem is formulated via the standard Galerkin method. For this case, the variables u (velocity) and p (pressure) are expressed with linear shape functions for velocity and constant shape functions for cell-centred pressure, and the incompressibility constraint is considered as an independent equation:

$$\begin{bmatrix} \mathbf{K}^e & \mathbf{G}^e \\ (\mathbf{G}^e)^T & \mathbf{0} \end{bmatrix} \begin{bmatrix} \mathbf{U}^e \\ \mathbf{p}^e \end{bmatrix} = \begin{bmatrix} \mathbf{f}^e \\ \mathbf{0} \end{bmatrix} \quad (3.54)$$

where

$$K_{ab}^e = \int_{\Omega^e} \hat{\mathbf{B}}_a^T \mathbf{D} \hat{\mathbf{B}}_b \Omega^e,$$

$$G_{ab}^e = - \int_{\Omega^e} \hat{\mathbf{B}}^T \mathbf{m} \mathbf{N}_p \Omega^e,$$

where $\mathbf{U}$ and $\mathbf{p}$ are the velocity and pressure vectors respectively, and

$$\mathbf{f}^e = \int_{\Omega^e} \begin{bmatrix} \mathbf{N}^T & f_x \\ \mathbf{N}^T & f_y \end{bmatrix} \Omega^e - \oint_{\Gamma^e} \begin{bmatrix} \mathbf{N}^T & t_x \\ \mathbf{N}^T & t_y \end{bmatrix} \Gamma^e$$

In a standard finite element formulation, $\mathbf{K}$ is the global stiffness matrix for an element and $\mathbf{G}$ is the discrete gradient operator, whereas $\mathbf{K}^e$ and $\mathbf{G}^e$ are the stiffness matrix and gradient operator for an element, $\mathbf{B}$ is a matrix including shape functions derivatives corresponding to a node (a) obtained from the constitutive laws, and $\mathbf{D}$ is a matrix of material properties. In the Underworld2 derivation, the material property matrix is considered to be an attribute of an individual particle together with its current state and history and not a mesh property (Moresi et al., 2007), so $\mathbf{K}_{ab}^e$ is thereby expressed as:

$$K_{ab}^e = \sum_{p=1}^{n_{ep}} \omega_p \hat{\mathbf{B}}_a^T(\mathbf{x}_p) \mathbf{D}_p \hat{\mathbf{B}}_b(\mathbf{x}_p) \quad (3.55)$$

where n_{ep} is the number of particles found in an element. Lagrangian elements carry out material properties and history variables and are advected via a second-order accuracy Runge Kutta method, whilst acting as integration points for the calculations.

Elements are filled with particles using a global space filler layout. This layout is based on a quasi-random particle distribution algorithm (Stegman et al., 2010), which

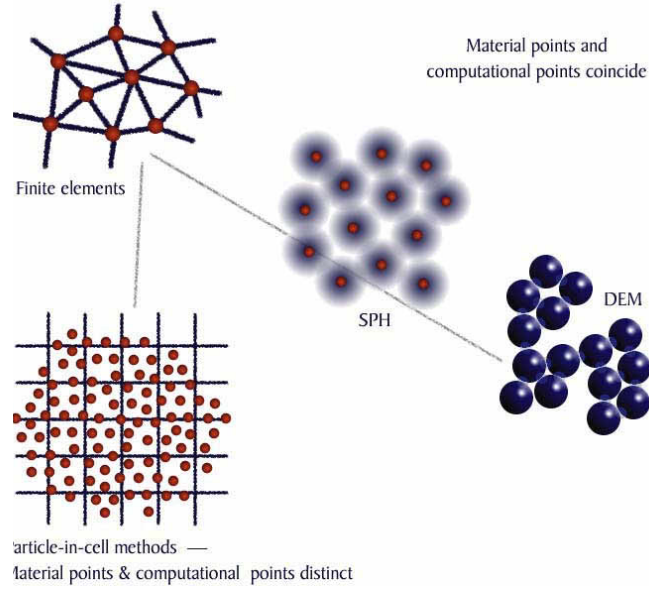


Figure 3.3: Underworld2 approach to discretise a mathematical problem with distinct controlling parameters allowing a natural representation of a system, taken from Moresi et al. (2003). Underworld2 combines a fixed FEM Eulerian mesh with mobile Lagrangian particles. The Lagrangian particles are advected using the velocity solution and their material information is mapped to Gauss points using a nearest neighbour scheme, whereby is then used for the integration in time. While in a fully FEM Lagrangian approach, the computational points are also material points, in the Lagrangian integration point method, computational points are not material points, but a set of material points is also tracked; conversely in an ALE and Eulerian FEM method, there is no tracking of material points.

allows new particles to fill in any eventual void that may result from the stretching of a fluid close to stagnation points in the velocity field, as the model progresses in time. The flow/addition of new particles in the system is controlled by employing a particle population control toll which maintains a balanced smooth density distribution of the particles in the system. Such as that, the addition of particles introduces new integration points into the system but no additional mass.

When solving the incompressible equation of motion with Underworld2, it is possible to use either the penalty method or the Uzawa method. The Uzawa method can be expressed as a systematical iterated conjugate gradient scheme for velocities which give the full solution and solve the pressure equation (Moresi and Solomatov, 1995).

$$\mathbf{G}^T \mathbf{K}^{-1} \mathbf{K} \mathbf{v} + \mathbf{G}^T \mathbf{K}^{-1} \mathbf{G} \mathbf{p} = \mathbf{G}^T \mathbf{K}^{-1} \mathbf{f} \quad (3.56)$$

In the Uzawa method (Moresi et al., 2007), velocity is apparently eliminated and converted in the expression entirely for the pressure $\mathbf{G}^T \mathbf{p}$:

$$\mathbf{G}^T \mathbf{K}^{-1} \mathbf{G} \mathbf{p} = \mathbf{G}^T \mathbf{K}^{-1} \mathbf{f} \quad (3.57)$$

The penalty method, instead, introduces a Lagrange multiplier (an incompressible constraint) directly into the variational form of the equation of motion. Both methods, Uzawa and penalty, are of direct applicability to Stokes' problems with a strong variation in layers viscosity (see Moresi and Solomatov, 1995; Stegman et al., 2010).

3.4.2 Non-dimensionalisation and Stokes analytical solution

The dimensional analysis is computed accordingly with the Newton second law motion see equation (3.3) and was implemented in all the research chapters (4-6).

Parameter	Characteristic			Reference			Scaling Factor	
	symbol	value	unit	symbol	value	unit	symbol	value
Length	l_D	1e5	[m]	l_{DL}	1.0	-	l_{SF}	l_{DL}/l_D
Viscosity	η_D	1e21	[Pa·s]	η_{DL}	1.0	-	η_{SF}	η_{DL}/η_D
Stress	σ_D	1e10	[Pa]	σ_{DL}	1.0	-	σ_{SF}	σ_{DL}/σ_D
Temperature	T_D	1e3	[K]	T_{DL}	1.0	-	T_{SF}	T_{DL}/T_D

Table 3.1: Characteristic, reference and scaling factor values for the chosen parameters.

Parameter	Basic Scaling Coefficient	
	symbol	value
Time	t_{SF}	η_{SF}/σ_{SF}
Force	F_{SF}	$\sigma_{SF}/(l_{SF} * l_{SF})$
Acceleration	g_{SF}	$l_{SF}/t_{SF}/t_{SF}$
Mass	m_{SF}	F_{SF}/g_{SF}

Table 3.2: Basic scaling coefficients calculated accordingly with the Newton's second law of motion.

3.4.2.1 Steady flow of a viscous fluid past a sphere

An important laminar flow problem with an analytic solution is that of a viscous sphere η_s moving with a small constant velocity through an incompressible fluid medium with viscosity η_m . The sinking Stokes velocity is given by:

$$V_{Stokes} = C \frac{\Delta \rho g r^2}{\eta_m} \quad (3.58)$$

with r being the radius of the sinking sphere, a viscosity ratio dependency can be expressed as follow:

$$C = \frac{2 + 2\eta'}{6 + 9\eta'} \quad (3.59)$$

where

$$\eta' = \frac{\eta_s}{\eta_m} \quad (3.60)$$

For a rigid sphere ($\eta' \rightarrow \infty$) and $C = \frac{2}{9}$, for a weak bubble ($\eta' \rightarrow 0$) and $C = \frac{1}{3}$ indicating that v_{Stokes} is mostly sensitive to η_m and not to η' . Different orientation of an elongated sinker also does not affect C by more than a factor of ~ 2 , (Batchelor, 2000). Then, the stokes velocity sets a time scale for any buoyancy dominated subduction problem ($V \propto \Delta\rho r^2/\eta_m$ and $\tau \propto \Delta\rho r$), (see Becker and Faccenna, 2009).

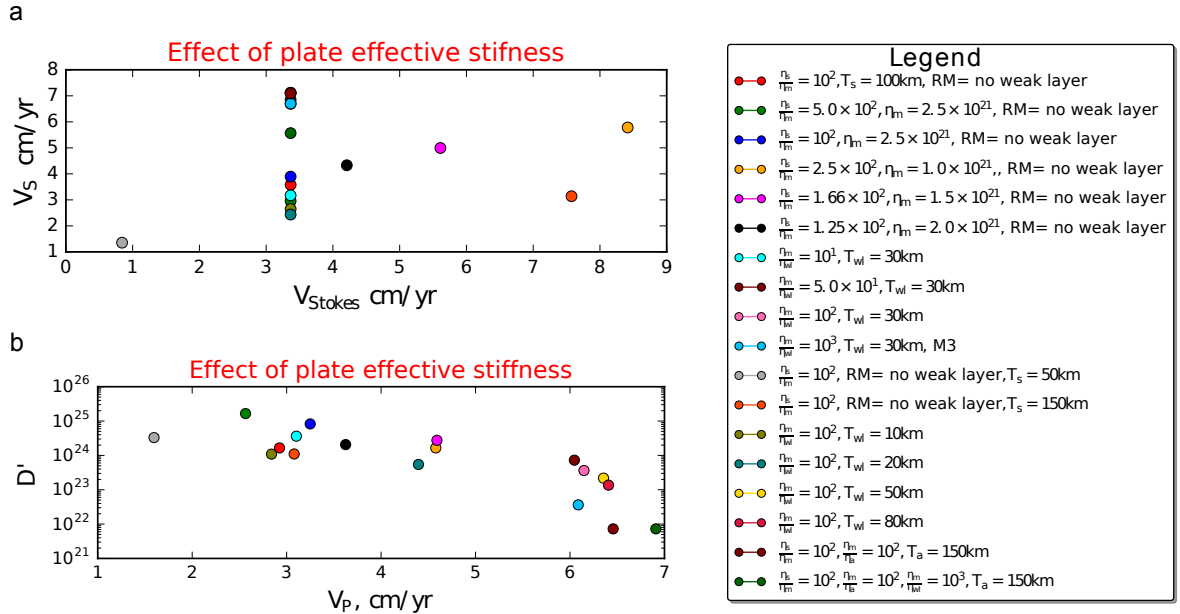


Figure 3.4: In a, relationship between numerical subduction velocity and analytical Stokes velocity (equation 3.58); in b, relationship between effective relative stiffness of the lithosphere, D' , (equation 4.2) and plate numerical velocity for the numerical simulations performed in chapter 4. While Stokes analytical solution well explains systems mainly driven by density and lithosphere-mantle bulk viscosity variation, such as the case of our reference model (in §4.6.1), other variables (e.g., D') could become relevant to describe the effect of the properties of the effective relative strength of the lithosphere relative to the underlying mantle on plate velocity, as shown in b (see also chapters 4-7). Further details about the numerical experiments and parameters varied are available in chapter 4.

4. **Chapter 4: The impact of a very weak and thin upper asthenosphere on subduction motions**

The impact of a very weak and thin upper asthenosphere on subduction motions ¹

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4.1 Highlights

- Numerical models show that a very weak and thin upper asthenosphere layer can alone reduce trench retreat and enhance plate motion
- The impact of this effect depends on the relative contrast between the effective stiffness of the lithosphere and the underlying mantle
- This regional mechanism explains the high convergence and low trench migrations rates observed globally at natural subduction zones

4.2 Abstract

Recent geophysical observations report the presence of a very weak and thin upper asthenosphere underneath subducting oceanic plates at convergent margins. Along these margins, trench migrations are significantly slower than plate convergence rates. We use numerical models to assess the role of a weak upper asthenospheric layer on plate and trench motions. We show that the presence of this layer alone can enhance an advancing trend for the motion of the plate and hamper trench retreat. This mechanism provides a novel and alternative explanation for the slow rates of trench migration and fast-moving plates observed globally at natural subduction zones.

4.3 Plain Language Summary

Plate tectonics relies on the concept of a rigid surface layer (the lithosphere) fragmented in a series of major and minor tectonic plates moving over a weaker and buoyant layer of asthenospheric material. The motion and deformation of the lithosphere are primarily

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driven by the sinking of the colder and denser oceanic lithosphere through time into the Earth’s interior at subduction zones. A range of geophysical regional studies has recently brought considerable attention to the detailed structure of the asthenosphere at these zones. These studies report the presence of a thin and even weaker asthenospheric layer at the base of the subducting lithosphere. We investigate the role of this layer on subduction dynamics using geodynamics numerical models. Our models apply knowledge of physics, chemistry, mathematics and geology, and study the dynamics of the Earth’s deep interior and their feedback on surface deformation. Our numerical results demonstrate that a very weak and thin upper asthenosphere affects the dynamics of sinking plates. It acts as a slippery base for the motion of the plate significantly increasing its speed and deformation. Our research provides a novel explanation to interpret some of the subduction phenomena at the regional scale observed on Earth, which have proven numerical difficulties until now.

4.4 Introduction

The boundary between the Earth’s tectonic plates and the underlying mantle, the lithosphere-asthenosphere boundary (LAB), is a fundamental element of plate tectonics as it critically controls the mechanical coupling between the motion of the tectonic plates and the flow in the underlying mantle (Anderson, 1995). The motions and deformation of the solid plates are primarily driven by the downward pull of slabs (Forsyth et al., 1975; Harper, 1975; Chapple and Tullis, 1977; Conrad and Lithgow-Bertelloni, 2002) and by the coupling with the viscous mantle flow surrounding them (Lithgow-Bertelloni and Richards, 1995, 1998; Conrad and Hager, 1999; Becker and O’Connell, 2001). Forces associated with the sinking of slabs into the mantle propagate to the surface controlling plate motions towards the trench zone and mantle flow around subducting slabs. Therefore, a very weak and thin upper asthenosphere (WL) may contribute to the drag exerted at the base of subducting plates having an impact on the force balance around the oceanic lithosphere. This could play an important role in decoupling the motion of tectonic plates from the flow in the underlying mantle, promoting or opposing subduction motions with respect to the mantle underneath.

The subduction of an oceanic plate into the mantle occurs concurrently with the motion of trenches at the surface. These are mobile and can either retreat, advance, or have a mixed type of motion in time with respect to a specific reference frame. We refer to the percentage ratio between the horizontal plate (V_P) and trench migration (V_T) velocities as the subduction partitioning (V_T/V_P). The subduction partitioning, as well as the subduction dip angle and the morphology assumed by the subducting slab while interacting with the 660 km transition zone are considered diagnostic elements to identify the subduction style (e.g. Christensen, 1996; Zhong and Gurnis, 1997; Bellahsen et al., 2005; Capitanio et al., 2007; Faccenna et al., 2007; Di Giuseppe et al., 2008; Funicello

et al., 2008; Capitanio et al., 2009; Stegman et al., 2010; Goes et al., 2011; G  rault et al., 2012). Most of the Earth’s slab and trench motions occur in conjunction with roll-back and retreat ($-V_T$), however advancing plate and trench motion ($+V_T$) towards the upper plate can also be found in some trench zones, e.g. at some of the edges of the Pacific plate (Schellart et al., 2008). Furthermore, the average absolute trench velocity observed globally is one third or less of the total convergence at the present, consistently across the majority of plate motions reference frames (e.g. Elsasser, 1971; Jarrard, 1986; Schellart et al., 2008; Becker and Faccenna, 2009; Goes et al., 2011; Holt and Becker, 2016). Numerical and analogous subduction models have previously helped explain the fundamental force balance of plate motions, yet the observed slow retreat with fast advance does not emerge as a general feature (Holt and Becker, 2016; Coltice et al., 2017). Geoid observations (Moresi and Gurnis, 1996; Mitrovica and Forte, 1997; Zhong and Davies, 1999), and numerical and geophysical studies (e.g., Schellart, 2004b; Bellahsen et al., 2005; Di Giuseppe et al., 2008; Funiciello et al., 2008; Schellart et al., 2008; Wu et al., 2008; Capitanio et al., 2009) constrain the viscosity contrast between the subducting oceanic plates and the mantle to be within a range of 100-500. These show significant slab roll back and trench retreat (i.e. $V_T/V_P < -0.3$). In most of these models, trench advance occurs for stiff, slowly migrating plates such as those with a viscosity contrast between the subducting lithosphere and the mantle greater than 1000. However, this is not easily reconciled with what is generally observed in nature and might apply to regional cases only (e.g. Garfunkel et al., 1986; Jarrard, 1986; Bellahsen et al., 2005; Lallemand et al., 2005; Sdrolias and M  ller, 2006; Faccenna et al., 2007; Di Giuseppe et al., 2008; Funiciello et al., 2008; Schellart et al., 2008; Stegman et al., 2010; Goes et al., 2017). This still presents the difficult task of explaining the low apparent strength of tectonic plates relative to the underlying mantle and the efficient slab pull propagation needed in subduction models to match plate velocities. Some solutions to explain such a slow retreating motion have been proposed, such as the role of slab rheology (e.g. Billen and Hirth, 2007; Capitanio et al., 2007, 2009) or a complex mantle rheology (e.g. Holt and Becker, 2016), the effect of slab and trench widths (e.g. Schellart et al., 2007, 2010), the presence of an overriding plate (e.g. Rodr  guez-Gonz  lez et al., 2012; Holt et al., 2015b), multiple plates (e.g. Yamato et al., 2009) or slabs (e.g. Holt et al., 2017; Faccenna et al., 2018b), the effect of the interaction between the slab and the upper/lower mantle discontinuity (e.g. Goes et al., 2008; Garel et al., 2014; Yang et al., 2017; Billen and Arredondo, 2018) and the effect of ridge push and a low viscosity asthenosphere (e.g. Capitanio et al., 2007; Stegman et al., 2010).

Some of these studies (e.g. Capitanio et al., 2007; Holt and Becker, 2016; Billen and Arredondo, 2018) have also provided insights into how an asthenosphere of diverse origin could diminish trench retreat rates. However, the globally observed trend in trench and plate motions and their partitioning at subduction zones is yet to be investigated,

especially in light of the recent geophysical observations of a very weak and thin upper asthenosphere (Kawakatsu et al., 2009; Schmerr, 2012; Naif et al., 2013; Stern et al., 2015; Hawley et al., 2016). Common to all these observations is the evidence of a seismic velocity drop (5 to 10%) across a sharp boundary, no more than 30 km thick, at base of the subducting lithosphere at different locations in the Pacific plate. The presence of this layer, which may support the theory of a plume-fed asthenosphere (e.g., Morgan, 1971, 1972; Phipps Morgan et al., 1995), is attributed to a significant number of volatiles and/or hydrated mantle phases. These are thought to have lowered density and viscosity, increasing the partial melt fraction, and resulting in partial melt "ponded" in lenses or channels at the base of the oceanic lithosphere (e.g., Schmerr, 2012; Sakamaki et al., 2013; Hawley et al., 2016). Furthermore, at these localities, we generally observe $|V_T/V_P| < \frac{1}{3}$, which applies in the majority the reference frames (Schellart et al., 2008; Coltice et al., 2017; Goes et al., 2017). A thin layer with a high fraction of partial melt is expected to have significantly reduced mechanical strength that could potentially decouple plate motions from mantle tractions and strongly influence the force balance between slabs, plate and the mantle.

Previous global numerical models (e.g., Lithgow-Bertelloni and Richards, 1995; Bercovici, 1996; Zhong et al., 1998; Tackley, 2000; Capitanio et al., 2007, 2009; Stegman et al., 2010; Höink et al., 2011; G  rault et al., 2012; H  ink et al., 2012; Becker, 2017; Goes et al., 2017) analyzed the role of a low viscosity asthenosphere (LVZ) but only a few have accounted for an LVZ of a limited thickness within the range of 100-200 km (Becker, 2017; Richards and Lenardic, 2018; Semple and Lenardic, 2018). Generally, an asthenosphere seems to better characterize plate-like behaviour improving our ability to match with observed surface velocities (Tackley, 2000; G  rault et al., 2012). Interestingly, a thinner LVZ with a stronger viscosity contrast shows to further promote plate motions leading to a stress magnification at the base of the lithosphere and helping overcome the viscous resistance at the base of the lithosphere (Richards and Lenardic, 2018). However, this has also shown to have a smaller effect on the global net rotation and observed global seismic anisotropy, as well as on global plate and subduction velocities, where large-scale plate dynamics seem most likely governed by broad continent-ocean asthenospheric viscosity and buoyancy contrasts (e.g. Capitanio et al., 2007; Becker, 2017; Goes et al., 2017).

On the other hand, if an LVZ seems to have a negligible effect on the large-scale subduction dynamics or represents a global feature, a very thin and weak top asthenosphere could still play a fundamental role for deciphering some of the observations of subduction partitioning at the regional scale, and provide relevant insights on global subduction zones on Earth. Such as those dominated by fast-moving plates and slow migrating trenches, which have proven numerically difficult until now. Yet, the role of a very thin ($h_{WL} \sim 10$ -100 km) and weak ($\eta_M/\eta_{WL} \sim 10$ -1000) WL on the subduction partitioning (V_T/V_P) has remained unexplored to date. Here we aim to address the impact of a WL on trench and

plate velocities and their partitioning at subduction zones. Additionally, our work builds upon on the recent geophysical inferences of a very weak and thin upper asthenosphere at convergent margins in the Pacific plate. We use a series of numerical experiments carried out in 2D to investigate the impact of a WL on the velocity and morphology of subducting slabs, adopting the geological and geophysical constraints described above. We then discuss the inferences these models yield on subduction dynamics.

4.5 Method

To investigate the role of a WL on subduction dynamics, we use the numerical code Underworld, described in detail by Moresi et al. (2007) and Stegman et al. (2006). The code follows a continuum mechanics approximation, which is widely used to describe geological and geophysical processes, and solve the conservation equations of mass, momentum and energy. We assume incompressibility and neglect temperature diffusion.

The initial model setup follows the approach of Enns et al. (2005); Schmeling et al. (2008); Yamato et al. (2009) and is adapted to solve the problem under investigation. We model subduction dynamics and mantle flow in a Cartesian box that extends 4000 km in the horizontal (X) direction and 660 km in depth (Y). Our models are made of 4 layers. There is an oceanic crust, an oceanic plate, a weak upper asthenospheric layer and an underlying mantle, as shown in figure 4.1b. Although similar, this setup differs from previous studies (e.g. Stegman et al., 2006; Schellart et al., 2007) because the oceanic plate is made of one single layer and the WL is characterized by a lower viscosity value than in the underlying asthenosphere or deeper mantle. The initial condition for the subducting plate includes an initial trench position 1000 km away from the wall and a slab tip already penetrating the mantle, 257 km deep with a dip angle of 34° to the horizontal. The modelled WL is located at the base of the oceanic plate and extends underneath the whole length of the oceanic plate, and has a thickness of 30 km. We use a model domain of 256×128 elements with a uniform grid spacing in each coordinate direction. This ensures sufficient numerical resolution to promote and properly-resolve the development of a weak hinge zone. Our 2-D experiments use a pseudo-plastic (PP) constitutive law with visco-plastic rheology near the surface and a Newtonian formulation deeper into the lithosphere and mantle. The visco-plastic rheology of the crust promotes the development of a weak hinge zone, which simulates the free surface and enables the oceanic lithosphere to detach from the top surface of the model and subduct freely into the mantle (Schmeling et al., 2008; Crameri et al., 2012). Plasticity is implemented through a depth-dependent yield criterion (Moresi and Solomatov, 1998). We use a cohesion of a 20 MPa, a friction angle of 30 degrees and a density contrast between the oceanic lithosphere and the underlying mantle of 80 kg/m^3 . The visco-plastic crustal layer controls the kinematics (Enns et al., 2005; Royden and Husson, 2006); however, the description of

this phenomena is outside the scope of this study. The velocity boundary conditions are free-slip everywhere to minimize the influence of the box sidewalls (a free-slip condition means no tangential stress and zero normal velocity). We do not apply any kinematic boundary and subduction is thus dynamically initiated. The reliability of the model has been tested against previous work as shown in section 4.6. The assigned values result in subduction velocities of 10-2 cm/yr in line with the average values predicted on Earth (figure 4.1c, 4.2 and 4.3).

We aim to identify end-member cases that support the description of the mechanical effect of a WL at the LAB on a subduction zone. For this goal, we use a simplified rheological and geometrical model. This very setup is necessary to constrain only the impact of a WL and to prove that subduction motions can be the result of a simple force balance around subducting plates, while no external forces are needed. In our study, we vary the properties of the WL at the base of the oceanic lithosphere, i.e. its viscosity, density and thickness. Additionally, we also explore the effects of different viscosity and thickness contrast between the plate and the mantle, and the effect of the crustal friction angle. A small number of more complicated models explores the extent to which these simplifications may influence our conclusions, such as the use of a more realistic visco-elastic-plastic temperature (T)-dependent rheology (VEP) with and without a WL. Finally, we omit the study of the mechanical effects given by variation in buoyancy of plates, as these are well documented in the literature (e.g. Christensen, 1996; Bellahsen et al., 2005; Capitanio et al., 2007; Faccenna et al., 2007; Di Giuseppe et al., 2008; Funiciello et al., 2008; Capitanio et al., 2009; Stegman et al., 2010; Goes et al., 2011; G  rault et al., 2012).

The type of flow in the asthenosphere can promote or resist plate motions with respect to the underlying mantle. This flow can be driven by the plates themselves (Couette flow), pressure driven by the flow within the mantle (Poiseuille flow) or a combination of the two (combined Couette-Poiseuille flow). Several authors (e.g. Ribe, 1992; Conrad and Hager, 1999; Ribe, 2001, 2010; Stegman et al., 2010; Buffett and Becker, 2012; Gerardi and Ribe, 2018; Semple and Lenardic, 2018; Gerardi et al., 2019) show that the force balance of the oceanic lithosphere is controlled by the balance of both bending-stretching momentum and boundary and internal stresses (flow in the boundary layer).

Among these authors (e.g. Ribe, 1992, 2001; Gerardi and Ribe, 2018; Richards and Lenardic, 2018; Gerardi et al., 2019) show that the motion of a mantle fluid in an interface layer, can be approximated with the motion of a Stokes fluid in the lubrication limit. This procedure is valid when the thickness of the layer is much thinner than the horizontal length, if $l_{WL}/h_{WL} \ll 1$. An approximation of the tangential forces representing the motion of a Stokes's fluid in a thin layer in the lubrication limit led to the definition of the dimensionless stiffness of a surface shell (Ribe, 1992, 2001) and an interface layer (Gerardi and Ribe, 2018; Gerardi et al., 2019). In our study, we are interested in describing

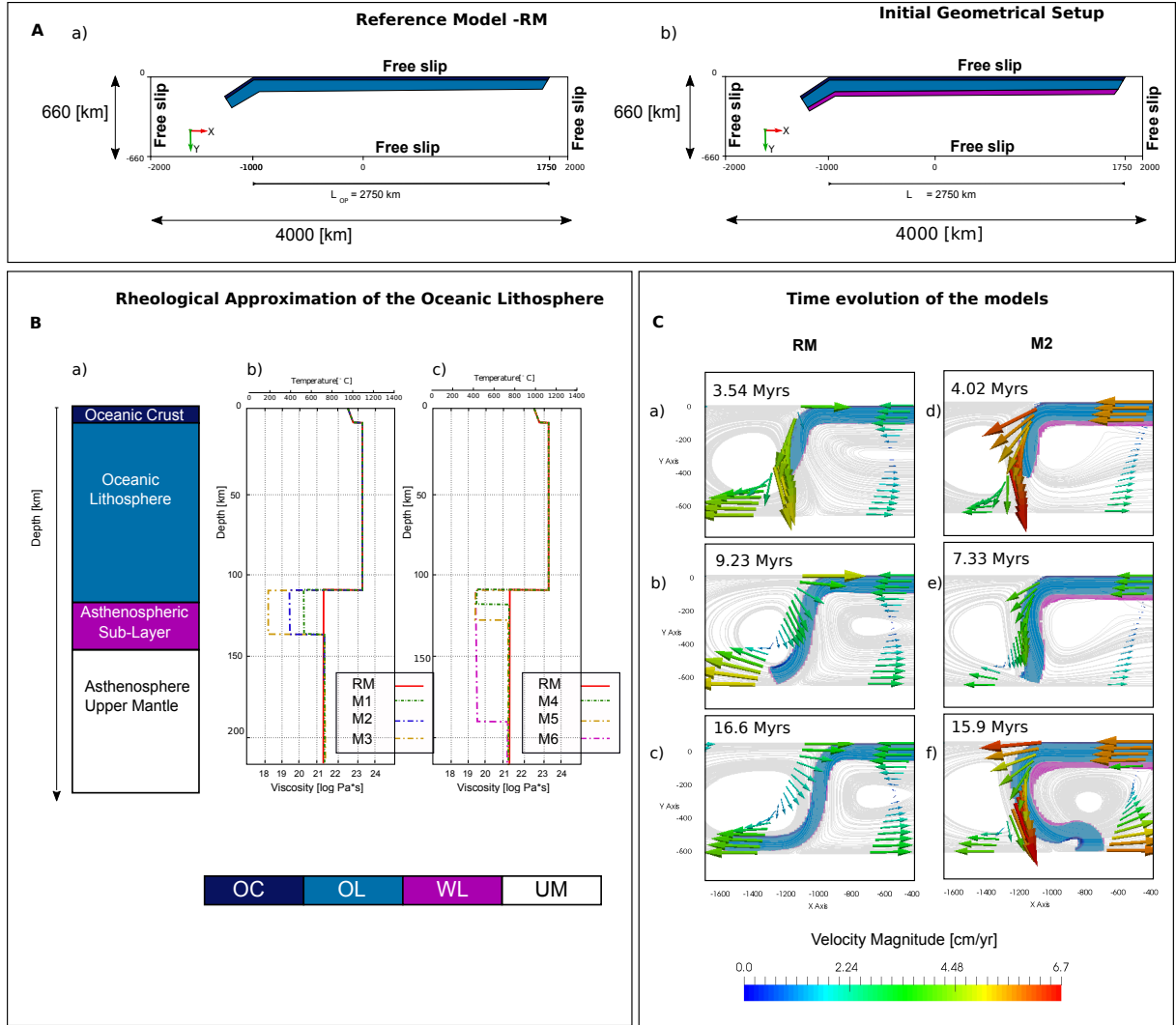


Figure 4.1: In A, a schematic representation of the geometrical setup for the reference model (RM) and the models with a weak layer (WL) in a and b, respectively. In B, the material field in a, the initial rheological setup for the pseudo-plastic (PP) with different viscosity contrast mantle and WL in b and WL thickness in c. In C, the subduction time evolution in a comparison between the RM and a model with a WL (M2, $\eta_M/\eta_{WL}=100$).

the deformation in the effective lithosphere relative to the underlying mantle, which may or may not contain a WL on the partitioning of trench and plate motions. Based on our numerical experiments, we represent a first order approximation of the deformation occurring in the system (with and without a WL) using a series of parameters related to the dimensionless stiffness of the effective lithosphere.

We empirically find that the effective relative strength, S' , and thickness, h' , of the lithosphere relative to the underlying mantle (figure 4.3a and b, and figure 7SI in the supporting information) can be taken into account as:

$$S' = \frac{\eta_P \eta_{WL}}{\eta_M^2}, \quad h' = \frac{h_P}{h_M} \left(1 + \frac{h_{WL}}{h_M} \right) \quad (4.1)$$

where η_P , η_{WL} and η_M indicate the viscosity of the plate, weak layer and mantle, and with h_P , h_{WL} and h_M being thicknesses of the plate, the WL and the mantle respectively. Note that when a WL is not included $\eta_{WL} = \eta_M$ and $h_{WL} = 0$, allowing us to characterise all the experiments with and without a WL.

We also find that the combined effect of the relative contrast in strength and thickness between the effective lithosphere and the underlying mantle can be expressed by the effective relative stiffness of the lithosphere, D' , as:

$$D' = \log(S')h' \quad (4.2)$$

D' represents an approximation of the tangential forces acting on a subducting lithosphere (with and without a WL) and help to explain the decoupling effect of a WL at the base of the lithosphere. Our study represents a balance of forces problem where the subduction dynamics with a WL are governed by the balance of bending-stretching moment and internal and boundary stresses (flow in the lithosphere and asthenosphere), which controls dip and curvature of slabs, figure 4.3c, providing new insights on the partitioning of subduction surface motions.

Further in the text, these parameters are used to describe the deformation occurring in the system. Finally, in our simulations we refer to V_{OP} as the magnitude of the horizontal plate velocity and V_T as the trench migration velocity. The results obtained in our numerical simulations were systematically analysed, particularly through the analysis of the morphological and velocity fields through time evolution of the models. More information about the methodology adopted and the parameters used and varied in this study is available in the supporting information.

4.6 Results

Here, we firstly present the evolution and general aspects of the reference model (RM), which satisfy a range of geophysical and geological observations, then summarize the

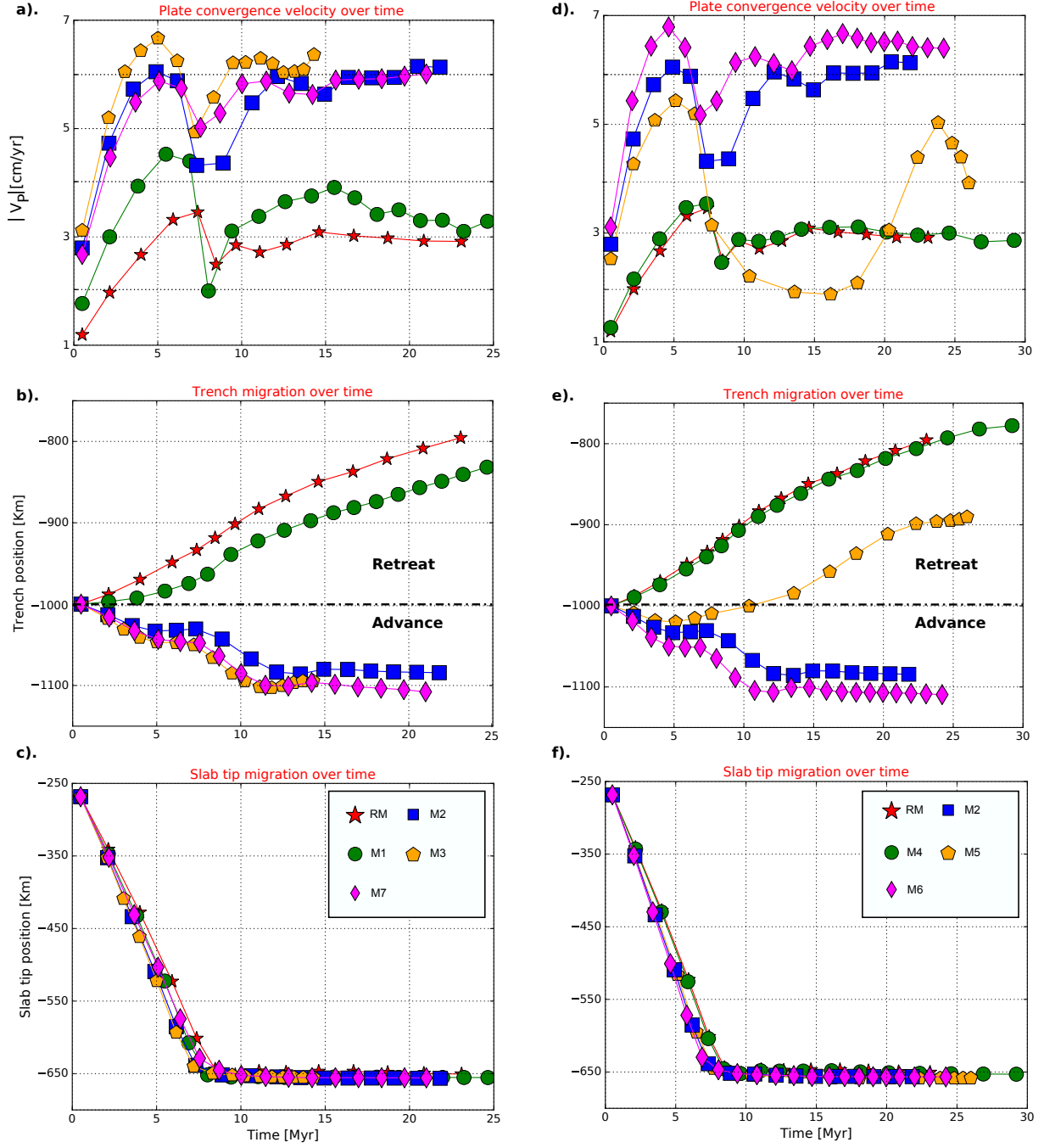


Figure 4.2: Evolution of plate velocities (V_P), trench and slab tip migrations through time, for a progressive increase in η_M/η_{WL} and $\Delta\rho_{(M-WL)}$, and in h_{WL} , as seen in a, b, c and d, e, f, respectively.

results obtained for all the subduction experiments in comparison to the RM case. Also, all the PP models presented here share three common subduction temporal evolution stages. These evolution stages are: (i) an initial period where subduction is dynamically initiated and the slab progressively accelerates while is sinking into the mantle; (ii) a transitional stage where the slab decelerates while interacting with the bottom of the model; and (iii) a final steady-state configuration at the mid-mantle boundary. These are shown in figure 4.1C and 4.2, and are in agreement with previous studies (e.g. Bellahsen et al., 2005; Yamato et al., 2009; Stegman et al., 2010).

4.6.1 Reference model evolution

The reference model has the same characteristic of the model setup presented above, but it does not contain a WL (figure 4.1Aa). The temporal evolution of this model is shown in figure 4.1Ca, b, and c. As the model progresses in time, we distinguish three stages of evolution given by the sinking of the oceanic plate into the mantle. In the first stage, the initiation of subduction follows the initial buoyancy configuration, with a slab tip immersed in the mantle. During this stage, the slab reaches a dipping angle of 60-70° and the magnitude of the plate velocity increases progressively. This stage is associated with trench retreat at the surface, while at depth, the progressive sinking drives a displacement of the mantle from beneath the slab (figure 4.1Ca). In the second phase, the slab tip interacts with the bottom of the model box. This phase is reached in 9.23 million years and is characterized by a general velocity drop (figure 4.1Cb). The final stage is defined by the sliding of the slab at the bottom of the model and the system reaches an approximate steady-state configuration (figure 4.1Cc). During this stage, the plate velocity increases toward an intermediate value between the two previous time phases and the slab continues to roll back while the trench retreats (figure 4.2a, b). Since the reliability and the evolution of this model are well known and in agreement with the literature (e.g. Bellahsen et al., 2005; Funicello et al., 2008), it was chosen as a reference for comparison with the models having a WL.

4.6.2 An advancing trend for plate and trench motions.

The primary effect of introducing a weak upper asthenospheric layer is to increase the subduction speed, particularly increasing plate convergence velocity and reducing trench retreat motion. In some models, it also increases the subduction dip angle and influences the shape assumed by the subducting slab, while interacting with the 660 km discontinuity (figure 4.1C and 4.3). A comparison between the RM and a WL model having a viscosity contrast between the mantle and the WL of 2 order of magnitude ($\eta_M/\eta_{WL}=10^2$) shows these effects in figure 4.1Ca, b, c and 4.1Cd, e, f, respectively.

The impact of these effects depends mainly on the viscosity between the mantle and the WL and on the thickness of the WL, and less on the density of the WL. In our PP models, we find that for moderate viscosity contrasts mantle-WL ($10^1 < \eta_M/\eta_{WL} < 10^2$) and a layer thickness between 10-20 km, V_P increases and trench retreat rate decreases, while the overall shape of the slab is not substantially affected. However, for larger viscosity contrasts ($\eta_M/\eta_{WL} \geq 10^2$) and thickness of the weak layer (>20 km), a significant change in velocity and trench migration mechanism develops and, the morphology of the subducting slab may also change (figure 4.2). For thinner ($h_{WL} < 10$ km), more buoyant ($\Delta\rho_{(M-WL)} \approx 20$) and stronger layers ($\eta_M/\eta_{WL} \leq 10$), the overall effect is reduced. However, thinner layers are sensitive to numerical resolution and, it is plausible that thinner layers may also play a relevant role on the V_T/V_P . It is also important to note that a WL has not a significant impact on the time necessary for the slab tip to reach the bottom of the model (figure 4.2c and f), which shows that a WL has a smaller influence on the vertical slab motion than it has on the horizontal plate motion.

All models with a WL show a significantly steeper velocity transition at the base of the lithosphere compared with models with no WL (figure 4.1C). This suggests that a weak and thin upper asthenosphere has the potential to effectively decouple the motion of the lithosphere from the flow into the underlying mantle. This process results in a more efficient asthenosphere drag, which decreases the resistance to shearing in the underlying mantle, lubricates the base of a plate and in turn contributes to increasing the plate speed (figure 4.1C and figure 4.2a,b,d and e). The WL acts as a slippery base for the motion of the oceanic plate while sinking into the mantle. Since a WL is lighter than the above lithosphere (Kawakatsu et al., 2009; Hawley et al., 2016), also the rising of lighter material contributes to the gravitational sliding of the plate (figure 4.1C). Furthermore, the deep entrapment of a WL on one side of the subducting slab also reduces the viscous drag and thus, facilitates the sinking of the slab into the mantle. Such vanishing drag on one of the slab's sides justifies the increase in sinking velocities up to a factor of ≈ 2 we observe in our models, (figure 4.1C and 4.2a and d).

In our models, we also observe a correlation between the value of the subduction partitioning and the effective relative strength, S' , and the inverse of the thickness, $1/h'$, of the lithosphere, as shown in figure 4.3a and b. Three important features are worth noting. First, this trend can be described by an inverse-tangent function for both S' and $1/h'$. Second, for low and high values of S' and $1/h'$, the subduction partitioning doesn't change significantly and the function trend is asymptotic. Hence a WL can increase plate velocities value up to a maximum of a factor of ≈ 2 and oppose trench motions. This shows that the effect of a WL on subduction motions is limited by the sinking dynamics. Third, this asymptotic trend is further constrained by the fact that areas having $S' > 5 \times 10^2$ and $S' < 10^{-2}$ are excluded from the graph (figure 4.3a). Such values of S' would correspond to plate-mantle and mantle-WL viscosity contrasts above 500 and 1000, respectively. These

are not found in nature and therefore not considered. For intermediate S' and $1/h'$ values, the subduction partitioning follows a co-tangential trend.

In this study, we perform simulations with varied parameters over a range that is Earth-like (figure 4.1b and supporting table T1). Our full set of experiments include cases where the initial viscosity of the plate and the mantle, the plate thickness and the crustal friction angle are varied to reproduce previous work (e.g., Bellahsen et al., 2005; Di Giuseppe et al., 2008; Funiciello et al., 2008), and expand upon these through the incorporation of a WL. On the basis of all our experiments we propose a regime diagram, which defines three distinctive subduction styles as a function of D' , subduction motions ($V_T/V_P\%$) and slab morphologies and dips, figure 4.3c. These subduction regimes are characterized by (i) a very steep slabs buckled at the mid-mantle discontinuity with advancing to quasi-stationary trenches; (ii) a relatively steep slab (but $< 90^\circ$) with quasi-stationary to moderately-retreating trenches, which continue flat at the 660 km discontinuity; and (iii) a shallower slabs with usually fast-retreating trenches, which buoyantly tend to stagnate at the mid-mantle boundary (figure 4.3c). D' allows us to capture and constrain these three subduction styles, which characterise the deformation in the system with and without a WL, as well as most of the variety of shapes and subduction partitioning mechanisms we observe in nature. These observations support a mechanism whereby the subduction style and motions are controlled by the amount of asthenospheric drag exerted at the base of the plate.

Finally, in most of our experiments we observe a correlation between V_P and V_T , which shows $V_T \pm 1/3 V_P$ (figure 4.3d). To test the validity of our models with relation to subduction zones on Earth, we use the estimate on V_P and V_T from Goes et al. (2008) for subduction zones in the Pacific plate and plot this versus our numerical experiments (figure 4.3d). Goes et al. (2008) uses the database of Sdrolias and Müller (2006), which is built on the Indo-Atlantic hot spots reference frame from O'Neill et al. (2005). The values of V_P and V_T emerging from our models mostly positively align with observed natural plate and trench migration rates. There are, however, zones that show poor correlation with our models, such as, the Tonga and Marianas trenches, which show $V_T > \pm 1/3 V_P$. At these localities, it is plausible that other mechanisms may be of major regional importance or coexist with the presence of a WL, and indeed further contribute to the complexity we observe in natural systems, e.g. double subduction systems, arcuate arcs and so on. Nonetheless, at the global scale, it appears that a regional WL mechanism may play a key role in dictating the partitioning of plate and trench motions and offers a better match to the average $V_T/V_P\%$ values we presently observe on Earth.

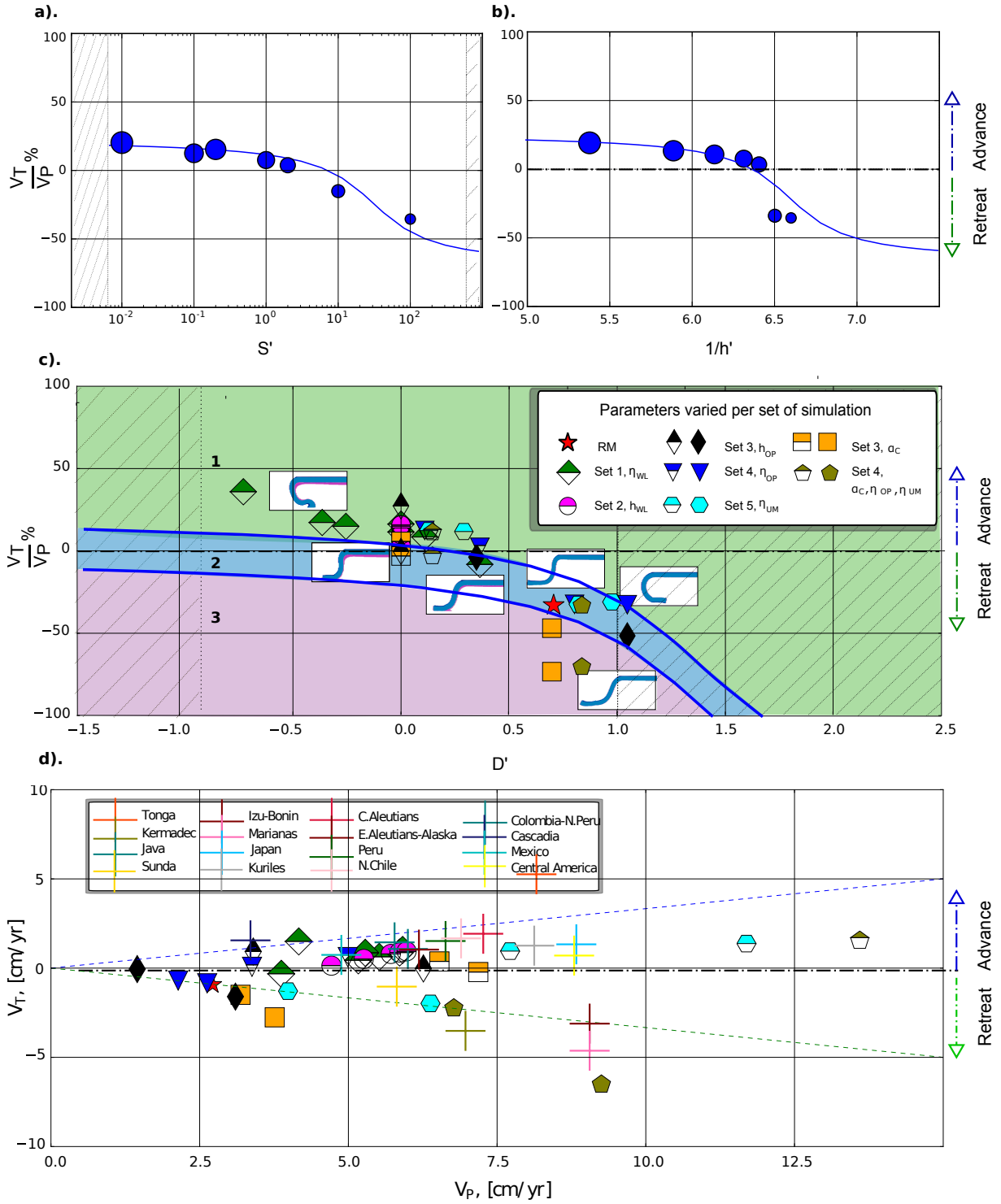


Figure 4.3: In a and b the subduction partitioning (V_T/V_P %) as a function of the effective relative strength (S') and thickness (h') of the lithosphere for a progressively weaker or thinner WL, respectively. The values for each simulation are plotted using a scatter plot function of S' and h' in a and b, respectively. In c, variations of V_T/V_P % with respect to the effective relative stiffness of the lithosphere (D') for all the simulations performed in this study. The different coloured fields define the three areas having different subduction styles. In d, the relationship between V_P and trench velocity V_T , as found in our experiments and in natural cases (Goes et al., 2011). The majority of the points lies between the dotted lines ($V_T = m V_P$) having m equal to $\pm 1/3$.

4.7 Discussion and conclusion

Previous global numerical models, observations and laboratory-derived creep laws suggest a moderate bulk viscosity contrast of 1-2 orders of magnitude between the asthenosphere beneath the oceanic lithosphere and the underlying mantle (Becker, 2017; Richards and Lenardic, 2018) and asthenosphere viscosity between $\sim 0.5 \times 10^{18}$ - 10^{21} Pa.s in ~ 200 km thickness (Mitrovica, 1996; Hu et al., 2016b; Freed et al., 2017). Such inferred large-scale asthenosphere weaknesses are generally explained by invoking the role of temperature or pressure; however, uncertainties on the true values remain, e.g. choice of appropriate rheologies and parameters. Also, the non-uniqueness relating to viscosity and thickness of an LVZ can confound the ability to distinguish between a thin layer of very low viscosity and a thick layer of greater viscosity (Cathles, 1975; Richards and Lenardic, 2018). Moreover, the trade-off between the resolving power and lateral extent of seismic studies may hinder the ability to distinguish narrow channels at the base of the lithosphere, such as that of a WL. Kawakatsu et al. (2009); Schmerr (2012); Naif et al. (2013); Stern et al. (2015) and Hawley et al. (2016) show that asthenospheric local and regional configurations may lead to up to a ≈ 4 orders of magnitude melt-induced viscosity reduction into one or multiple thin and more buoyant layers at the base of the lithosphere; e.g. Hawley et al. (2016) suggests that the enhanced melt content and/or presence of hydrated phases would have significantly lubricated the base of the subducting lithosphere of the Farallon slab with potential implications for subduction motions. These asthenospheric configurations seem to be compatible with global plates and seismic anisotropy models as long as their lateral extent remains limited compared to the rest of the plate (Becker, 2017).

In our study, we show that the presence of a thin WL at the base of the oceanic lithosphere critically affects subduction dynamics, particularly influencing the subduction speed, reducing trench migration rates and, in some cases, controlling the morphology assumed by a subducting slab. Our results show that for large viscosity contrasts ($\eta_M/\eta_{WL} \geq 10^2$) and a WL thickness greater than 20 km, the morphology of the subducting slab can be affected. In such cases, the subduction system may likely develop a quasi-stationary to an advancing moving trench with a moderate to fast convergence rate for the plate. For moderate viscosity contrasts between the mantle and the WL ($10^1 < \eta_M/\eta_{WL} < 10^2$), the subduction velocity increases and trench retreat rate decreases without significantly changing the overall shape of the slab. The overall effect is reduced where thinner and more buoyant layers exist (figure 4.2c, d and e, and 4.3b). This effect is constrained within end-member cases, and a description of the full range of slab behaviours is captured by the regime diagram in figure 4.3c. We also find that all the simulations with a WL are confined by velocity variations of a factor of ≈ 2 maximum, and all show a low absolute subduction partitioning component, i.e. $|V_T/V_P| < 40\%$ (figure 4.3a and b).

The overall impact of a WL on subduction dynamics depends on the relative contrast between the effective properties of the lithosphere and the mantle. We have proposed the effective relative stiffness of the lithosphere (D'), as a diagnostic parameter for this effect. D' not only describes the results found in this study and thus the effect of a WL but also includes plate strength values, which have previously been examined in a broad range of published numerical and analogical studies (e.g. Bellahsen et al., 2005; Di Giuseppe et al., 2008; Funiciello et al., 2008). All our experiments create a comprehensive subduction regime diagram that allows one to distinguish three subduction styles as a function of the relationship between D' , subduction motions ($V_T/V_P\%$), and slab morphologies and dips. This characterises and constrains the deformation in the system, as well as most of the variety of shapes, velocities and trench migration mechanisms we observe in nature (figure 4.3c). Although our study does not exclude the potential for alternative mechanisms, the role of the effective relative stiffness of the lithosphere seems to be of significance consistently with previous studies (Capitanio et al., 2007; Stegman et al., 2010) and due to lithosphere-mantle decoupling processes, thus affecting trench and plate motions at subduction zones.

The setup chosen has allowed us to demonstrate that low trench retreat and fast plate speed can be the result of a simpler force balance around the subducting plate alone, whilst no extra forces or factors are needed (e.g. an overriding plate and an upper-lower mantle discontinuity); however we do not exclude that other factors which have already shown to reduce trench motions may additionally affect or coexist with the mechanism illustrated here. Nonetheless, we also employed a small number of more complicated models to explore the extent to which our simplifications may influence our conclusions (supporting figure S3-S6.). We tested the role of a VEP T-dependent rheology, which would naturally lead to weakening at the base of the lithosphere, the inclusion of the upper/lower mantle transition zone and the effect of an extended modelled box. While rheology and choice of parameters can be different, the effect of a WL shows a consistent signature.

In the literature, many models have been made to describe subduction dynamics. While these models have provided useful insights toward the modelling of natural subduction systems and mantle convection processes, the modelling of subduction zones has not yet resolved clear trends for the subduction style, convergence velocity, trench motion, and slab dip, especially in the case of slow retreating or advancing trenches in natural settings (e.g. Garfunkel et al., 1986; Jarrard, 1986; King, 2001; Heuret and Lallemand, 2005; Lallemand et al., 2005; Sdrolias and Müller, 2006; Capitanio et al., 2007). Some more recent studies have also brought considerable attention to the effect of an asthenosphere of some origin in reducing trench retreat (e.g. Holt and Becker, 2016; Billen and Arredondo, 2018) and also enhancing plate motion when a ridge push condition is applied (e.g. Capitanio et al., 2007). This, along with the addition of an overriding plate

to the system seems to be relevant for reducing $|V_T/V_P|$ observed in numerical modelling (Rodríguez-González et al., 2012; Holt et al., 2015a,b; Holt and Becker, 2016).

Here, we suggest that the presence of a weak upper asthenosphere underneath subducting plates in a free subduction model is already alone of importance in decreasing the high absolute values of the subduction partitioning observed in many precedent analogues and numerical models and to level more in line with the average values presently seen on Earth. Specifically, in our models the average absolute trench velocity is approximately 30 per cent or lower of the total convergence plate velocity (figure 4.3d), as found in many plate motion reference frames, e.g. (Goscombe and Gray, 2008; Schellart et al., 2008; Becker and Faccenna, 2009). Furthermore, while a maximum increase in convergence velocity of a factor of ≈ 2 due to the weak layer is likely negligible in global models, its role becomes relevant at a regional scale influencing subduction motions and mantle flow dynamics. Thus, the regional mechanism proposed here provides a novel and explanation to interpret the apparent dichotomy in subduction partitioning values and slab dip angles globally observed along the edges of the Pacific Plate (Bellahsen et al., 2005; Lallemand et al., 2005; Coltice et al., 2017), such as the localities where the presence of a WL is confirmed by geophysical studies (e.g. Kawakatsu et al., 2009; Schmerr, 2012; Naif et al., 2013; Stern et al., 2015; Hawley et al., 2016).

4.8 Acknowledgments

We are grateful to A. Popov, A. Pusok and J. Yang for helpful discussions and inputs while at the Johannes Gutenberg University in Mainz. We thank T. Yang and other colleagues at Melbourne University for valuable comments. Additionally, we are grateful to two anonymous reviewers for their helpful comments. We further thank the co-authors and current code developers for providing the code Underworld2. Some computations were performed using Mogon, Raijin and Pawsey High-Performance Computing Clusters. This manuscript complies with FAIR Data standards. Software and code are open sources. We provide the DOI 10.5281/zenodo.1434224 of the 2.5.1 version of the Underworld2 code used to run our models and the DOI 10.5281/zenodo.3390014 which contains all our models scripts.

4.9 Supporting Information for “The impact of a very weak and thin upper asthenosphere on subduction motions”

Contents of this file

1. Methodology

2. Table S1, S2 and S3

3. Figures from F1 to F8

4.9.1 Methodology

This supporting information provides additional details about the methodology adopted and further material to support the conclusions proposed by this study.

4.9.1.1 Geometrical model set-up

Our models consist of a 2D Cartesian domain of 4000 km in length and 660 km in depth. The initial model setup is a further development of previous studies, (e.g. Enns et al., 2005; Stegman et al., 2006; Schmeling et al., 2008; Yamato et al., 2009; Schellart et al., 2010). It is made of four different layers. There is an oceanic crust, an oceanic tectonic plate, a weak upper asthenospheric layer (WL) and an upper mantle. Although similar, this setup differs from previous studies because the modelled oceanic plate is made of one single layer and the WL is characterized by a lower viscosity value than in the underlying asthenosphere or mantle. The initial model geometry includes an oceanic lithosphere with an additional slab length of 256 km dipping 34 degrees into the underlying mantle to initiate progressive subduction, as illustrated in figure 4.4. We use a model domain of 256×128 elements with a uniform grid spacing in each coordinate direction. This gives a grid resolution of 5.2 km in the vertical direction and a horizontal resolution of 15.6 km. This ensures sufficient numerical resolution to promote and resolve the development of a weak hinge zone. The weak hinge crustal zone is characterized by a visco-plastic rheology which simulates free surface and enables the oceanic lithosphere to freely detach from the top surface of the model and subduct into the mantle (Cramer et al., 2012). The geometrical setup is based on (Schmeling et al., 2008) and there is shown that the model space size and the free-slip boundary conditions do not affect the velocities largely or the induced flow significantly. Therefore, we apply free-slip velocity boundary conditions everywhere which minimize the influence of the box sidewalls (a free-slip condition which means there is no tangential stress and zero normal velocity). We do not apply any further conditions and subduction is dynamically initiated. To demonstrate that the vicinity of the slab to the front wall has a negligible effect on our outcomes of subduction partitioning we perform a series of experiments shown in supporting figure 4.9. Figure 4.9 shows a comparison between i) our current setup ($L_{BOX} = 4000$ km); ii) an extended model ($L_{BOX} = 6000$ km), iii) an extended model ($L_{BOX} = 6000$ km) where we slightly increase the resolution to 384×128 cells; and iv) a model with periodic boundary conditions at the sidewalls. These simulations were performed for the RM (no WL) and the WL cases. Further details about the parameters varied are available in table 4.9.2. These results show that if expand the modelled box and do not change the numerical resolution, this causes

a slightly lower resolution for the slab region (figure 4.9a) and the small discrepancies we observe in plate velocities over time (figure 4.9a); however these do not significantly affect our outcomes or the partitioning of plate and trench motion (figure 4.9 c). Moreover, if we keep the size of the grid cell the same (approx. 15 km as for the default setup used in the paper), implying that we use more grid points in the wider box, these differences become completely negligible (figure 4.9a). Furthermore, if we apply periodic boundary conditions at the side walls and maintain the same resolution as for our current setup (256*128), this again show that the vicinity of the slab to the front wall has a negligible effect on our outcomes (figure 4.9a, b and c). While there are some minor differences between the cases, the results consistently show that the presence of a WL strongly affects plate and trench motion, and their partitioning at subduction zones, while the size of the model box remains a second order effect.

We test, show and describe the reliability and reproducibility of our reference model (RM) against previous work (e.g. Funiello et al., 2004; Schellart, 2004a; Stegman et al., 2006; Capitanio et al., 2007; Schellart et al., 2007; Goes et al., 2008; Capitanio et al., 2009), see figure 1 in the main manuscript and figure 4.4 and in this supporting documentation. Therefore, we chose the RM as a base for comparison with the simulations having a WL. In the interest of simplicity, we avoid considering the presence of other factors that could influence plate and trench motions and strictly relate to locale configurations. Among these, there are the presence of an overriding plate, the upper/lower mantle transition zone, complex mantle rheology and the lateral extension of a slab (e.g. Schellart et al., 2007; Capitanio et al., 2010b; Garel et al., 2014; Holt and Becker, 2016). This very setup is necessary to show that low trench retreat and fast plate speed can be the result of a simpler force balance around the subducting plate and no extra forces are needed. Also, we do not consider the presence of an overriding plate, as this helps the single plate to subduct freely and in isolation, and is not affected by induced mantle flow from neighbouring slabs or large-scale convection. In addition, recent global statistical investigations also show that the role of the overriding plate in determining the trench velocity is of subordinate importance (Schellart, 2008b) and the subducted slab predominantly determines the trench velocity. Furthermore, while the role of slabs' interactions with the lower mantle and the role of complex mantle and lithosphere rheologies remain out of the scope of this paper, in this supporting information we show that independently by the choice of parameters, the effect of a WL shows a consistent signature. This can be seen in some of our models that allow for some of this further complexity and still validate the conclusions proposed by this study (supporting figure 4.4, 4.5, 4.6, 4.7 and 4.8).

4.9.1.2 Pseudo-plastic (PP) Models

We consider subduction and plate motions in an incompressible Boussinesq fluid at infinite Prandtl number and very low Reynolds number. We solve the Stokes equation that

balances buoyancy forces with viscous stresses and pressure gradients.

The numerical implementation is the particle in cell finite element method in Underworld in which material properties are carried by Lagrangian integration points and mapped to element properties through a local discrete-Voronoi integration scheme. This numerical approach is designed for accurately tracking distinct material domains in the lithosphere as they undergo large deformation due to creeping flow. Here, we solve the conservative equations for mass and momentum together with their constitutive relationships. The non-dimensional equations of motion for an incompressible fluid are:

$$\frac{\partial v_i}{\partial x_i} = 0 \quad (4.3)$$

and

$$\frac{\partial \tau_{ij}}{\partial x_j} - \frac{\partial p}{\partial x_i} + \rho g_i = 0 \quad (4.4)$$

Where i and j represent spatial directions and follow the Einstein summation convention, $\tau_{ij} = \sigma_{ij} + p\delta_{ij}$ is the Cauchy stress deviator, ρ is the fluid density, v_i is the local velocity vector of a material, g_i is the gravitational acceleration and x is the spatial coordinate. Further details describing the non-dimensional equations and the numerical technique in underworld can be found in Moresi et al. (2003) and Stegman et al. (2006)

The rheology used for the majority of the experiments performed in this study is Newtonian in all but the uppermost layer (representing the oceanic crust and any associated sedimentary layers) which is visco-plastic. The visco-plastic crust acts as a lubricating phase to simulate the free surface and to allow the hinge to detach from the surface (Schmeling et al., 2008; Crameri et al., 2012). We refer to these models as pseudo-plastic (PP).

When a Newtonian constitutive relationship is used (for an incompressible fluid) the relationship between the stress and strain rate is defined as:

$$\tau'_{ij} = 2\eta^{vs}\dot{\varepsilon}_{ij} \quad (4.5)$$

being τ'_{ij} the deviatoric stress and $\dot{\varepsilon}_{ij} = \frac{1}{2} \left(\frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i} \right)$ the strain rate tensor.

When plastic failure is considered for the uppermost layer, the total effective viscosity is taken as the minimum between the viscous and the plastic terms, as follow:

$$\eta_{\text{eff}} = \min[\eta^{vs}, \eta^{\text{pl}}] \quad (4.6)$$

We assume Byerlee's law which describes the increase in the maximum strength of the upper crustal rocks with increasing confining pressure (depth). We also assume an approximation of the Drucker-Prager yield criterion, the so-called depth-dependent (henceforth

denoted as DD), see Moresi and Solomatov (1998); Tackley (2000); Stein et al. (2004) for reference, which introduces a friction angle and depends only on the lithostatic pressure and neglects the dynamic component. In this case, the yield stress, τ_y , and the effective viscosity of the plastic branch, η_{pl} , are implemented as follow:

$$\tau_y = \tau_0 + \mu p_{lith} \quad (4.7a)$$

$$\eta_{pl} = \frac{\tau_y}{2\dot{\epsilon}_{II}} \quad (4.7b)$$

being $\dot{\epsilon}_{II} = \left(\frac{1}{2}\dot{\epsilon}_{ij}\dot{\epsilon}_{ij}\right)^{(1/2)}$ the second invariant of the strain rate, τ_0 the cohesive strength, μ the friction coefficient and $p_{lith} = \rho g z$ the reference lithostatic pressure at depth z . We use a cohesion of a 20 MPa, a friction angle of 30 degrees and a density contrast between the oceanic lithosphere and the underlying mantle of 80 kg/m^3 , in line with the values calculated in the detailed study of Cloos (1993).

We neglect temperature diffusion as the duration of the experiment is approximately 20 million years and we are modelling the evolution of an established subduction zone (see Toth and Gurnis (1998); Regenauer-Lieb et al. (2001); Gurnis et al. (2004) for subduction initiation). We also assume that the boundary layer can be represented by a plate of fixed thickness, as this does not significantly alter over the duration of the experiment for a plate of 70 million years old. The thermal age of the oceanic lithosphere is used to compute its initial thickness (Turcotte and Schubert, 2014) for a 70 Myr old plate, but we also perform simulations corresponding to a lithospheric thickness of 50 and 120 km. The sets of parameters used and varied are shown in table 4.9.2 and additional results are shown in supporting figure 4.4 and 4.6. Furthermore, since a visco-elastic-plastic (VEP) rheology can naturally lead to a layer of low viscosity the base of the subducting lithosphere (Billen and Hirth, 2005), we also perform a small number of more complicated models to explore the extent to which our simplifications may or may not influence our conclusions. For this further set of numerical simulation, we use a full half-space cooling model temperature dependent approach and implement VEP rheology for all the phase involved, as described below. These further calculations support and are in agreement with the main conclusions proposed by this study.

4.9.1.3 Visco-Elastic-Plastic (VEP) T-dependent models

Our VEP T-dependent rheology calculations assume that temperature plays a role in establishing the initial density, strength distribution and thickness of the subducting lithosphere and upper mantle. The temperature structure of the oceanic plate is a full half-space cooling model consistent with a conductive boundary layer but constant in the underlying asthenosphere and deeper mantle. In this case, the solution of the equation is given by:

$$T(z, t) = T_M \operatorname{erf} \frac{z}{2\sqrt{kt}} \quad (4.8)$$

with T_M being the temperate of the up-welling mantle, k the thermal diffusivity that we consider isotropic, and t the thermal age of the oceanic lithosphere. The thermal age of the oceanic lithosphere is used to compute its initial thickness, such as $h_L = 2.32(kt)^{1/2}$ (Turcotte and Schubert, 2014) with k 1 mm/s² at an age of 70 million years. We also perform calculation considering plates of variable age of 50 and 100 million years, as shown in figure 4.5 and table 4.3. We take density anomalies to be temperature dependent:

$$\Delta\rho = \alpha\rho_0\Delta T \quad (4.9)$$

with α being the thermal expansion ($3 \times 10^{-5} \text{K}^{-1}$) and $\Delta\rho$ the density difference from the reference state ρ_0 . The total weight of a column of the lithosphere is computed by integrating its density in equation (4.9), where $\Delta T = T - T_m$ and T is substituted by equation 4.8 (e.g. Becker and Kaus, 2010).

We model crust, oceanic plate, WL and underlying mantle as visco-elastic-plastic materials. The visco-elastic-plastic constitutive equation for the deviatoric stress is given by:

$$\dot{\varepsilon}_{ij} = \dot{\varepsilon}_{ij}^{\text{vs}} + \dot{\varepsilon}_{ij}^{\text{el}} + \dot{\varepsilon}_{ij}^{\text{pl}} = \dot{\varepsilon}_{II}^{\text{vs}} \frac{\tau_{ij}}{\tau_{II}} + \frac{\Delta}{2G} \frac{\tau_{ij}}{\tau_{II}} + \dot{\varepsilon}_{II}^{\text{pl}} \frac{\tau_{ij}}{\tau_{II}} \quad (4.10)$$

where $\dot{\varepsilon}_{ij} = \frac{1}{2} \left(\frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i} \right) - \frac{1}{3} \frac{\partial v_k}{\partial x_k} \delta_{ij}$ is the deviatoric strain rate tensor, $\dot{\varepsilon}_{ij}^{\text{vs}}$, $\dot{\varepsilon}_{ij}^{\text{el}}$, $\dot{\varepsilon}_{ij}^{\text{pl}}$ are the viscous, elastic and plastic components, respectively. $\frac{\Delta}{\tau_{ij}} = \frac{\partial \tau_{ij}}{\partial t} + \tau_{ik} w_{kj} - w_{ik} \tau_{kj}$ is the Jaumann objective stress rate, $w_{ij} = \frac{1}{2} \left(\frac{\partial v_i}{\partial x_j} - \frac{\partial v_j}{\partial x_i} \right)$ is the spin tensor, G is the elastic shear modulus and the subscript II denotes the square root of the second invariant of a corresponding tensor. In our study, the viscous diffusion and dislocation creep mechanisms of the oceanic crust, lithosphere and mantle are based on the parameters of dry olivine from Hirth and Kohlstedt (2003). The viscous rheology of the WL is computed either as Newtonian or as a combination of diffusion and dislocation creep processes for a wet olivine material (Hirth and Kohlstedt, 2003). The magnitude of the viscous creep strain rate takes into account both diffusion ($\dot{\varepsilon}^{\text{diff}}$) and dislocation ($\dot{\varepsilon}^{\text{disl}}$) creep mechanisms as:

$$\dot{\varepsilon}^{\text{vs}} = \dot{\varepsilon}^{\text{diff}} + \dot{\varepsilon}^{\text{disl}} = A_{\text{diff}}(\tau_{II}) + A_{\text{disl}}(\tau_{II})^n \quad (4.11)$$

with n being the stress exponent of the dislocation creep. The pre-exponential factor (A) of each creep component is defined as:

$$A_{\text{diff}} = B_{\text{diff}} \exp \left[-\frac{E_{\text{diff}} + PV_{\text{diff}}}{RT} \right], \quad A_{\text{disl}} = B_{\text{disl}} \exp \left[-\frac{E_{\text{disl}} + PV_{\text{disl}}}{RT} \right] \quad (4.12)$$

B, E, and V denote the creep constant, activation energy and activation volume, respectively, of the corresponding creep mechanisms and R is the gas constant. In addition, the creep constant (B) also takes into account the effect of grain size, H₂O and tensor correction (given that the creep-laws are typically measured under uniaxial or simple shear), as defined in Gerya (2009).

We employ a Maxwell visco-elastic relationship (equation 4.10) which has been used in studies of lithospheric deformation before Peltier (1974); Schmalholz and Podladchikov (2001) following previous work (e.g. Kaus et al., 2016; Pusok et al., 2018). In such a system, the effective visco-elastic viscosity is computed as:

$$\eta^{ve} = \eta_{min} \left[\frac{1}{\left(\frac{1}{\eta_{diff}} + \frac{1}{\eta_{disl}} \right), \frac{1}{\eta_{el}} + \frac{1}{\eta_{max}}} \right] \quad (4.13)$$

η_{min}, η_{max} are the minimum and maximum viscosity cut-off values of 10^{18} Pa·s and 10^{24} Pa·s respectively, which ensure the numerical stability of the non-linear solution (iterate between $\dot{\varepsilon}_{II}^*$ and τ_{II}). The individual creep and elastic viscosities become:

$$\eta_{diff} = \frac{1}{2}(A_{diff})^{-1}, \quad \eta_{disl} = \frac{1}{2}(A_{disl})^{-\frac{1}{n}}(\dot{\varepsilon}_{II}^*)^{\frac{1}{n}-1}, \quad \eta_{el} = \frac{1}{G\Delta t^e} \quad (4.14)$$

with Δt^e being the elastic time step, $\dot{\varepsilon}_{II}^* = (\dot{\varepsilon}_{ij}^* \dot{\varepsilon}_{ij}^*)^{1/2}$ the second invariant of the effective strain rate $\dot{\varepsilon}_{ij}^* = 2\dot{\varepsilon}_{ij}^{*(t+\Delta t)} + \frac{1}{G\Delta t^e}\tau_{ij}^t + \frac{1}{\mu}(w_{ik}\tau_{kj} - \tau_{ik}w_{kj})$, in which the subscripts t and Δt^e indicate values at the current and future time step, respectively. In our calculations we omit the spin tensor w_{ij} and assume it not fundamental to the extent of the problem investigated (Moresi et al., 2003). The implementation of viscoelasticity follows Moresi et al. (2003); Farrington et al. (2014) and modifies the momentum equation. Such as that, during the advection, the stress field of the future time step is corrected to account for an effective viscosity, an additional force term dependent on material properties, the time step size, and the stress history. The updated momentum equation coupled with the conservation of mass provides the solution for a Maxwell viscoelastic material deforming under the Boussinesq approximation for infinite Prandtl, low Reynolds incompressible flow. The plastic failure of rocks is approximated by a DD Coulomb failure criterion, as described in equation 4.7a and combined this with the definition of lithostatic pressure given in equation 4.9.

The effective viscosity, η^* , and the updated deviatoric stresses, τ_{ij} , are computed from the effective strain rates, using the standard quasi-viscous expression:

$$\tau_{ij} = 2\eta^* \dot{\varepsilon}_{ij}^*, \quad \eta^* = \min \left[\left(\frac{1}{\eta_{ve}} \right)^{-1}, \frac{\tau_y}{2\dot{\varepsilon}_{II}^*} \right] \quad (4.15)$$

with ε_{II}^* ensuring yielding for stresses equal to or greater than τ_y . Then, the value of η^* is iterated to allow stress to redistribute from points which become unloaded. The iteration is repeated until the velocity solution is unchanged to within the error tolerance required for the solution as a whole.

4.9.1.4 Scaling Analysis

The degree to which the motion of plates is promoted or impeded with respect to the flow in the underlying mantle depends upon the type of flow within the asthenosphere. This flow can be driven by the plates themselves (Couette flow), pressure driven by the flow within the mantle (Poiseuille flow) or a combination of the two (combined Couette-Poiseuille flow). Several authors (e.g. Ribe, 1992; Conrad and Hager, 1999; Ribe, 2001, 2010; Stegman et al., 2010; Buffett and Becker, 2012; Gerardi and Ribe, 2018; Semple and Lenardic, 2018; Gerardi et al., 2019) show that the force balance of the oceanic lithosphere is controlled by the balance of both bending-stretching momentum coupled with boundary and internal stresses (flow in the boundary layer).

Among these authors (e.g. Ribe, 1992, 2001; Gerardi and Ribe, 2018; Richards and Lenardic, 2018; Gerardi et al., 2019) show that the motion of a mantle fluid in an interface layer, can be approximated with the motion of a Stokes fluid in the lubrication limit. This procedure is valid when the thickness of the layer is much thinner than the horizontal length, if $l_{WL}/h_{WL} \ll 1$. An approximation of the tangential forces representing the motion of a Stokes's fluid in a thin layer in the lubrication limit led to the definition of the dimensionless stiffness of a surface shell, f , (Ribe, 1992, 2001) and an interface layer, γ , (Gerardi and Ribe, 2018; Gerardi et al., 2019) as:

$$f = \frac{h_s}{h_{bg}} \left[\frac{\eta_s}{\eta_{bg}} - 1 \right], \quad \gamma = \frac{\eta_{in}}{\eta_{bg}} \frac{h_P}{h_{in}} \quad (4.16)$$

where η_s , η_{bg} and η_{in} indicate the viscosity of the shell, background fluid/ mantle and interface, and with h_s , h_{bg} , h_P and h_{in} being thicknesses of the shell, background fluid/ mantle, plate and interface, respectively. In our study, we aim to address the role of a very thin and weak upper asthenosphere on subduction dynamics, and particularly on the partitioning of plate and trench motions at subduction zones. Hence, we are interested in describing the deformation in the effective lithosphere relative to the underlying mantle, which may or may not contain a WL, on the partitioning of trench and plate motions. Based on our numerical experiments, we represent a first order approximation of the deformation occurring in the system (with and without a WL) using a series of parameters related to the dimensionless stiffness of the effective lithosphere.

We empirically find that the effective relative strength, S' , and thickness, h' , of the lithosphere relative to the underlying mantle (figure 3a and b, and figure 4.10a and c) can be taken into account as:

$$S' = \frac{\eta_P \eta_{WL}}{\eta_M^2}, \quad h' = \frac{h_P}{h_M} \left(1 + \frac{h_{WL}}{h_M} \right) \quad (4.17)$$

where η_P , η_{WL} and η_M indicate the viscosity of the plate, weak layer and mantle, and with h_P , h_{WL} and h_M being thicknesses of the plate, the WL and the mantle respectively. Note that when a WL is not included $\eta_{WL} = \eta_M$ and $h_{WL} = 0$, allowing us to characterise all the experiments with and without a WL. We also find that the combined effect of the relative contrast in strength and thickness between the effective lithosphere and the underlying mantle can be expressed by the effective relative stiffness of the lithosphere, D' , as:

$$D' = \log(S')h' = \log \left(\frac{\eta_P \eta_{WL}}{\eta_M^2} \right) \frac{h_P}{h_M} \left(1 + \frac{h_{WL}}{h_M} \right) \quad (4.18)$$

Our study represents a balance of forces problem where the subduction dynamics are governed by the balance of stretching moment and internal and boundaries stresses. This has shown to critically control the dip and curvature of slabs, and hence the partitioning of surface subduction motions (figure 3 in the main manuscript), providing new insights on the role of a weaker upper asthenosphere on subduction dynamics. D' allows us to fit our experiments into a regime diagram that shows the combined effects of viscosity and thickness variations in the effective lithosphere (with and without a WL), the partition of subduction motions (V_T/V_P) and the slabs' morphologies and dips.

Quantities related to the resistance to shearing and the mechanical strength of the lithosphere or asthenosphere have been described in previous work (e.g. Karato et al., 2001; Ribe, 2001; Stegman et al., 2010; Turcotte and Schubert, 2014; Richards and Lenardic, 2018). Therefore, we compare our approach with others which would favour the bending or the shearing at the base of the lithosphere as the dominant deformation mechanism to define the impact of a WL on subduction motions. In our study, we refer to the effective relative flexural stiffness of the lithosphere, D^* , as:

$$D^* = S' h'^3 = \left(\frac{\eta_P \eta_{WL}}{\eta_M^2} \right) \left[\frac{h_P}{h_M} \left(1 + \frac{h_{WL}}{h_M} \right) \right]^3 \quad (4.19)$$

We also refer to the effective stress at the base of the lithosphere, τ^* , as:

$$\tau^* = \frac{S'}{h'} = \frac{\left(\frac{\eta_P \eta_{WL}}{\eta_M^2} \right)}{\frac{h_P}{h_M} \left(1 + \frac{h_{WL}}{h_M} \right)} \quad (4.20)$$

While the cotangent trend that defines the relationship between V_T/V_P and S' or $1/h'$ (figure 4.10b and c) does not vary significantly when plotted against $\log(S')$ or $1/h'^3$ (figure 4.10 a and d), the comparison between D' (figure 4.11a) with others mechanisms that

would favour the bending D^* (figure 4.11b) or the shearing at the base of the lithosphere, τ^* (figure 4.11c) demonstrates how WL's effects and deformation are better described by D' . D' represents an approximation of the tangential forces acting on a subducting lithosphere (with and without a WL), which has shown to be the dominant deformation mechanism for the decoupling effect of a WL at the base of the lithosphere. D' allows us to capture and constrain three subduction styles, which characterise the deformation in the system with and without a WL, as well as most of the variety of shapes, dips and subduction partitioning mechanisms we observe in nature. By contrast, these cannot be adequately represented using D^* or τ^* as a relationship that link either of these variables to V_T/V_P , and slabs' morphologies and dips does not emerge with a clear signature (figure 4.11 b and c). These observations suggest a WL mechanism whereby the style of subduction and the subduction partitioning are controlled by the type of flow active in the effective lithosphere. Although our study does not exclude the potential for alternative mechanisms, our results suggest that a WL provides a regional mechanism to explain the slow rates of trench migration and fast-moving plates that we observe globally at natural subduction zones.

4.9.1.5 Velocity Calculations

In all our sets of simulations, we extract plate velocity, V_P , as the mean of the magnitude velocity within the horizontal section of the oceanic plate. Trench migration rates, V_T , are calculated as the difference in the trench X coordinate between the old and new time step. Our subduction partitioning calculations (V_T/V_P) use the mean of V_P and V_T calculated over the first subduction stage of the evolution of the models (≈ 6 Myr). While V_T/V_P resulting from later stages of the evolution in time of the models can be polluted by the interaction of the dip of the slab with the bottom of the model, the subduction motions resulting from the initial stage of the evolution of the models are purely the result of subduction dynamics and allow us to retrieve a representative value of V_T and V_P , see figure 4.7.

4.9.2 Table S1, S2 and S3.

4.9.3 Figure S1 to S8

PP Model parameters varied per set of simulation in this study									
# Set	Model ID	η_M/η_{WL}	$\Delta\rho_{(M-WL)}$ [kg/m ³]	h _{WL} [km]	h _{OL} [km]	η_{OL} [Pa s]	η_M [Pa s]	η_{OL}/η_M	α
1	RM, M1, M2, M3 M8,M9,M10 M7	10 ⁰ , 10 ¹ , 10 ² , 10 ³	0	30	106	2.5 × 10 ²³	2.5 × 10 ²¹	10 ²	30
		5 × 10 ¹ , 5 × 10 ² , 10 ⁴	0	30	106	2.5 × 10 ²³	2.5 × 10 ²¹	10 ²	30
		10 ²	20	30	106	2.5 × 10 ²³	2.5 × 10 ²¹	10 ²	30
2	M4,M5,M6 M11,M12	10 ²	0	10,20,80	106	2.5 × 10 ²³	2.5 × 10 ²¹	10 ²	30
		10 ²	0	50,150	106	2.5 × 10 ²³	2.5 × 10 ²¹	10 ²	30
3	M13, M14 M15,M16	10 ⁰ , 10 ²	0	0,30	50	2.5 × 10 ²³	2.5 × 10 ²¹	10 ²	30
		10 ⁰ , 10 ²	0	0,30	150	2.5 × 10 ²³	2.5 × 10 ²¹	10 ²	30
4	M17, M18 M19,M20	10 ⁰ , 10 ²	0	0,30	106	2.5 × 10 ²⁴	2.5 × 10 ²¹	10 ³	30
		10 ⁰ , 10 ²	0	0,30	106	5 × 10 ²³	2.5 × 10 ²¹	5 × 10 ²	30
5	M21, M22 M23,M24	10 ⁰ , 10 ²	0	0,30	106	2.5 × 10 ²³	1 × 10 ²¹	2.5 × 10 ²	30
		10 ⁰ , 10 ²	0	0,30	106	2.5 × 10 ²³	1.67 × 10 ²¹	1.5 × 10 ²	30
6	M25, M26 M27,M28	10 ⁰ , 10 ²	0	0,30	106	2.5 × 10 ²³	2.5 × 10 ²¹	10 ²	0
		10 ⁰ , 10 ²	0	0,30	106	2.5 × 10 ²³	2.5 × 10 ²¹	10 ²	15
7	M29,M30 M31, M32	10 ⁰ , 10 ²	0	0,30	106	1 × 10 ²³	1 × 10 ²¹	10 ²	0
		10 ⁰ , 10 ²	0	30	106	1 × 10 ²³	1 × 10 ²¹	10 ²	30

Table 4.1: Parameters used and varied for the PP rheological experiments.

Supporting PP Model parameters varied per set of simulation in this study											
# Set	Model ID	η_M/η_{WL}	h_{WL} [km]	h_{OL}	η_{OL}/η_M	α	η_{UM}/η_{LM}	L_{BOX}	Res	Side walls BCs	Surface walls BCs
8	M33,M34	$10^0, 10^2$	0,30	106	10^2	30	30	4000	256*192	free slip	free slip
9	M35,M36	$10^0, 10^2$	0,30	106	10^2	30	-	6000	256*128	free slip	free slip
	M37, M38	$10^0, 10^2$	0,30	106	10^2	30	-	6000	384*128	free slip	free slip
10	M39,M40	$10^0, 10^2$	0,30	106	10^2	30	-	4000	256*128	periodic	free slip

Table 4.2: Parameters used and varied for the PP rheological experiments provided in this supporting information. Res and BCs refer to numerical resolution and model boundary conditions, respectively.

VEP parameters varied per material modelled in this study								
# ID	Material	ρ_0 [kg/m ³]	C_p [J/kg/K]	k [W/m/K]	T Profile	Viscous Flow Law	C [MPa]	α G [Pa]
0 OC	Crust	3000	914	3	Half-Space (50-70-100 Myr)	Hirth and Kohlstedt (2003) Dry Olivine	20	30×10^{10}
1 OL	Lithosphere	3250	914	3	Half-Space (50-70-100 Myr)	Hirth and Kohlstedt (2003) Dry Olivine	20	30×10^{10}
2 WL	Weak Layer	3250	914	3	0 [°C/km]	Hirth and Kohlstedt (2003) Wet Olivine/ Const	20	30×10^{10}
3 UM	Mantle	3250	914	3	0 [°C/km]	Hirth and Kohlstedt (2003) Dry Olivine	20	30×10^{10}

Table 4.3: Parameters used and varied for the VEP rheological experiments. Density linearly depends on temperature and depth.

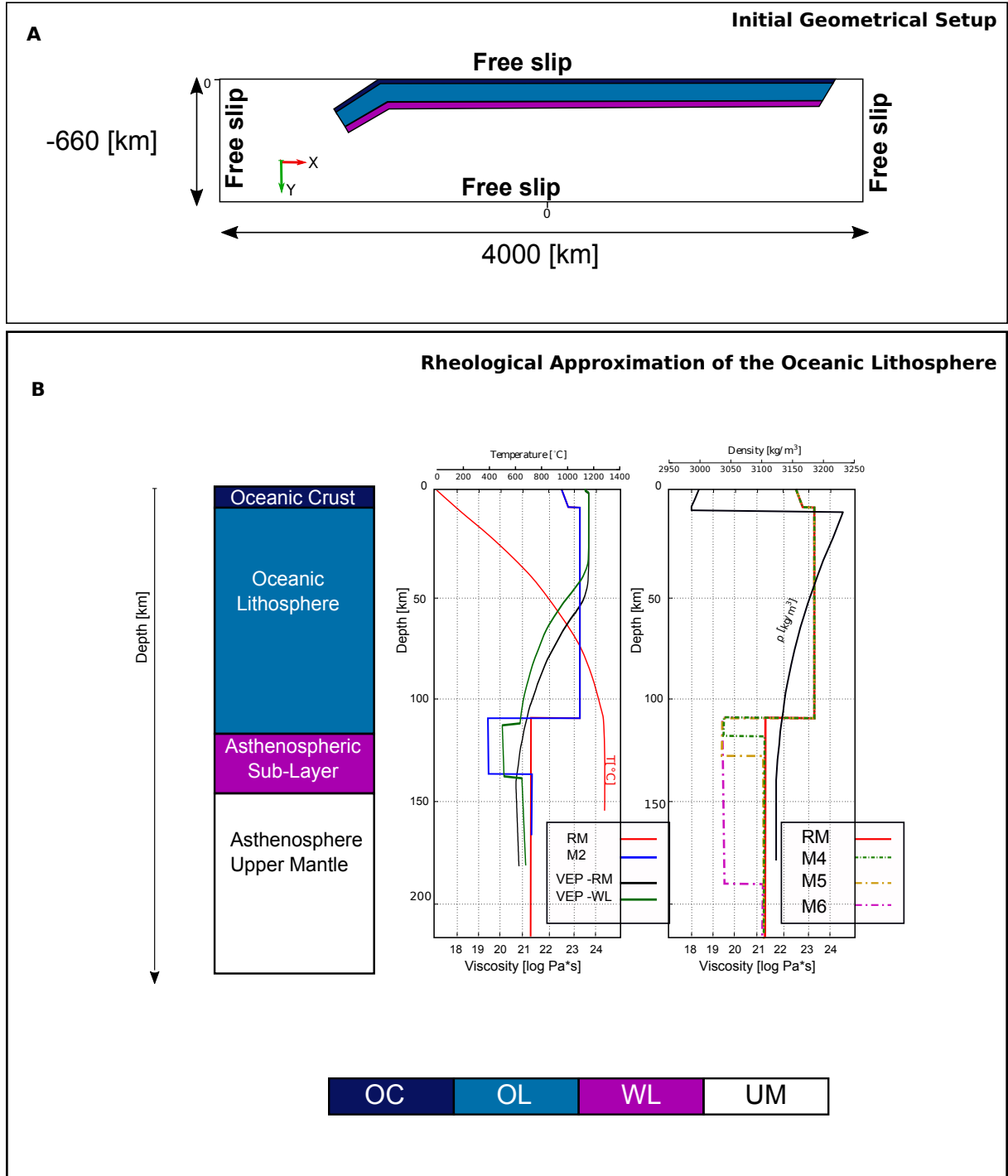


Figure 4.4: A) A schematic representation of the geometrical setup for the models with a weak layer (WL) in A. B) The material field through depth, the initial rheological setup for the pseudo-plastic (PP) and the visco-elastic-plastic T-dependent (VEP) experiments with different viscosity contrast mantle and WL, the density and the temperature profiles and the initial setup for models with different WL thickness.

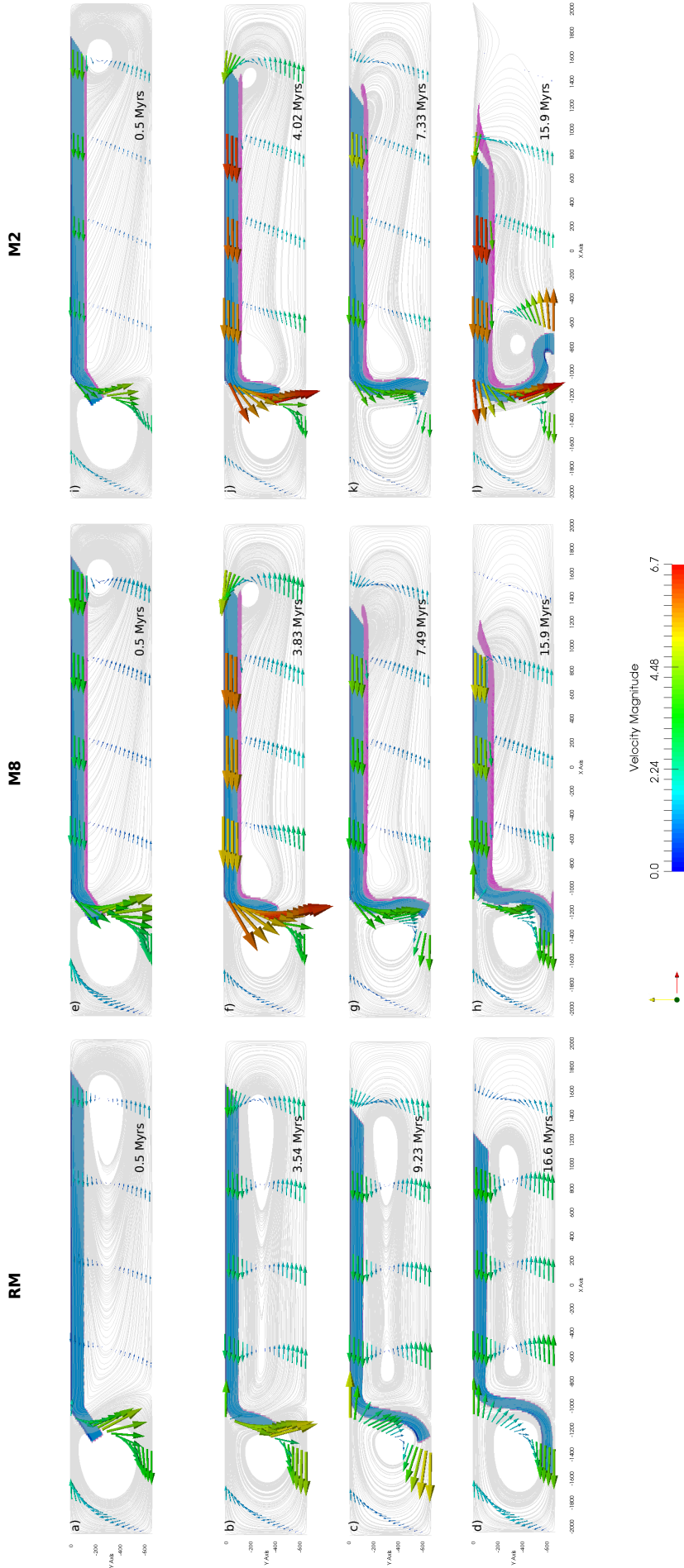


Figure 4.5: Time evolution of subduction models shown in a comparison between the RM ($\eta_M/\eta_{WL} = 1$), M8 ($\eta_M/\eta_{WL} = 5 \times 10^1$) and M2 ($\eta_M/\eta_{WL} = 10^2$). This shows the typical three stages of the evolution in time of subduction zone: incipient subduction zone (b, f, j), interaction of the slab tip with the bottom of the model (c, g, k) and steady state stage (d, h, l). M8 and M2 show higher velocity for the plate, a different trench migration mechanism and a different slab morphology and dip. The velocity arrows size of the third column (M2) was decreased to guaranty an appropriate representation of what occurs in all the models in the figure.

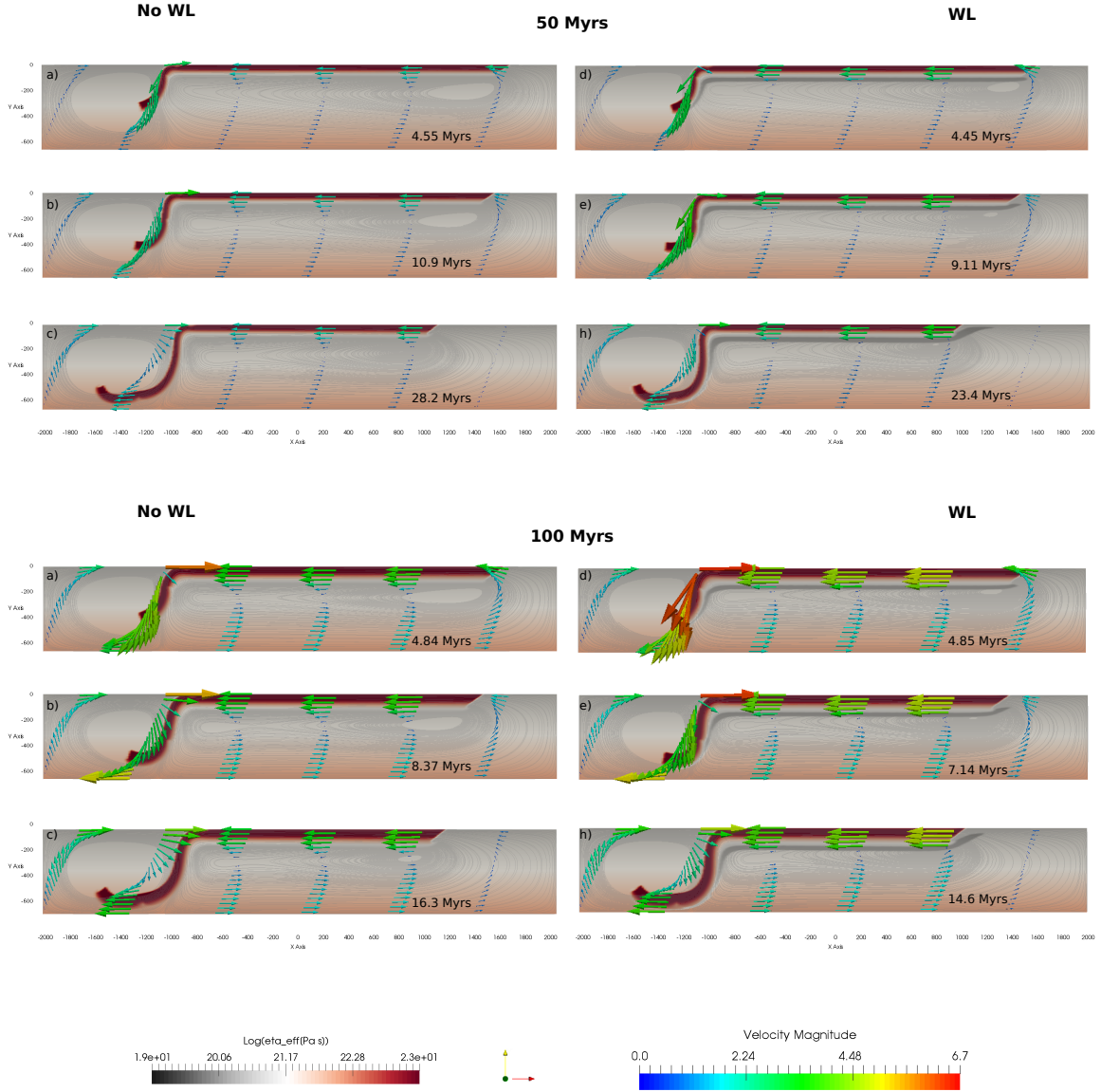


Figure 4.6: Time evolution of subduction models in a comparison between a 50 and 100 million years old oceanic lithosphere with visco-elastic-plastic (VEP) T-dependent rheology. Experiments on the left side do not include a weak layer and experiments on the right include a WL with a wet olivine rheology. We show the typical three stages of the evolution in time of subduction zone: incipient subduction zone, the interaction of the slab tip with the bottom of the model and steady state stage. Models on the right side (with a WL) show higher velocity for the plate, different trench migration mechanisms and different slab morphologies and dips.

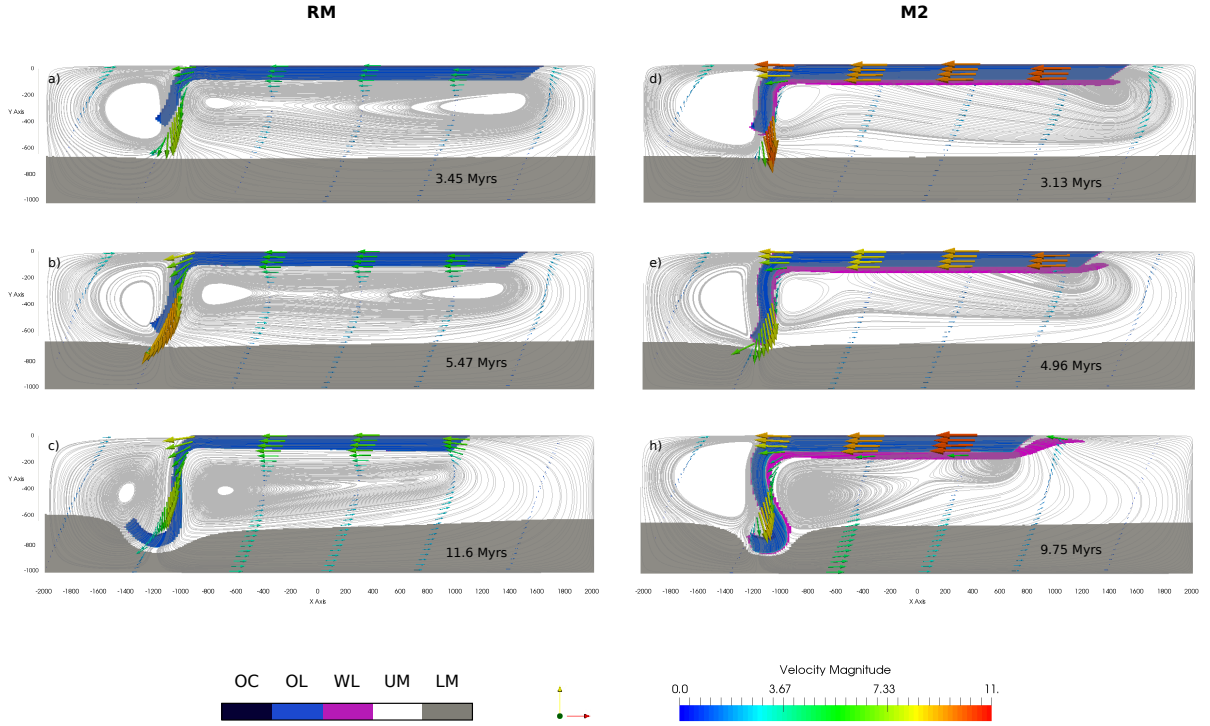


Figure 4.7: Time evolution of subduction models with a lower mantle (LM) shown in a comparison between the RM ($\eta_M/\eta_{WL} = 1$) and M2 ($\eta_M/\eta_{WL} = 10^2$). The upper-lower mantle transition zone is characterised by a viscosity contrast of 30. The evolution of the models show again the typical three stages of the evolution in time of subduction zone: incipient subduction zone, interaction of the slab tip with the bottom of the model and steady state stage. M2 show higher velocity for the plate, different trench migration mechanism and different slab morphology and dip.

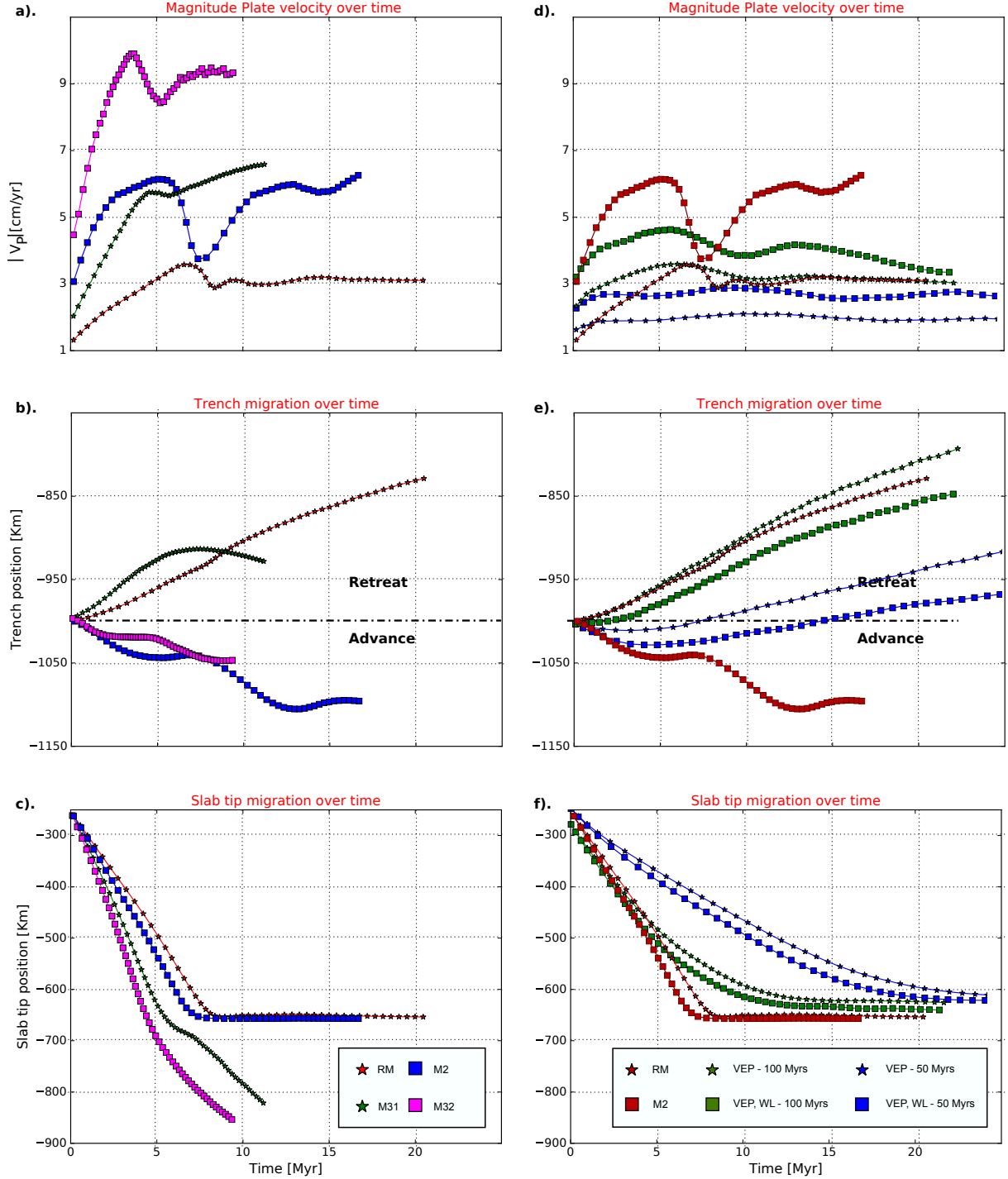


Figure 4.8: Evolution of plate velocities (V_P), trench and slab tip migrations through time for simulations with and without the presence of the lower mantle in a), b), c), and for simulations with PP and VEP rheology with and without a WL in d), e) and f).

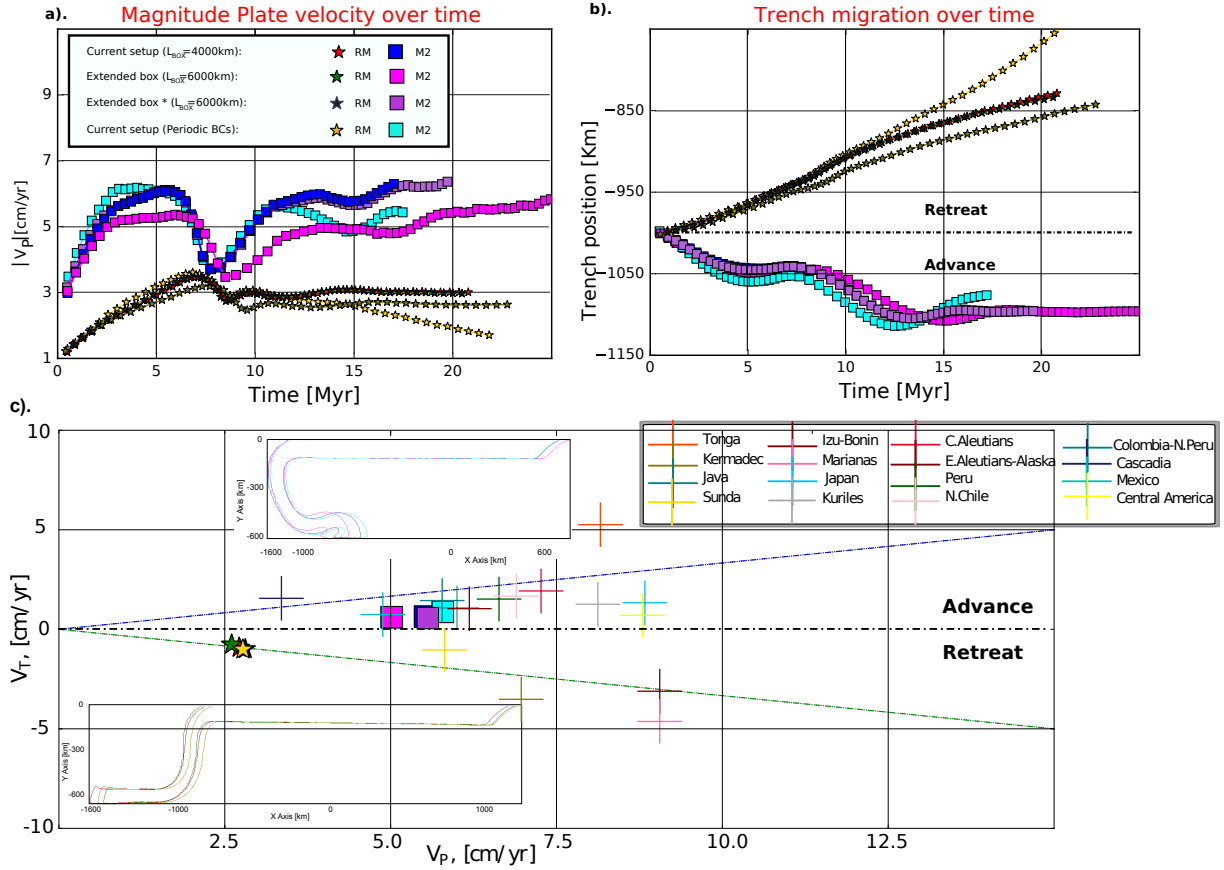


Figure 4.9: Evolution of plate velocities (V_P) and trench migrations through time for the default, the extended and the periodic boundary conditions model setups without WL (RM, $\eta_M/\eta_{WL} = 1$) and with a WL (M2, $\eta_M/\eta_{WL} = 10^2$) in a) and b), respectively. In c) V_T and V_P for the experiments considered in a comparison to natural cases.

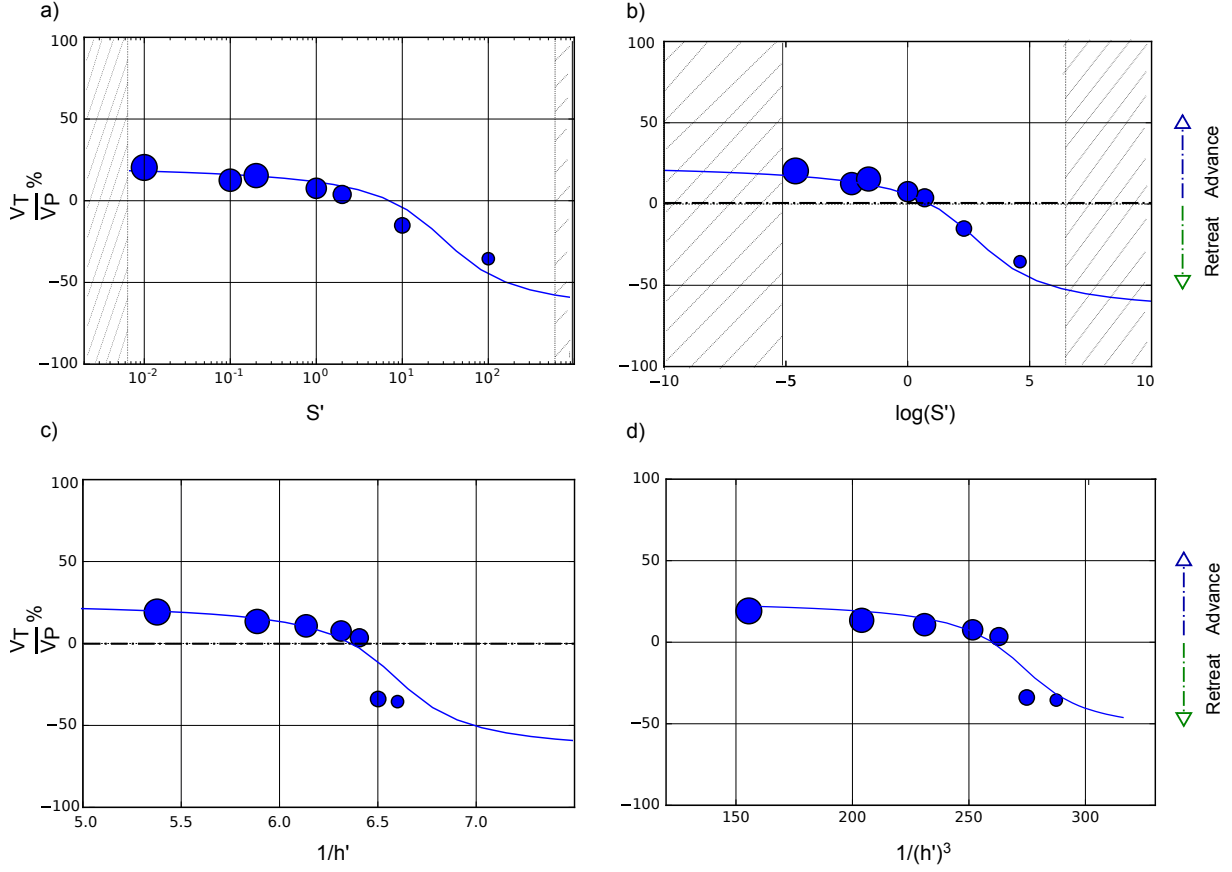


Figure 4.10: In a and b, the subduction partitioning (V_T/V_P %) as a function of the effective relative strength (S') and thickness (h') of the lithosphere for a progressively weaker or thinner WL in a) and c), respectively. In b) and d), the subduction partitioning (V_T/V_P %) as a function of $\log(S)$ and $1/(h')^3$ of the effective lithosphere for a progressively weaker or thinner WL, respectively. For the same set of experiments, the values are plotted using a scatter plot function of S' and h' in a) and c), and $\log(S)$ and $1/(h')^3$ in c) and d). The scatter plots size increases progressively with the viscosity and thickness of the WL.

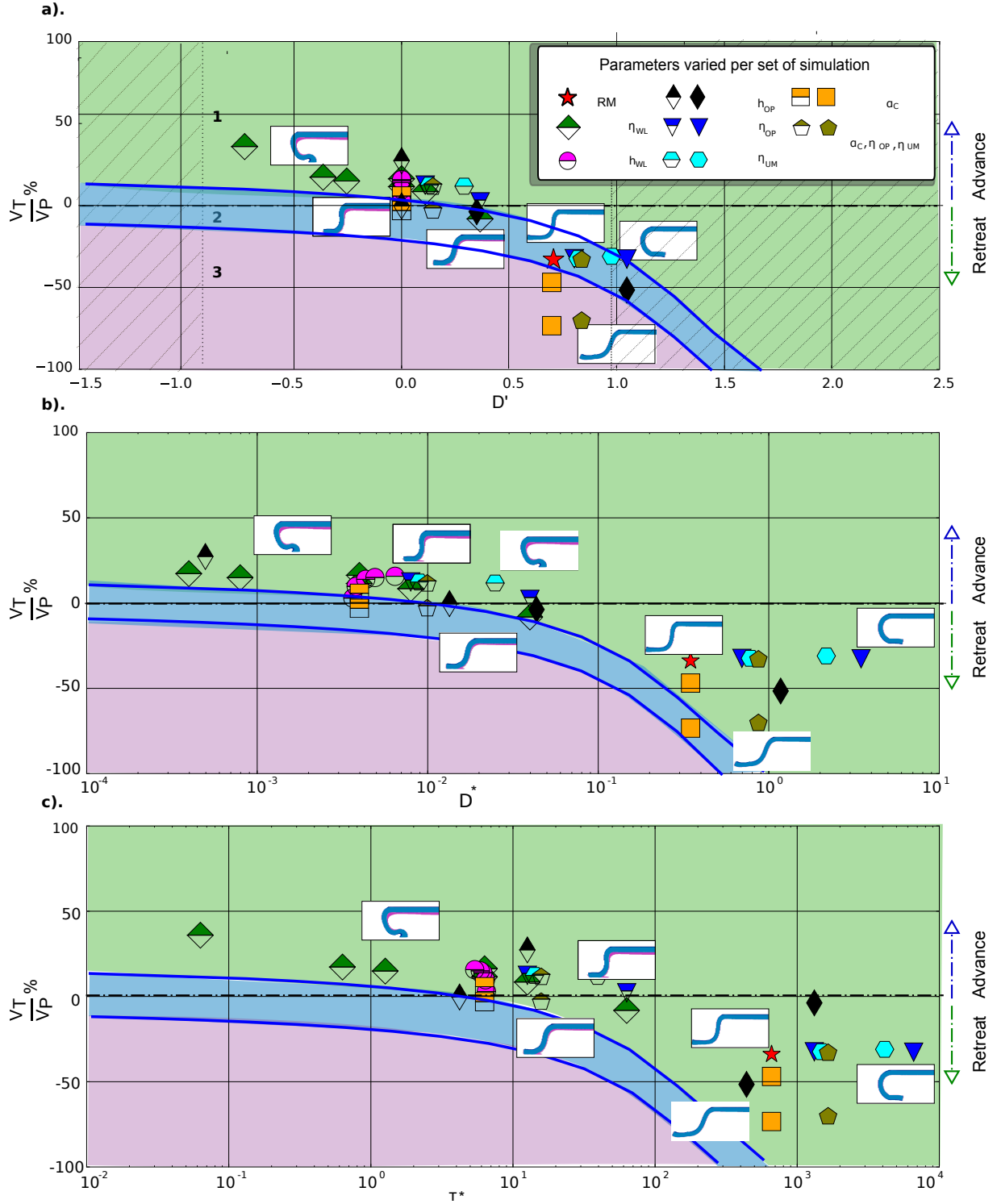


Figure 4.11: Subduction partitioning, $V_T/V_P\%$, with respect to the effective relative stiffness of the lithosphere D' , D^* and τ^* in a), b) and c), respectively for all the PP simulations performed in this study. In a) the different coloured fields define three areas having different subduction styles and surface motions. The different colour fields in b) and c) show our attempt to identify different subduction styles and motions based on other mechanisms, where a relationship does not emerge with a clear signature.)

4.10 Other unpublished work

Three different model setups were implemented to address some additional aspects of this chapter, such yield criteria, plate-mantle viscosity contrasts and buoyant overriding plate effects.

4.10.1 Yield criteria effects

The effect of implementing different yield criteria for the crust was explored because the weakening of the crust in the models is necessary to detach the slab from the surface and to allow the plate to move freely. Thus, in this subsection, the effects of different friction angles in a depth-dependent rheology are explored in a comparison between the RM and the uppermost asthenospheric weak layer (henceforth denoted as WL or SLL) having ($\eta_m/\eta_{WL} = 10e3$) case. These effects can be summarized as follow:

1. for low values of the friction angle ($\alpha < 10$), trench retreat in both the RM and the SLL cases; however the amount of trench retreat is smaller for the WL case. WL models additionally develop the highest velocities in comparison to the relative RM, consistent with the outcome presented in chapter 4, as shown in figure 4.10.1a);

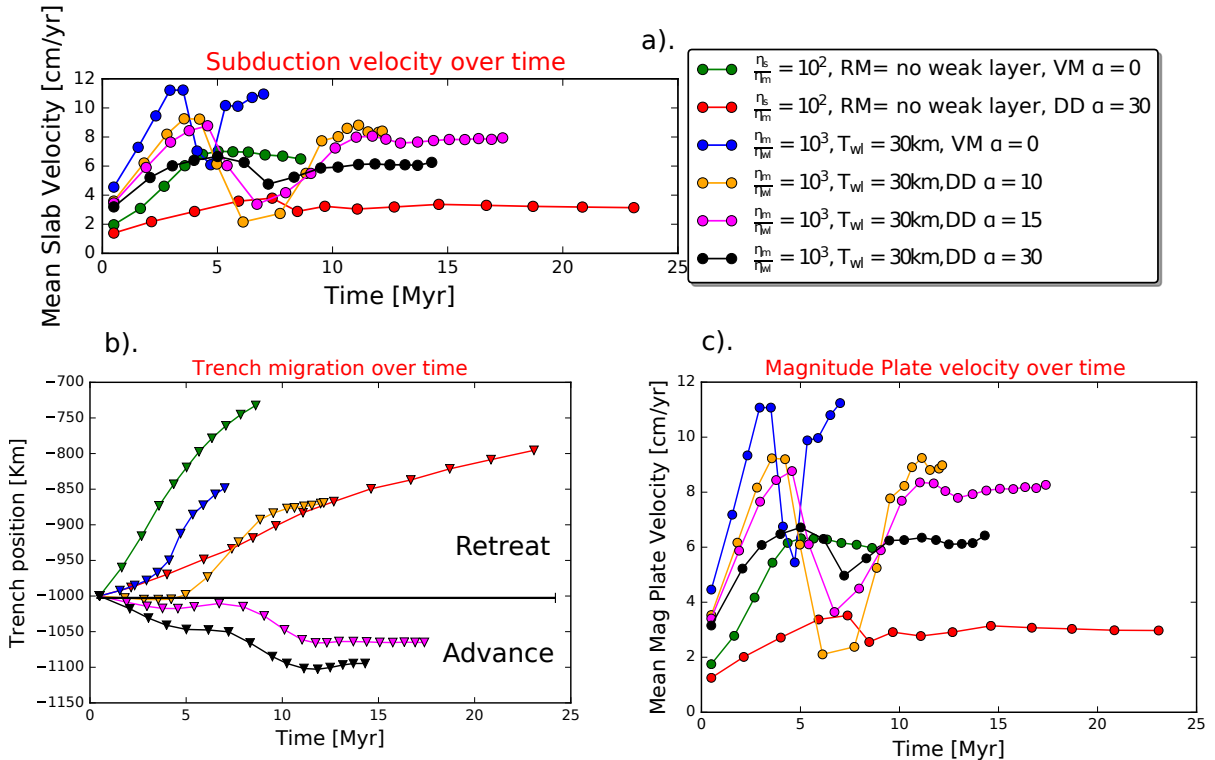


Figure 4.12: Effects of crustal friction angle on magnitude of main slab velocity, and trench and slab tip migrations in a), b) and c), respectively. These experiments adopt a Depth-Dependent Von-Mises criterion.

2. for moderate friction angles ($\alpha \sim 10$) an intermediate trench motion between advance and retreat is observed for the WL case (figure 4.10.1b). The trench is quasi stationary at the beginning of the evolution of the model, and retreats later after the slab tip has reached the bottom of the model. Also, slab velocities are intermediate in magnitude compared to the other end-member cases;
3. while for larger friction angles ($\alpha > 10$) overall slab and trench migration velocities are reduced, WL cases still present larger values of plate velocity and reduced trench migration compared to the RM case;
4. such as that, independently from the value assumed for α , the models with a WL show both higher subduction velocities and either a slow trench retreat or trench advance mechanisms, see also figure 4.10.1c) and b) respectively.

4.10.2 Subducting plate viscosity and thickness effects

To further prove the reliability of the model using previous works, the case of a stiffer plate (e.g. $\eta_s/\eta_m = 10e3$) developed by increasing the viscosity of the plate was also tested (figure 4.10.1). Hence, a stiffer plate generates a primary order well known effect on reducing trench migration and plate movements, a reverse slab morphology at the 660 km discontinuity (Bellahsen et al., 2005; Di Giuseppe et al., 2008; Funicello et al., 2008). All these effects were reproduced by our studies. However, when a WL is included in the model, it maintains its original signature by producing an increase in slab velocity, plate velocity V_P and affecting trench migration V_T (figure 4.10.1 a,b and c).

The oceanic plate thickness of the oceanic also affects subduction dynamics, as seen in previous work (e.g. Funicello et al., 2008). For our choice of parameters, thinner plates show lower subduction and trench retreat velocities, whilst thicker plate present larger velocities and retreating trenches (see figure 4.10.2a,b and c). Also during this set of experiments a WL has shown to keep a constant signature, as shown in figure 4.3d and subsection 4.6.2).

4.10.3 Effect of a buoyant overriding plate

The aim of using a more complex geometrical setup including an overriding plate was to test if a weak, buoyant channel at the LAB could still affect subduction dynamics in this case. In order to study this problem, a model setup slightly modified to the previous one was implemented as shown in figure 4.10.3. The new model setup differs from the previous one for the size and resolution of the modelled box, the length of the subducting plate (and weak layer) and presence of an overriding plate with Newtonian rheology. The modelled box now extends 7000 km in the X direction and 660 km in depth. The overriding plate and subducting plate (and weak layer) extend 2500 km and

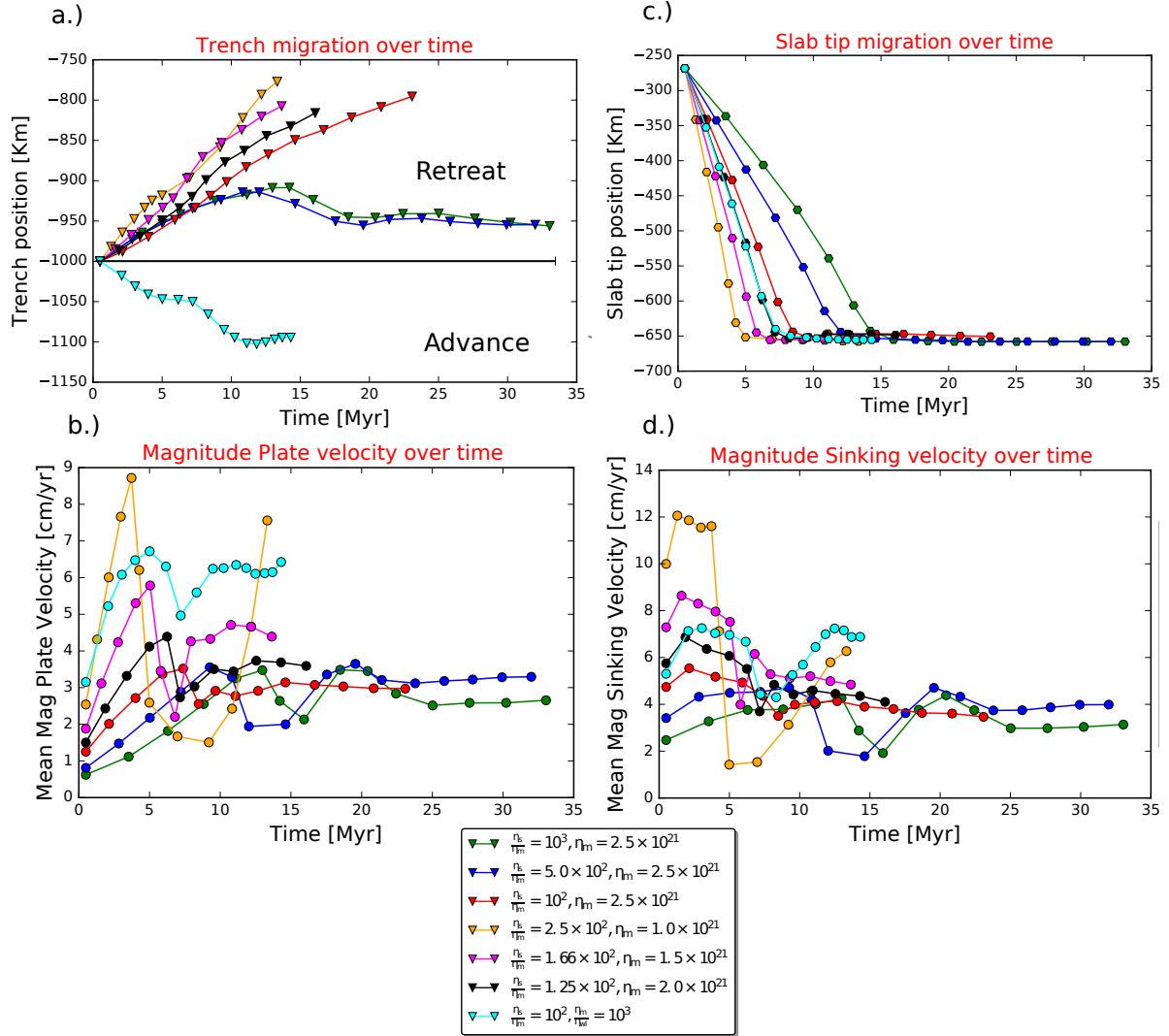


Figure 4.13: A comparison of the effects of lithosphere-mantle viscosity contrasts for different values of both plate and mantle viscosity. The effects of plate and mantle viscosity contrast are tested on slab velocity, trench and slab tip migrations in a), b) and c), respectively.

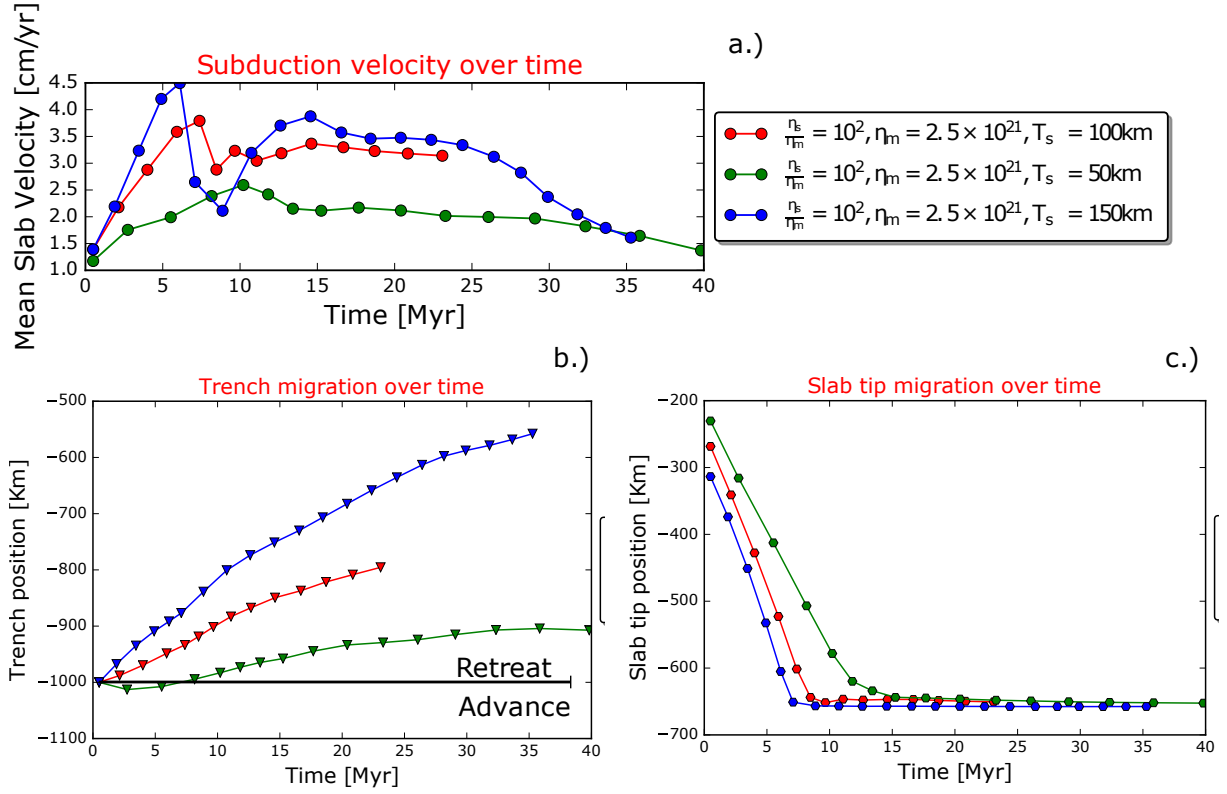


Figure 4.14: Effects of lithosphere thickness (with no weak layer) on slab velocity, trench and slab tip migrations in a), b) and c), respectively.

3250 km in the X-direction, respectively. The thickness of the continental crust and root are 32 and 100 km, respectively, the viscosity of the overriding plate is the same as for the subducting plate and all the other properties of the systems were left untouched. The modelled upper plate has no density contrast to the mantle, a thickness of 32 km crust and 100 km continental root and a constant viscosity of 2.5 Pa·s, thus reducing trench curvature to the minimum, similarly to the numerical approach of (Capitanio and Faccenda, 2012). The overriding plate is a passive agent in the models and no further boundary conditions are applied, subduction is thus dynamically initiated.

Models were made to determine the effects of the overriding plate on the morphology of the subducting slab and plate horizontal velocity. The temporal evolution of the mechanical models with the presence of an overriding plate for the RM and WL ($\eta_s/\eta_m = 10e2$) cases are shown in a comparison figure 4.10.3 a, b, d, f and h, i, j, m, respectively. While these models are slightly slower (less than 1 Myr) compared to the models without an overriding plate, the WL effect remains consistent. A WL promotes higher plate velocity and steeper slab morphology, as well as reduced rates of trench retreat (figure 4.10.3h, i, j, m and k, l, n).

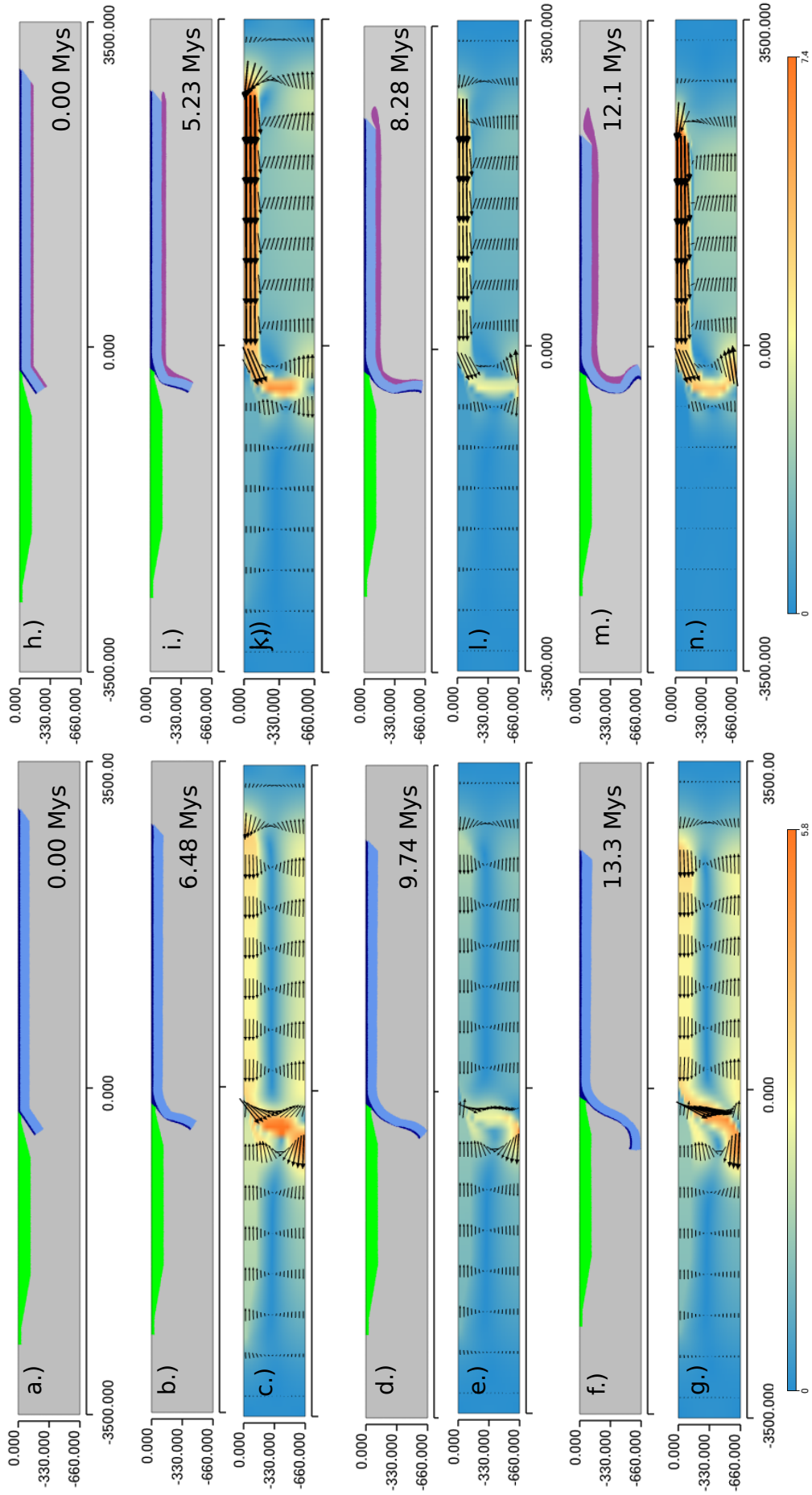


Figure 4.15: A comparison of the velocity field time evolution of mechanical models with an overriding plate for the RM and WL ($\frac{\eta_m}{\eta_{WL}} = 10^3$) cases with the addition of streamlines.

5. **Chapter 5: On the Influence of a Sub-Lithospheric Layer Width on the Partitioning of the Subduction Induced Mantle Flow**

On the influence of a sub-lithospheric layer width on the partitioning of the subduction induced Mantle Flow^{1 2}

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5.1 Highlights

- A range of geophysical observations suggests the presence of a very thin and weak sub-lithospheric layer (SLL) beneath the base of the subducting lithosphere, but little is known about its lateral extent.
- very little is also known about the three dimensional effects of an SLL on the partitioning of the deep and shallow deformation fields into its poloidal and toroidal components.
- Regional variations in SLL width provide a novel framework to interpret patterns of mantle circulation in natural subduction zones characterised by some degree of toroidal motion.

5.2 Abstract

The partitioning of the three-dimensional mantle flow into its toroidal and poloidal components is a diagnostic element to characterise present-day and past plate tectonics regimes. It provides a relevant driving force for the evolution of the geological mantle circulation and many geological processes. For example, the poloidal component is responsible for the sinking of the subducting lithosphere into the mantle and toroidal component allows for lateral transport of melt and volatiles throughout the subduction zone, as well as for the development of strike-slip motion. Hence, subduction-induced mantle flow plays a critical role in the evolution of convergent margins. Recent

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² *Keywords: subduction modelling, mantle flow, plate motions*

geophysical observations suggest the presence of a very weak and thin sub-lithospheric layer (SLL) beneath certain segments of the subducting Pacific plate; however very little is known about the lateral extent of this feature. To address the role of the SLL on mantle circulation, we perform buoyancy-driven numerical models where we systematically vary the width and the viscosity of the SLL. Our results show that a non-uniform SLL induces a consistent toroidal component underneath the base of the subducting plate even in models where the slab has no free edge, where the velocity field is otherwise poloidal. Our results present an average T/P average of 0.2-0.5 near the surface and 2.0 further at depth, which fit well with the expected values for a no-net reference frame. Our outcomes provide a novel and alternative explanation for the apparent lack of a relation between the slab pull driving force in buoyant convection and the generation of toroidal motion. Additionally, these findings support the interpretation of lateral slab dip, morphology and trench migration variations, as those observed along some of the subducting margins of the Pacific plate, where the presence of an SLL has been confirmed by geophysical studies.

5.3 Introduction

The lithosphere, the Earth’s uppermost surface layer, is fragmented into a series of strong major and minor tectonic plates moving over a weaker and generally more buoyant material known as the asthenosphere. At subduction zones, the negative buoyancy of slabs, resisted by the drag exerted from the surrounding viscous mantle, drives the motion and the deformation of the tectonic plates and mantle circulation (Jacoby, 1973; Forsyth et al., 1975; Lithgow-Bertelloni and Richards, 1995). The resistance to shearing acting at the lithosphere-asthenosphere boundary (LAB) plays a critical role influencing the force balance around the subducting plate and the degree of mantle-lithosphere coupling processes (Anderson, 1995; Natarov and Conrad, 2012). Beneath the base of the subducting lithosphere a very thin and weak sub-lithospheric layer (SLL) has shown to significantly reduce the mechanical coupling between plate motions and mantle tractions; hence impacting the partitioning of surface motions (Carluccio et al., 2019b) in the two dimensions (2D). Therefore, combined SLL viscosity and width anomalies could play a fundamental role in the three-dimensional (3D) force balance around a subduction zone; thus influencing how the deformation flow induced in the mantle by the sinking slab is split between the vertical and rotational components of the three-dimensional velocity field (Ribe, 2001; Höink et al., 2012; Billen and Arredondo, 2018; Richards and Lenardic, 2018). As such, the partitioning of induced mantle flow into its toroidal and poloidal components is critically important for our understanding of the mantle circulation around the trench zone (Faccenda and Capitanio, 2012; Long, 2013). It is also particularly relevant for back-arc basin formation, melt- and volatiles-transport in the backarc, as well as overriding

plate and subducting slab deformations (Sleep and Toksöz, 1971; Forsyth et al., 1975; Uyeda and Kanamori, 1979). Yet the impact of the SLL width on the partitioning of the subduction-induced mantle flow and how it relates to surface motions remains to be fully investigated.

Any 3D velocity field associated with plate motions can be divided into poloidal and toroidal components (Forsyth et al., 1975). Poloidal flow is associated with divergence in a horizontal plane (e.g., spreading centres and back-arc spreading at subduction zones). It corresponds to the vertical mass transport and thus promotes the deformation of the overriding plate (Sleep and Toksöz, 1971; Uyeda and Kanamori, 1979; Dvorkin et al., 1993; Funiciello et al., 2003; Schellart and Moresi, 2013). Toroidal flow is associated with vertical vorticity (rotation in the horizontal plane around a vertical axis) and generates transform (i.e., strike-slip) motion or plate rotation (Tackley, 2000). The partitioning of the induced mantle flow is henceforth denoted as the amplitude ratio between the toroidal and poloidal (T/P) components of the Earth’s deformation field. The present-day decomposition of the velocity field shows as much poloidal as toroidal kinetic energy comparable at most wavelengths (e.g., Hager and O’Connell, 1978; Forte and Peltier, 1987; O’Connell et al., 1991; Lithgow-Bertelloni et al., 1993; Dumoulin et al., 1998). While, the present-day average toroidal motion depends on the choice of reference frame adopted and changes during geologic time (Čadek and Ricard, 1992; Lithgow-Bertelloni et al., 1993; Tackley, 2000), no plate boundary is purely poloidal or toroidal but usually displays some combination of the two. Both flow fields are currently observed on Earth and strongly localized along plate boundaries with comparable peak values (O’Connell et al., 1991; Bercovici, 2003). The superficial toroidal component is a relevant fraction of the poloidal flow (Hager and O’Connell, 1978), although smaller, with a T/P having ranged from approximately 0.25 and 0.5 (excluding net rotation) in the last 120 million years (Olson and Bercovici, 1991; Lithgow-Bertelloni et al., 1993).

Buoyancy driven fluids with only vertically layered viscosity do not generate any toroidal motion and lateral variable viscosity convection (e. g., with constant, depth-dependent or temperature-dependent viscosity) calculations typically produce only a limited amount of T/P (Christensen and Harder, 1991; Tackley, 2000). There is apparently no direct link between the driving slab pull force in buoyant convection (which drives only vertical and divergent motion) and the development of toroidal energy. The challenge of how to trigger buoyancy-driven poloidal motion to induce and self-sustained toroidal flow, either through horizontal viscosity variations or through boundary conditions (i.e., imposing plates, faults or strain weakening effects at plate margins) represents the so-called ”poloidal–toroidal coupling problem” and is one of the most critical issues in understanding the generation of Earth’s plate-tectonic style of mantle convection (Bercovici et al., 2015).

It has been proposed (e.g., Gable et al., 1991) that the degree of coupling between the lithosphere-mantle could contribute to the large toroidal component that we currently observe at natural tectonic boundaries (O’Connell et al., 1991; Bercovici and Wessel, 1994). Global convection models show that the inclusion of a thickened low viscosity zone (LVZ) - the asthenosphere - favours plate-like behaviour; however, this also seems to have a smaller effect on the global net rotation and observed patterns of global seismic anisotropy (Tackley, 2000; McNamara et al., 2010). On the other hand, a very weak and thin sub-lithospheric layer has shown the potential to substantially accommodate large amounts of strain drop through horizontal mechanical shearing and gravitational sliding directly underneath the base of the lithosphere (Scoppola et al., 2006; Carluccio et al., 2019b). Thus, variations in the lateral properties of the SLL are expected to result in differential flow at the boundary between tectonic plates and underlying mantle and could play an important role in the partitioning of the induced mantle flow and subduction surface motions.

The subduction of the oceanic lithosphere into the mantle is accompanied by trench surface motion. Trenches are mobile with respect to the plate and other plate boundaries and can either retreat, advance or have a quasi-stationary motion in time relative to a specific reference frame. We refer to the partitioning of subduction surface motions, or subduction partitioning, as the ratio between the horizontal plate (V_P) and trench migration (V_T) velocities, (V_T/V_P). Most of the Earth slab and trench is characterised by roll-back and retreat ($-V_T$); however, advancing plate and trench motions ($+V_T$) towards the upper plate can also be found in some trench zones. Furthermore, the average absolute subduction partitioning observed globally is one third or less of the total convergence at the present, consistent across the majority of plate motion reference frames (Elsasser, 1971; Jarrard, 1986; Schellart et al., 2008; Becker and Faccenna, 2009; Goes et al., 2011; Holt and Becker, 2016). While many previous models have provided useful insights into the modelling of natural subduction systems and mantle convection processes (Garfunkel et al., 1986; Jarrard, 1986; King, 2001; Heuret and Lallemand, 2005; Lallemand et al., 2005; Sdrolias and Müller, 2006; Capitanio et al., 2007), the modelling of subduction zones in the case of slow retreating or advancing trenches in natural settings has only recently started to resolve a clear trend. Some of these works have shown the effect of slab and mantle rheology in reducing trench retreat (Holt and Becker, 2016; Billen and Arredondo, 2018), ridge push condition in enhancing plate motion (Capitanio et al., 2007), as well as an SLL in decreasing the high absolute values of subduction partitioning (Carluccio et al., 2019b). These, along with the addition of an overriding plate to the system (Holt et al., 2015b), seem to be relevant factors for reducing $|V_T/V_P|$ to a level more in line with those currently observed on Earth (Coltice et al., 2017). Yet, the relationship between V_T/V_P and induced mantle flow, T/P , and the subduction style due to an SLL width remains to be elucidated.

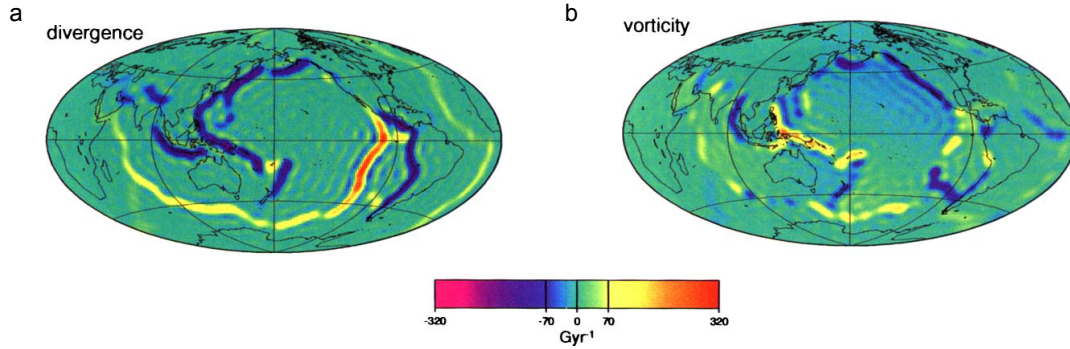


Figure 5.1: Horizontal divergence (left) and radial vorticity (right) fields for the discontinuous plate-tectonic model. This model was derived from a spherical-harmonic analysis of O’Connell et al. (1991) and edited later by Bercovici (2003).

Previous 3D subduction modelling work has shown how the subduction of an oceanic plate into the underlying mantle is accommodated by both the poloidal and the toroidal components of the velocity field (e.g., Becker et al., 1999; Conrad and Hager, 1999; Piro-mallo et al., 2006; Kneller and Van Keken, 2008; Király et al., 2016). The extent of this contribution has been shown to depend upon lateral variations in plate and trench widths (Funiciello et al., 2004; Stegman et al., 2006; Billen and Hirth, 2007; Schellart et al., 2007), the distribution of temperature around the slab, age and strength of a plate (Kincaid and Griffiths, 2003, 2004; Morra et al., 2006), viscosity and buoyancy contrasts between slab and mantle (Becker et al., 1999; Conrad and Hager, 1999; Piro-mallo et al., 2006; Capitanio et al., 2007; Kneller and Van Keken, 2008; Király et al., 2016), lithosphere-mantle rheology (Billen and Hirth, 2005; Pusok et al., 2018), spherical spontaneously driven subduction initial conditions (Cramer and Tackley, 2014), inclusion of more buoyant surface material (Moresi et al., 2014; Pusok and Kaus, 2015), slabs interaction with each other (Király et al., 2017) and with the overriding plate (Sternai et al., 2014), double subduction systems (Pusok and Stegman, 2019) as well as plate geometry (Bercovici and Wessel, 1994; Kneller and Van Keken, 2008).

These factors have shown to influence the mantle displacement around a slab free edge which substantially contribute to the curvature and migration of a subduction zone (e.g., Funiciello et al., 2008; Faccenna et al., 2010; Guillaume et al., 2010; Jadamec and Billen, 2010; Long, 2013). While each ingredient of the subduction system is fundamental and contributes to the evolution of mantle flow and slabs deformation, a clear pattern in the relative control of the partitioning of mantle circulation and its relationship with subduction surface motions on subduction dynamics has to be resolved. A very thin and weak SLL of variable width could exert an important control on the partition of the induced pattern of mantle flow, significantly influencing the amplitude T/P, as well as slab and trench deformation styles. This is especially salient in light of recent geophysical

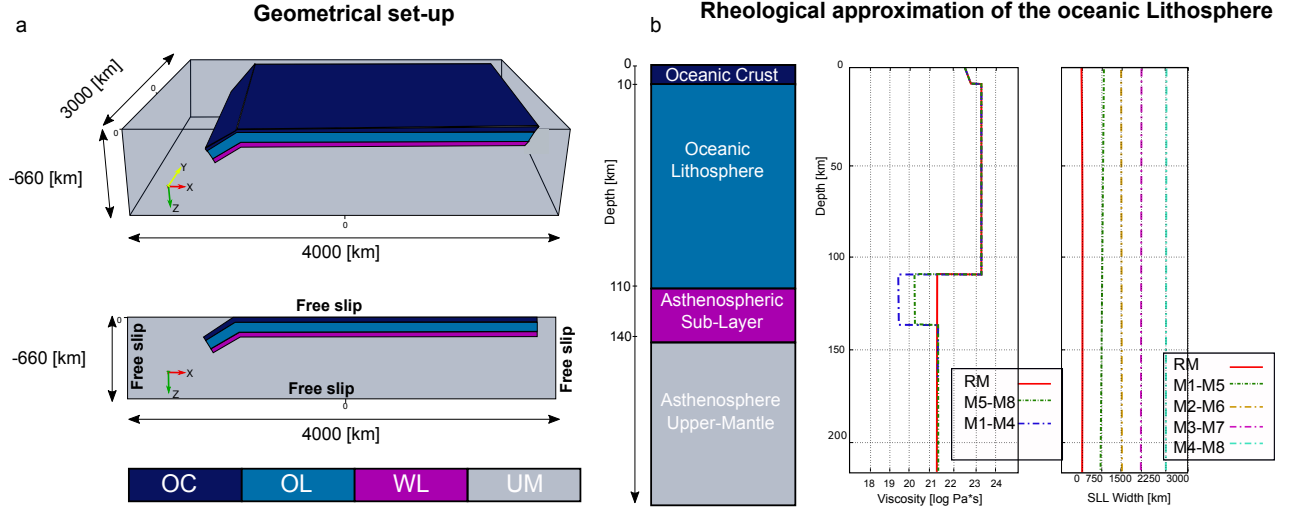


Figure 5.2: In a, simplified representation of the geometrical setup for the reference model with a sub-lithospheric layer (SLL). In b, material field profile through depth, including the initial rheological setup for different viscosity contrasts mantle-SLL (10^0 , 10^1) and 10^2) and SLL-oceanic lithosphere width ratio (1, 1/4, 1/2, 3/4 and 0) of the models presented in this paper.

observations having brought considerable attention to the structure of the asthenosphere at convergent margins (Kawakatsu et al., 2009; Schmerr, 2012; Naif et al., 2013; Stern et al., 2015; Hawley et al., 2016). These report regionals abrupt seismic velocity decrease and high electrical conductivity increase within a 30 km thick channel underneath several segments of the subducting Pacific plate. Melt-induced viscosity reduction into one or multiple thin and more buoyant layers are thought to lead up to a ≈ 4 orders of magnitude viscosity jump between the lithosphere and underlying uppermost asthenosphere. While the enhanced melt content and/or presence of hydrated phases could have significantly lubricated the base of the subducting lithosphere with potential implications for global subduction dynamics, the complicated physical nature of the LAB and the trade-off that exists between power and lateral coverage of seismic methods have hindered the ability to image the full extent of these features. Since little is known about the lateral extent of an SLL, here we use 3D numerical models to address the influence of SLL width on induced patterns of deep and surface mantle circulation and explore the relationship between them.

We build upon the work of Carluccio et al. (2019b) by varying the viscosity and width of the SLL to determine the relative controls on slab and trench curvature and shape, and partitioning of mantle flow.

5.4 Method

To investigate the role of an SLL on subduction dynamics, we use the numerical code Underworld2, described in detail by Moresi et al. (2007) and Mansour et al. (2020). Un-

derworld follows a continuum mechanics approach that solves the conservation equations of mass, momentum and energy. We consider subduction and plate motions in an incompressible Boussinesq fluid at infinite Prandtl number and very low Reynolds number. We then solve the Stokes equation that balances buoyancy forces with viscous stresses and pressure gradients. Underworld uses a particle-in-cell finite element method in which material properties are carried by Lagrangian integration points and mapped to element properties through a local discrete-Voronoi integration scheme. This numerical approach is designed for accurately tracking distinct material domains in the lithosphere as they undergo large deformation due to creeping flow. Here, we neglect temperature diffusion and solve the conservation equations for mass and momentum together with their constitutive relationships. The non-dimensional equations of motion for an incompressible fluid are:

$$\frac{\partial v_i}{\partial x_i} = 0 \quad (5.1)$$

and

$$\frac{\partial \tau_{ij}}{\partial x_j} - \frac{\partial p}{\partial x_i} + \rho g_i = 0 \quad (5.2)$$

Where i and j represent spatial directions and follow the Einstein summation convention, $\tau_{ij} = \sigma_{ij} + p\delta_{ij}$ is the Cauchy stress deviator tensor, ρ is the fluid density, v_i is the local velocity vector of a material, g_i is the gravitational acceleration and $\{x_i\}$ is the spatial coordinate system. Further details describing the non-dimensional equations and the numerical technique in underworld can be found in Moresi et al. (2003) and Stegman et al. (2006)

5.4.1 Rheological model

We adopt a pseudo-plastic rheology approximation (PP) described in detail in Carluccio et al. (2019b) and shown in figure 5.2c. This approach uses Newtonian rheology everywhere but the uppermost layer of the subducting plate (representing the oceanic crust and any associated sedimentary layers) which is visco-plastic. The visco-plastic layer acts as a lubricating phase to simulate the free surface and to allow the hinge of the bending plate to detach from the surface (Schmeling et al., 2008; Cramer et al., 2012). When a Newtonian constitutive relationship is implemented (for an incompressible fluid) the relationship between the deviatoric stress, τ_{ij} , and strain rate, η^{vs} , tensors is defined as:

$$\tau_{ij} = 2\eta^{vs}\dot{\epsilon}_{ij} \quad (5.3)$$

where $\dot{\epsilon}_{ij} = \frac{1}{2} \left(\frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i} \right)$ is the strain rate tensor.

When plastic failure is considered for the uppermost layer, the total effective viscosity, η_{eff} , is taken as the minimum between the viscous, η^{vs} , and the plastic, η^{pl} , terms, as follow:

$$\eta_{\text{eff}} = \min[\eta^{\text{vs}}, \eta^{\text{pl}}] \quad (5.4)$$

We assume Byerlee’s law which describes the increase in the maximum strength of the upper crustal rocks with increasing confining pressure (depth). This is implemented by assuming the so-called depth-dependent approximation of the Drucker-Prager yield criterion (see Moresi and Solomatov (1998); Tackley (2000); Stein et al. (2004) for details) which introduces a friction angle and depends only on the lithostatic pressure and neglects the dynamic component. In this case, the yield stress, τ_y , and the effective viscosity of the plastic branch, η_{pl} , are implemented as follow:

$$\tau_y = \tau_0 + \mu p_{\text{lith}} \quad (5.5a) \quad \eta_{\text{pl}} = \frac{\tau_y}{2\dot{\epsilon}_{II}} \quad (5.5b)$$

where $\dot{\epsilon}_{II} = \left(\frac{1}{2}\dot{\epsilon}_{ij}\dot{\epsilon}_{ij}\right)^{(1/2)}$ is the second invariant of the strain rate, τ_0 the cohesive strength, μ the friction coefficient and $p_{\text{lith}} = \rho g z$ the reference lithostatic pressure at depth z . We use a cohesion of a 20 MPa, a friction angle of 30 degrees and a density contrast between the oceanic lithosphere and the underlying mantle of 80 kg/m^3 , in line with the values calculated in the detailed study of Cloos (1993).

5.4.2 Geometry of the model setup

We model subduction dynamics and mantle flow in a Cartesian box that extends 4000 km and 3000 km in the horizontal X and Y directions, respectively, and 660 km in depth (Z). Our models are made of 4 layers. There is an oceanic crust (OC), an oceanic plate (OP), a weak SLL and an underlying upper mantle (UM), as shown in figure 5.2b. This setup builds upon Carluccio et al. (2019b) to expand to the third dimension. Although similar, this setup differs from previous 3D studies (Stegman et al., 2006; Schellart et al., 2007) because the oceanic plate is made of one single layer and the SLL is characterized by a lower viscosity value than in the underlying asthenosphere or deeper mantle. The initial condition for the subducting plate includes an initial trench position 1000 km away from the wall, an initial plate width equal to the box width, and a slab tip already penetrating the mantle, 257 km deep with a dip angle of 34° to the horizontal. The modelled SLL is located at the base of the oceanic plate and has a variable width compared to the whole length of the oceanic plate (figure 5.2c and table 5.4.3), and a default thickness of 30 km. We use a model domain of $256 \times 128 \times 128$ elements with a uniform grid spacing in each coordinate direction which gives a grid resolution of 5.16 km (Z) and 23.4 km (Y) and 15.6 km (X). This ensures sufficient numerical resolution to promote and properly resolve

the development of a weak hinge zone. We verified the model outputs using a lower resolution grid of $256 \times 64 \times 64$ for a crustal layer and an SLL two times thicker. Note that the number of elements is constrained by our use of a simple geometrical-multigrid solver which dictates the resolution in each coordinate direction to be $N = m \times 2^n$ elements, where m is a small integer (typically 2, 3, 5) and n is the number of levels in the multigrid mesh hierarchy. The velocity boundary conditions are free-slip everywhere to minimize the influence of the box sidewalls (a free-slip condition means no tangential stress and zero normal velocity). We do not apply any kinematic boundary condition and subduction is thus dynamically initiated. The reliability of the model has been tested against previous 3D work as shown in section 5.5. The assigned values result in subduction velocities of 2-10 cm/yr (figure 5.4a, 5.5), and divergence and vorticity ± 200 Gyr (figure 5.6, 5.7, 5.9) both in line with the average values predicted on Earth O’Connell et al. (1991); Tackley (2000); Bercovici (2003); Stegman et al. (2010) (figure 5.2).

5.4.3 Subduction surface motions

The pattern of flow in the asthenosphere can promote or resist plate motions with respect to the underlying mantle. This flow can be driven by the plates themselves (Couette flow), by pressure gradients induced by global flow within the mantle (Poiseuille flow) or a combination of the two (combined Couette-Poiseuille flow). Several studies in the past (Conrad and Hager, 1999; Ribe, 2001) and more recently (Stegman et al., 2010; Richards and Lenardic, 2018; Carluccio et al., 2019b) outline the importance of the balance of bending moment and internal stresses for the description of the mechanical strength of the lithosphere and the asthenosphere.

To explore the relationships between the effective properties of the lithosphere (OP+SLL) relative to the underlying mantle, which may or may not contain an SLL, and surface motion we define a series of parameters related to the effective relative strength (S') and width (W') of the lithosphere. These quantities build upon those identified by Carluccio et al. (2019b), hence representing an empirical first-order approximation of the deformation occurring in the 3D subduction system in terms of dimensionless ratios among the physical properties of the materials modelled, as:

$$S' = \log \left(\frac{\eta_P \eta_{SLL}}{\eta_M^2} \right), \quad w' = \frac{w_P}{w_M} \left(1 + \frac{w_{SLL}}{w_M} \right) \quad (5.6)$$

where η_P , η_{SLL} and η_M indicate the viscosity of the plate, weak layer and mantle respectively, and w_P , w_{SLL} and w_M being widths of the plate, the SLL and the mantle respectively. Note that, when an SLL is not included, $\eta_{SLL} = \eta_M$ and $w_{SLL} = w_M$, which allows us to characterise all the experiments both with and without an SLL.

Model parameters varied per set of simulation in this study					
# Set	Model ID	W_{OP}/W_{box}	W_{SLL}/W_{OP}	η_M/η_{SLL}	$h_{SLL}[\text{km}]$
1	RM	1	0	10^0	0
	M1-M4	1	1/4, 1/2, 3/4,1	10^2	30
	M5-M9	1	1/4, 1/2, 3/4,1	10^1	0,30
	M10	1/2	0	10^0	0
	M11-M14	1/2	1/4, 1/2, 3/4,1	$10^0 - 10^3$	0,30
	M15-M19	1/2	1/4, 1/2, 3/4,1	$10^0 - 10^3$	0,30

Table 5.1: Parameters used and varied for the experiments performed during this study.

5.4.4 Induced flow characteristics

We decompose the 3D subduction deformation field into its poloidal and toroidal components to analyse the induced flow characteristics. The 3D decomposition is evaluated in a two-dimensional slice by evaluating the velocity gradients for each 2D grid point at the desired time and space in the evolution of the subducting slab. The surface integration of the poloidal and toroidal components yields to their 2D scalars, Φ and Ψ , respectively, consistent with previous work (Tackley, 2000; Piromallo et al., 2006; Stegman et al., 2006).

The horizontal divergence as:

$$\nabla_h = \left(\frac{\partial}{\partial x}, \frac{\partial}{\partial y} \right) \cdot \mathbf{u} \quad (5.7)$$

and the vertical component of vorticity as:

$$w_z = \hat{z} \cdot \nabla \times \mathbf{u} \quad (5.8)$$

In very symmetric cases it is possible to solve the Poisson's equation for the poloidal and toroidal potentials by integrating over a 2D surface. Such that, we obtain the poloidal velocity scalar Φ by integrating ∇_h over a surface of interest:

$$\Phi(t)|_{y=width} = \oint\!\!\!\oint_{H(W)} \nabla_h(t) \, dx \, dz \quad (5.9)$$

and the toroidal velocity scalar Ψ as:

$$\Psi(t)|_{z=depth} = \oint\!\!\!\oint_{W(L)} w_z(t) \, dx \, dy \quad (5.10)$$

One of the most traditional quantities invoked to quantify plate-like behaviour and long-term dynamics of the convecting system is the toroidal:poloidal (T/P) ratio (Tackley, 2000), which is simply:

$$T/P = \frac{\Psi}{\Phi} \quad (5.11)$$

The decomposition of the velocity field is performed on individual lateral or vertical planes. In the interest of this study we analyse mantle flow along phase boundaries through three individual horizontal planes (XY) of constant depths, which allows us to present the lateral variability we obtain in our system. These slices are located close to the surface (40 km), underneath the LAB (150 km) and in the vicinity of the 660 km discontinuity (450 km). The evaluation of the velocity field is carried out at the least 3 elements away from a phase boundary to avoid numerical noise given by the abrupt change in physical properties within two adjacent layers. We also explore the effect of SLL width on the vertical mass transport evaluating the poloidal component in three vertical planes located at the front, mid and back vertical XZ surfaces (figure 4.4 in supporting information).

By varying SLL width compared to Op width between 1 ($W_{SLL}=W_{OP}$) and 0 ($W_{SLL}=0$), we aim to identify end-member cases that outline the mechanical effect of variations in SLL width underneath a descending plate on slab and mantle dynamics. For this goal, we use a simplified rheological and geometrical model. This setup allows us to constrain only the impact of an SLL width and to show how the wide range of subduction behaviours and the patterns of mantle circulation can be the result of a simple force balance around subducting plates, while no external forces are needed. Our single slab model setup with no free-edge slab allows us to apply our decomposition to models which have no-lateral viscosity variations, such as the reference model, RM ($W_{SLL}/W_{OL} = 0$) and M4 ($W_{SLL}/W_{OL} = 1$); hence ideally having zero toroidal flow. As it would be expected, these modelling cases yield to very small T/P ratios (of order 10^{-4} – 10^{-5}) (figure 5.8 and 5.9), consistent with the accuracy of single-precision arithmetic, indicating that significant (e.g., of order 10^{-2} – 10^{-1}) toroidal flow, when found, is not an artefact of numerical leakage. The setup chosen, thus, allows us to verify the correctness of the decomposition of the velocity fields into its poloidal and toroidal components, surface motions and other plate-behaviours diagnostics.

In our study, we vary the properties of the SLL at the base of the oceanic lithosphere, i.e., its viscosity and width. A small number of more complicated models explores the extent to which these simplifications may influence our conclusions, such as the use of a single slab model setup with a free-edge slab and the inclusion of the lower mantle. The results obtained in our numerical simulations are systematically analysed through the analysis of the impact of SLL width on i) slab morphology and subduction surface motion

partitioning; ii) the evolution through time of induced poloidal and toroidal potentials and their partitioning, in addition to the investigation of the divergence and vorticity fields; and ultimately iii) the applicability of our numerical outcomes to natural subduction zones. More information about the parameters used and varied in this study is available in table 5.4.3 and figure 5.2, whereas more details about the methodology adopted can be found in Carluccio et al. (2019b).

5.5 Results

We start by presenting the general aspects of the subduction evolution of the reference model (RM), which satisfy a series of natural observations, and later outline the outcomes of the remaining models in relation to the RM case and, subsequently, to one another. All our numerical experiments share in common the three classical and distinct stages of a slab evolution (figure 5.4), also observed in previous studies (Bellahsen et al., 2005; Stegman et al., 2006; Yamato et al., 2009). These stages are: 1) a starting period, in which the perturbation provided by the initial model geometry is sufficient to dynamically initiate and self-sustain subduction and substantial mantle flow around the hinge and slab tip. During this period the slab accelerates while sinking into the underlying mantle. This stage is accompanied by the development and progressive increase in the intensity of the two poloidal cells located at both sides of the subducting slab; 2) a transitional phase where the slab decelerates due to the interaction with the bottom boundary of the model and, consequently, the size of the poloidal cells increases and their intensity decreases; and 3) a final stage in which the system reaches a steady-state configuration between the previous two, whilst the subducted slab approaches a resting position atop the bottom boundary of the model.

5.5.1 Reference model time evolution

The reference model has the same initial geometry as the model setup, but it does not contain an SLL (figure 5.2a). As the evolution through time of our numerical experiments follows the aforementioned classical description (§5.5), we continue by presenting the main key features of the RM and other models that include an SLL during the final temporal stage when the system settles into an approximate steady-state configuration.

The overall evolution of the RM exhibits trench retreat and slab roll-back mechanisms accompanied by the presence of two poloidal cells one ahead and one behind the subducting slab, present from the earliest point in the model evolution (figure 5.3-5.4). The slab shows an average dip-angle of 60-70°, which slightly increases when the slab tip interacts with the bottom of the modelled box in ~ 7.8 million years, after which a steady-state stage is reached. The final stage of the subduction evolution of the RM is defined by continuous trench retreat and slab roll-back motion, whilst the slab lies flat at the 660

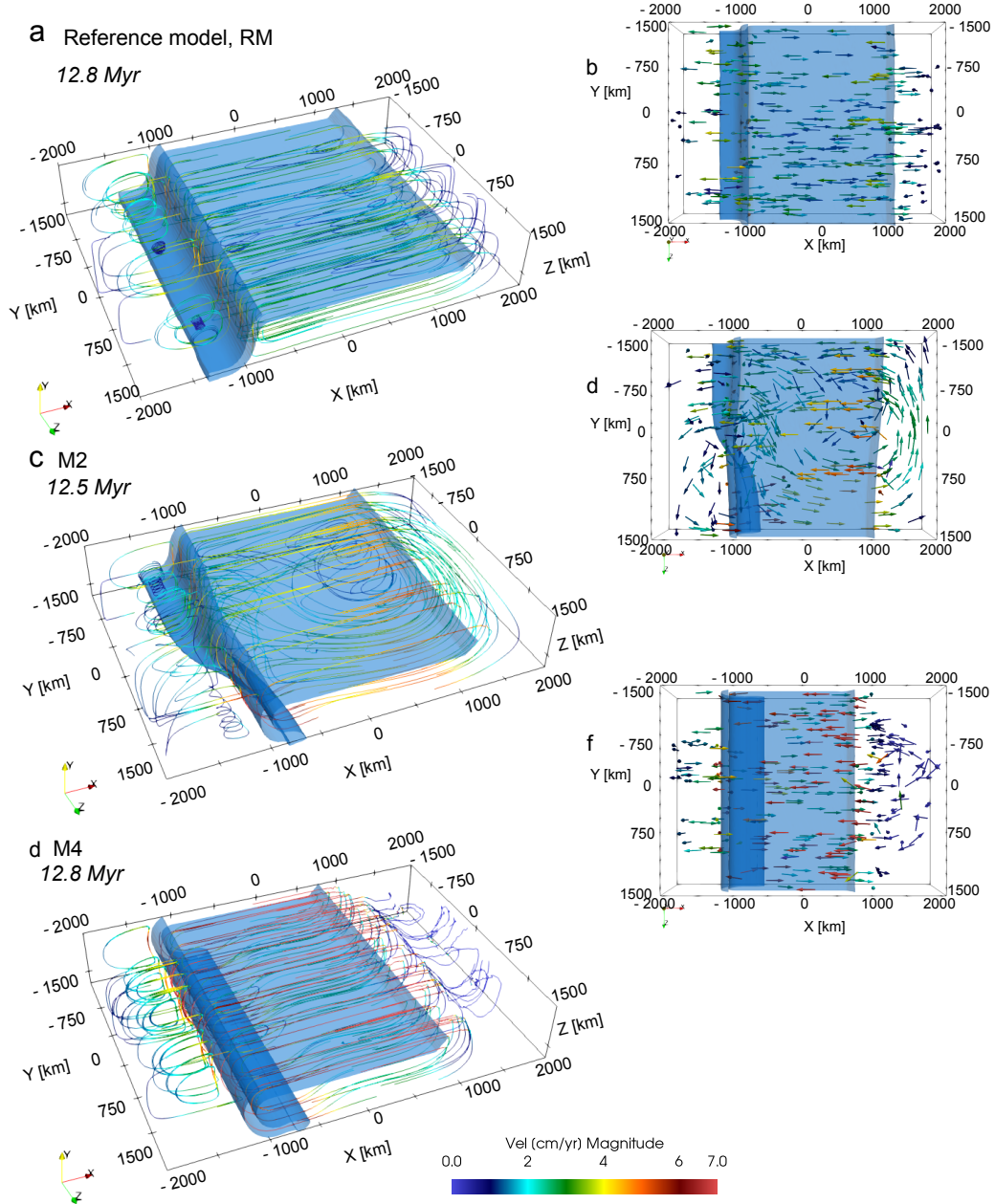


Figure 5.3: Subduction style comparison of (top to bottom) the RM, M2 ($\eta_M/\eta_{SLL} = 100$) of $W_{SLL} = 1/2W_{OP}$ and M4 ($\eta_M/\eta_{SLL} = 100$) with $W_{SLL} = W_{OP}$ at the steady-state stage of the time evolution. In a and b, the RM case shows slab and the trench rollback and retreat, respectively. In c and d, M2 exhibits asymmetric patterns of trench (retreat and advance) and slab (fat and buckled) migrations. In e and f, M4 develops trench advance and a buckled slab at the 660 km discontinuity.

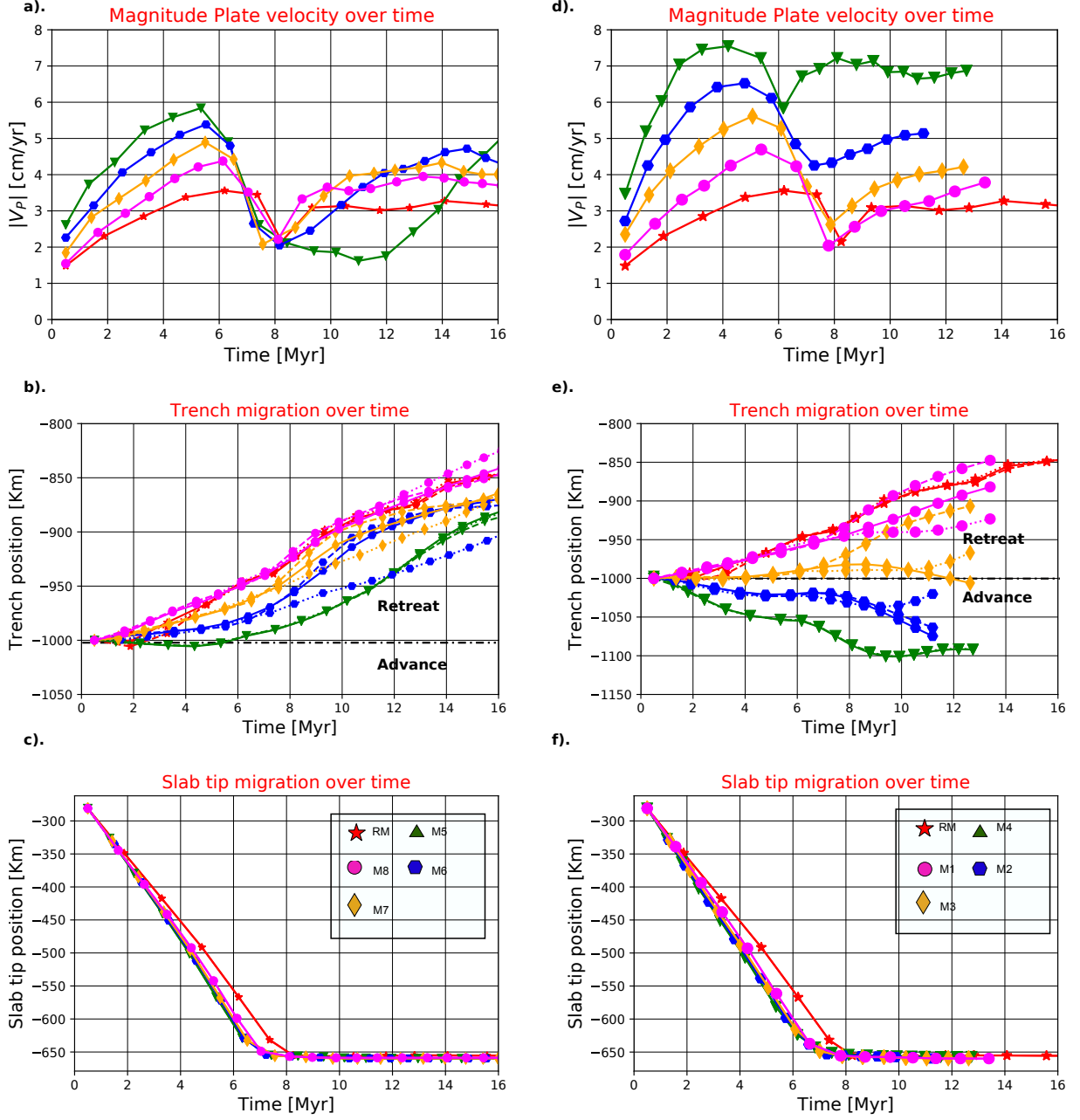


Figure 5.4: Evolution of plate velocities (V_P), and trench and slab tip migrations through time, for a progressive increase in SLL width, for $\eta_M/\eta_{SLL} = 10$ in a, b, c and $\eta_M/\eta_{SLL} = 10^2$ in d, e, f, respectively.

km discontinuity. The vertical vorticity and, thus the toroidal component throughout this model are both negligible, $|w_z| < 5$ and $\Psi \approx 10^{-3} - 10^{-4}$, respectively, regardless of depth (figure 5.8a and 5.9). While some numerical noise in relation to areas of high strain rates (i.e., at ridges and above the subducting slabs) can be observed, our RM provides a reliable base indicating that when significant toroidal flow component is found in other models, it is not an artefact or numerical leakage. As it would be expected for a model that does not assume lateral viscosity or temperature field variations, in agreement with previous work (Bellahsen et al., 2005; Stegman et al., 2006; Funicello et al., 2008), thus it was chosen as a reference for comparison with models having an SLL.

5.5.2 Effect of an SLL width on subducting slabs morphology and surface motions

We describe the combined impact of SLL width and viscosity on 3D subduction dynamics, in addition to the overall effect an SLL exerts on subduction surface motions to promote plate convergence velocity, hamper trench retreat motion and increase the subduction dip angle (figure 5.3C and 5.4c,f), consistent with the 2D numerical study Carluccio et al. (2019b).

The primary effect of changing SLL width is to produce a variation in the slab subducting speed up to two and a half times faster than the RM and to promote the development of substantial toroidal flow. The development of toroidal flow unevenly impacts the 3D slab structure and trench migration rates unevenly (figure 5.3b, 5.4 and 5.5). Conversely, this has only a limited effect on the time necessary for the slab tip to reach the base of the modelled box (figure 5.4c and f), and hence on the sinking subduction motion.

We divided our numerical experiments in two classes with two different viscosity contrasts between the SLL and mantle. The first case has low $\eta_M/\eta_{SLL} = 10^1$ (models M5-M8) and the second case has moderate $\eta_M/\eta_{SLL} \geq 10^2$ (models M1-M4) and we vary the ratio between SLL and OP width, W_{SLL}/W_{OP} , from 0 to 1 with an increment of 1/4 (figure 5.2b).

The models within class 1 show plate velocity that may increase to a maximum factor of ~ 1.7 compared to RM. Slab roll back and the trench has either a quasi-stationary type of motion, in M8 ($W_{SLL}/W_{OP} = 1$), or reduced trench retreat rates compared the RM. Models with larger SLL width overall exhibit steeper slabs and trench migrations variation ($\Delta X t$), across the 3D slab, up to a maximum of $\Delta X t_{z_{surface}} = 70$ km (figure 5.4b).

The second class of models and, so for moderate to high viscosity contrasts SLL-UM ($\eta_M/\eta_{SLL} \geq 10^2$), presents a greater change in the trench migration mechanism and slab morphology may vary withing the subduction domain (figure 5.4). M4 ($W_{SLL}/W_{OP} = 1$)

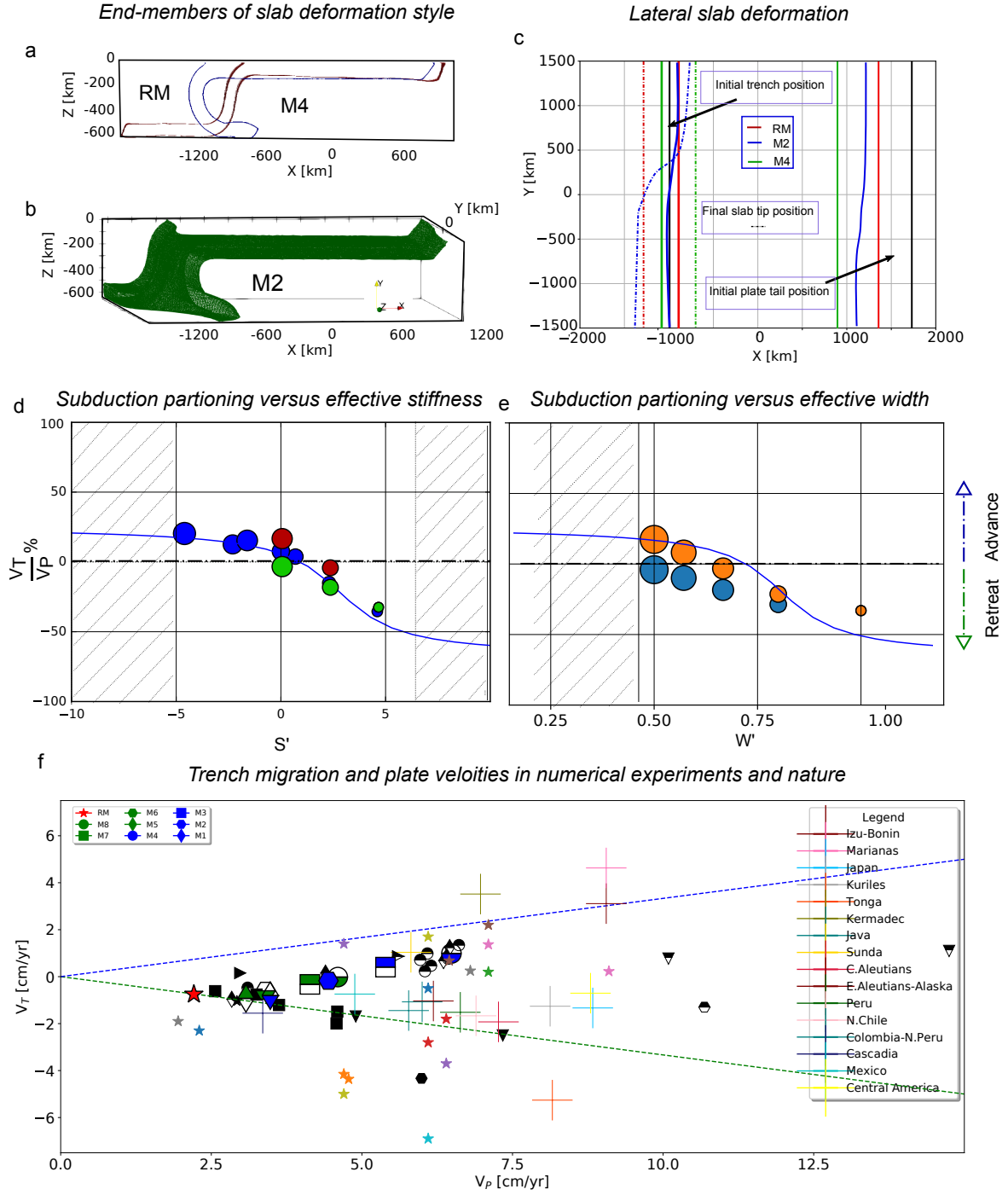


Figure 5.5: In a and b, a comparison among slab morphologies of RM, M2 and M4 cases at the steady-state stage of the subduction evolution. In c, a comparison among the initial and steady-state trench, plate tail and slab tip positions for the same 3D models. In d and e, the subduction partitioning (V_T/V_P %) as a function of the effective relative strength (S') and width (W') of the SLL progressively weaker and wider SLL (i.e., larger bullets). The 3D results for $W_{SLL} = W_{OP}$ (red) and $W_{SLL} = 1/2W_{OP}$ (green) are plotted on top of the values obtained in the 2D (blue) numerical study of Carluccio et al. (2019b). In f, the relationship between V_P and trench velocity V_T , as found in our 3D experiments (coloured polygons), in the 2D experiments of Carluccio et al. (2019b) (black polygons), in natural cases Goes et al. (2011) (coloured crosses) and in the no-net reference frame (coloured stars) with data taken from Schellart et al. (2008). The majority of the points lies between the dotted lines ($V_T = mV_P$) having m equal to $\pm 1/3$.

exhibits the largest plate velocity up to a factor of 2.5 compared to the RM case, continuous trench advance and a steady-state subducting slab that buckles at the 660 km discontinuity (figure 5.3c, 5.5a). Subduction experiments with $3/4W_{OP} \leq W_{SLL} < 1/4W_{OP}$, such as M2, shows large plate speeds and an overall advancing to quasi-stationary trench (figure 5.4d and e). In these models, we do not only observe variations in trench migrations but also in the morphology throughout the 3D slab. The area of the slab with no SLL shows a convex -s- shape, exhibits less trench advance and a slab flattening at the 660 km boundary; whereas, the remaining part of the slab exhibits a concave -c- morphology, has more trench advance and a steady-state slab buckling at the transition zone (figure 5.4e, 5.5b). The maximum variation in trench migration rate is once more $\Delta X t_{z_{surface}} \approx 70$ km showing that there is a limit on the amount of deformation an SLL can cause in the system. Ultimately, M1 ($W_{SLL}/W_{OP} \leq 1/4$) shows smaller lateral variations in trench retreating rates, more limited changed in slab morphology and plate velocities throughout the overall length of the subduction zone compared to the aforementioned models of this class.

We also observe a strong correlation between the value of the subduction partitioning and the effective relative width W' and strength, S' of the lithosphere, as shown in figure 5.5d and e). Here some important features are worth noting.

First, S' and W' share a similar inverse-tangent trend. Second, the function trend is asymptotic, which means that for low and high values of S' and W' the subduction partitioning doesn't change significantly. This shows again that the effect of an SLL on subduction motions is limited by the sinking dynamics, consistent with previous numerical work (Carluccio et al., 2019b). The increase in plate velocities can be up to a maximum of a factor of ≈ 2.5 , which occurs concurrently with the hampering of trench motions also consistent with previous findings (Carluccio et al., 2019b). Additionally, figure 5.5d outlines that the experimental curve co-tangent trend for W' remains the same for different viscosity contrasts UM-SLL ($10 - 10^2$).

Second, the asymptotic trend is further constrained by having $S' > 6.2$ and $S' < -5$ excluded from the graph (figure 5.5a). Such values of S' would correspond to plate-mantle and mantle-SLL viscosity contrasts above 500 and 1000, respectively. These are not found in nature and therefore not considered (Moresi and Gurnis, 1996; Goes et al., 2017). Similarly, values of widths larger or smaller than the oceanic plate and modelled box are also excluded as they are outside the extension of the modelled box, and hence have no physical meaning (figure 5.5d and e). Ultimately, for intermediate S' and W' , the subduction partitioning follows a co-tangential trend. Our results in figure 5.5d have been plotted on top of the 2D values obtained in the study of Carluccio et al. (2019b) to show that even if the magnitude of SLL effects on surface motions increases slightly from 2D to 3D, it maintains a constant trend.

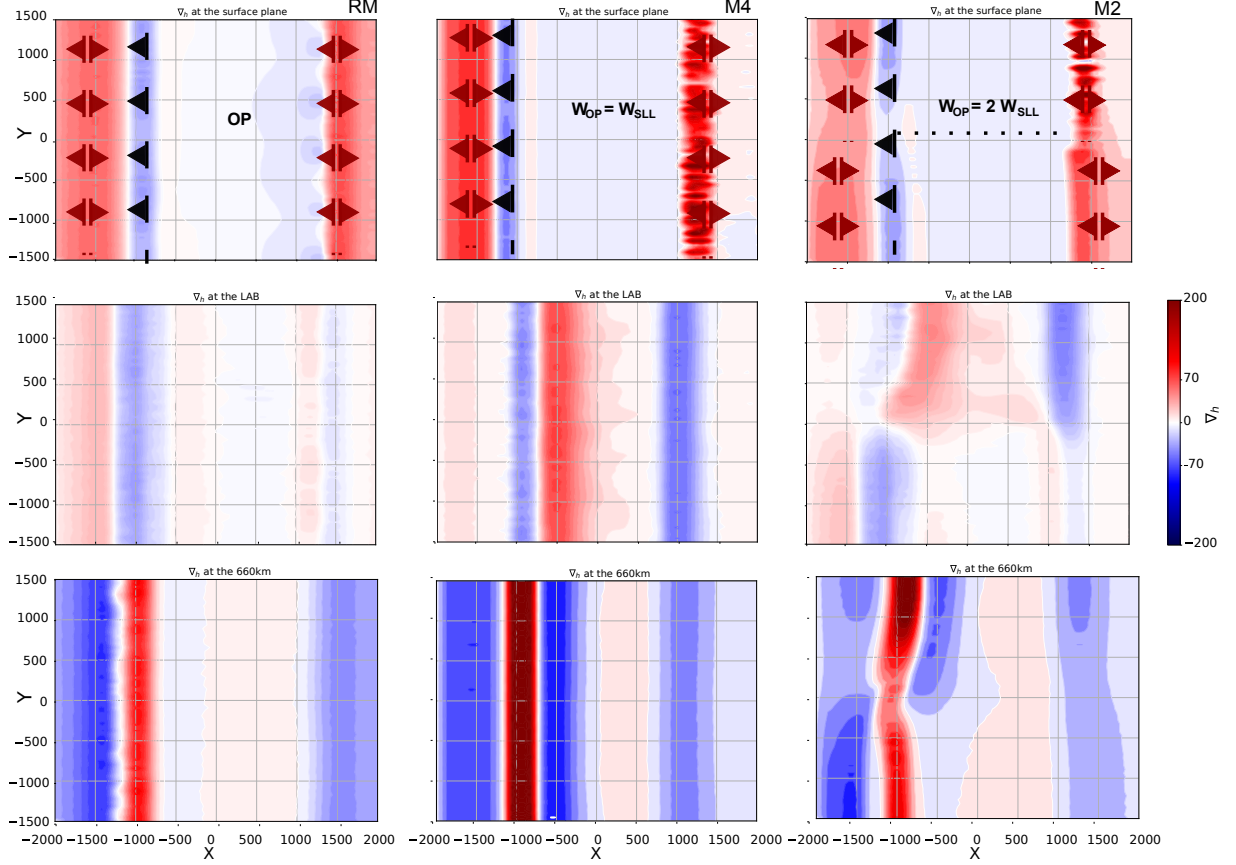


Figure 5.6: Horizontal divergence fields (left to right) for the RM, M4 and M2 at the steady-state stage in the proximity of (top to bottom) the surface (a, d, g), lithosphere-asthenosphere boundary (b, e, h) and 660 km bottom boundary of the model (c, f, h).

We also observe a striking linear relationship between plate and trench migration velocities in our numerical experiments (figure 5.5f). Trench and plate velocities increase linearly with both SLL width and viscosity. Our results suggest that W' and S' can be of crucial importance for the better understanding of the 3D mechanical role of a decoupling sub-lithospheric layer, and consequently the extent to which lithosphere-asthenosphere interactions can influence surface deformation and geological processes. Our findings (figure 5.5) support a mechanism whereby the subduction style and motions are controlled by both slab pull and asthenospheric drag exerted at the base of the plate balanced by the response of the viscous-flow induced in the adjacent mantle, in good agreement with previous studies (e.g., Ribe, 2001; Capitanio et al., 2007; Carluccio et al., 2019b; Coltice et al., 2019).

5.5.3 Effect of SLL width on the induced deformation flow

The temporal evolution of our models is accompanied by flow in the upper mantle around the subducting slab which can be decomposed into poloidal and toroidal flows for analysis.

The strength and intensity of each of these components may vary during the subduction process.

To explore the extent to which an SLL may influence the vertical transport of mass in a subduction model, we analysed the divergence of the velocity field which is associated with the poloidal component. We observe two main characteristics in common to all the models: i) surface divergence is higher at ridges and above the slab, and smaller at the trench (i.e., at the convergence zone) and ii) this trend is attenuated at the LAB; and iii) the trend in the values flips deeper in the upper mantle (figure 5.6). The addition of an SLL causes variation in surface divergence field around the plane of symmetry of the subduction zone, $Y \sim 0$ (figure 5.6 top-right) and due to very low and asymmetric trench motion (Figure 5.4e), concurrent with the development of toroidal flow (Figure 5.7f). The lateral variation in ∇_h close to the surface are milder than at deeper depths, but stronger than in RM and M4. Behind the subducting slab, (figure 5.6 mid and right columns), the addition of an SLL generally causes the development of a region of strong divergent motion in M4 (figure 5.6 mid-column). The M2 case exhibits a lateral variation in the sign of the divergence field accordingly with SLL lateral extension at the LAB (figure 5.6 right). Towards the bottom boundary, an additional lateral change in the divergence field is due to slab variation in morphology from -c- to -s-shape from one side to the other of the subduction zone.

The change in polarity and strength of ∇_h at XZ planes underneath the three-dimensional slab may be recorded in the orientation of seismic anisotropy and help to explain peculiar signatures encountered in the geological record throughout a subduction zone. The comparison of the vorticity field for Rm, M4 and M2 is shown in figure 5.7. Figure 5.7 allows us to test and verify that there is no substantial vorticity in models with no lateral viscosity variations regardless of depth, such as RM and M4. Conversely, it is possible to observe the development of a strong toroidal cell at the base of the oceanic lithosphere for M2, which is active throughout the whole model domain (figure 5.9). The results show that the toroidal component of the deformation flow is responsible for lateral variations in trench migrations and plate velocities surface motions (figure 5.4), as well as slab morphology (figure 5.3 and 5.5b).

5.5.4 Temporal variation of the induced deformation flow

The trends in the evolution through time of the poloidal potential, as well as the intensity of the poloidal cells, follow that three classical stages in the evolution of the subduction speed, as shown in figure 5.4. The largest values of poloidal potentials are generally registered at the base of the model caused by viscous friction due to the interaction of the slab with the 660 km discontinuity. Intermediate and lower values are found at the surface and the LAB, respectively. Common to all the simulations with an SLL is the development of a more intense poloidal cell behind the subducting slab. In all these

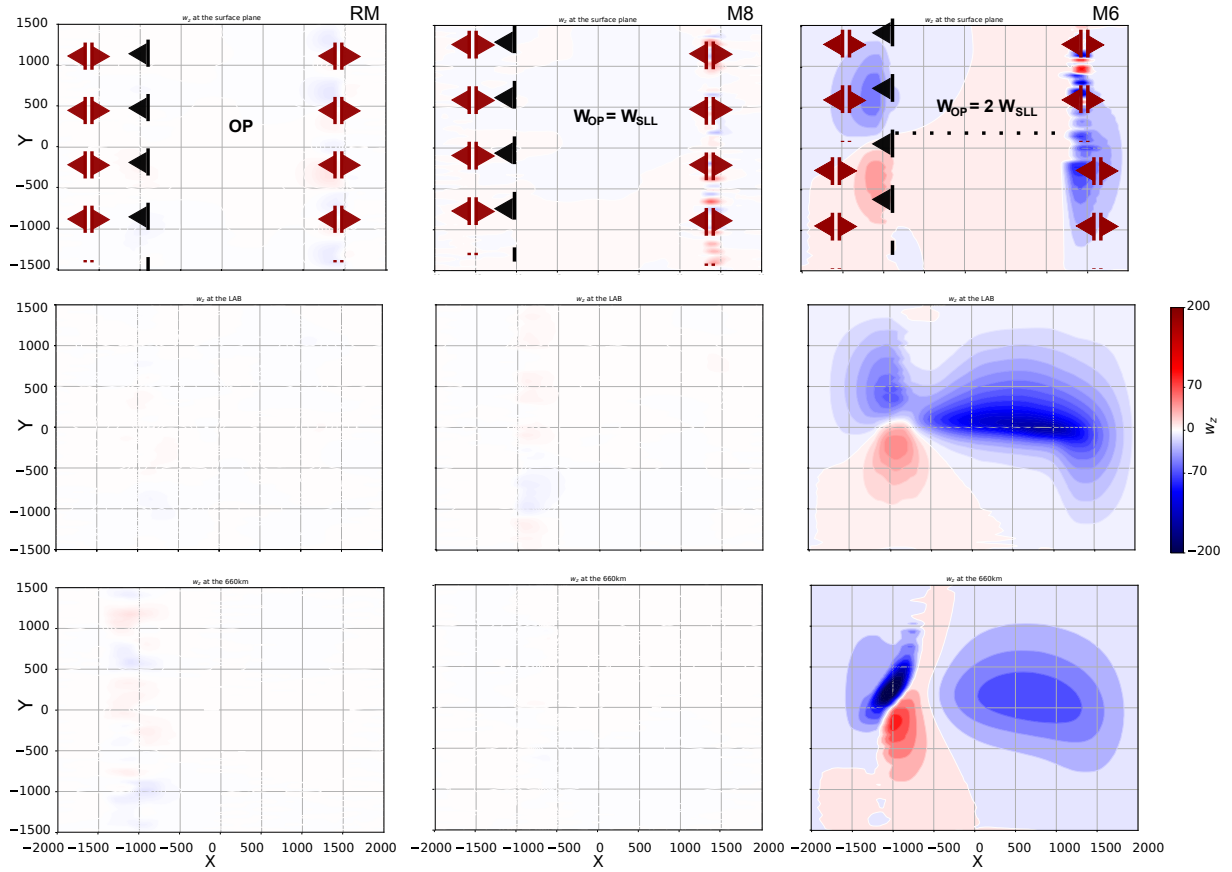


Figure 5.7: Radial vorticity fields (left to right) for the RM, M4 and M2 at the steady-state stage in the proximity of (top to bottom) the surface (a, d, g), lithosphere-asthenosphere boundary (b, e, h) and 660 km bottom boundary of the model (c, f, h).

models both the components of the deformation field are active from the early stages of the evolution of the model (figure 5.8, 5.9). Overall the largest poloidal potential is observed for M4, which is also the model characterised by the largest velocity (figure 5.3).

The poloidal component represents the predominant flow for all the models, except for when is compared to SLL toroidal flow at the LAB. In this case, the toroidal:poloidal ratio can reach values up to a factor 2 (figure 5.9c). The toroidal cell can generate a significant toroidal motion around the LAB, $\Psi_{z=LAB}$ followed by $\Psi_{z=660km}$ and then $\Psi_{z=surface}$ (figure 5.5, 5.8, 5.9). We find maximum values of $T/P \approx 0.5$ and ≈ 2.4 at the surface and at the LAB, respectively. The RM and M4 show $T/P \approx 0$ stable through time, since they have no associated lateral viscosity variations in the along-strike direction due to the lack of symmetry breaking. Figure 5.9 ultimately summarises our end-members results and show a striking relationship that relates SLL width to the partitioning of subduction surface motions and induced mantle flow. Hence, highlighting the significance of thin, uneven and weak decoupling layers embedded underneath the base of a subducting plate for the better understanding of mantle-lithosphere interaction, and their impact on surface deformation. Thus, these results do not only suggest that a decoupling sub-lithospheric layer affects surface motions and induced patterns of mantle flow, but it may also be crucial for our understanding of the secular evolution of the planet, as it can exert a critical control on T/P .

5.5.5 Application to natural subduction zones

Our numerical simulations employ varied parameters over a range that is Earth-like (figure 5.2b and table T1). Our full set of models include the results obtained for a different initial plate width with a free-edge, other crustal rheological parameters. To test the validity of our 3D experiments, we use the estimates on V_P and V_T from Goes et al. (2008) on the Indo-Atlantic hotspot, as well as Schellart et al. (2008) built on the no-net reference frames from (O'Neill et al., 2005) for major subduction zones in the Pacific. The values of V_P and V_T from our models most positively align with the observed present-day plate and trench migration rates constrained within $V_T \pm 1/3V_P$ (figure 5.5e). However, along some of the subducting segments of the Pacific rim, there are zones showing little correlation with our models (i.e., $V_T > \pm 1/3V_P$), such as the Tonga and Sunda trenches. Nonetheless, many of the remaining subduction regions experience significant lateral changes in slab morphology and trench migrations mechanisms, such as Mariana, Izu-Bonin and Nazca, among others. If an SLL would present at these zones, it could provide a means to help to explain some the intricacy in patterns of deformation we observe in natural systems. While it is plausible that other mechanisms may be of major regional importance or coexist with the presence of an SLL (i.e., double subduction systems, arcuate arcs), an SLL would be expected to play a crucial role in the partitioning of surface and deep induced motions, as well as broad subduction dynamics. Additionally, our resultant values of vorticity and

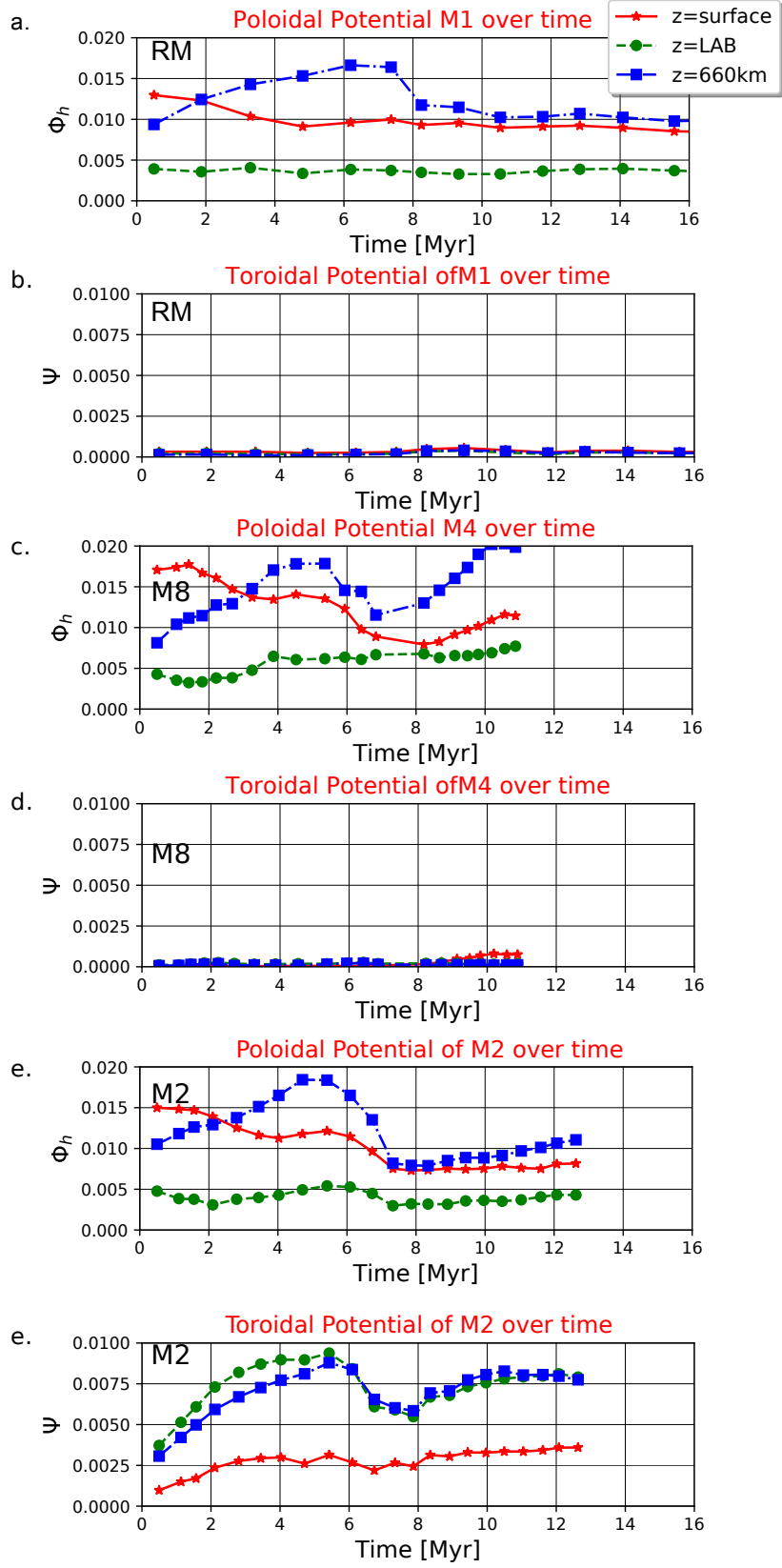


Figure 5.8: A comparison between the time evolution of the poloidal potential, Φ_h , and the toroidal potential, Ψ , for the RM (a, b), M4 (b, c) and M2(d, e) cases.

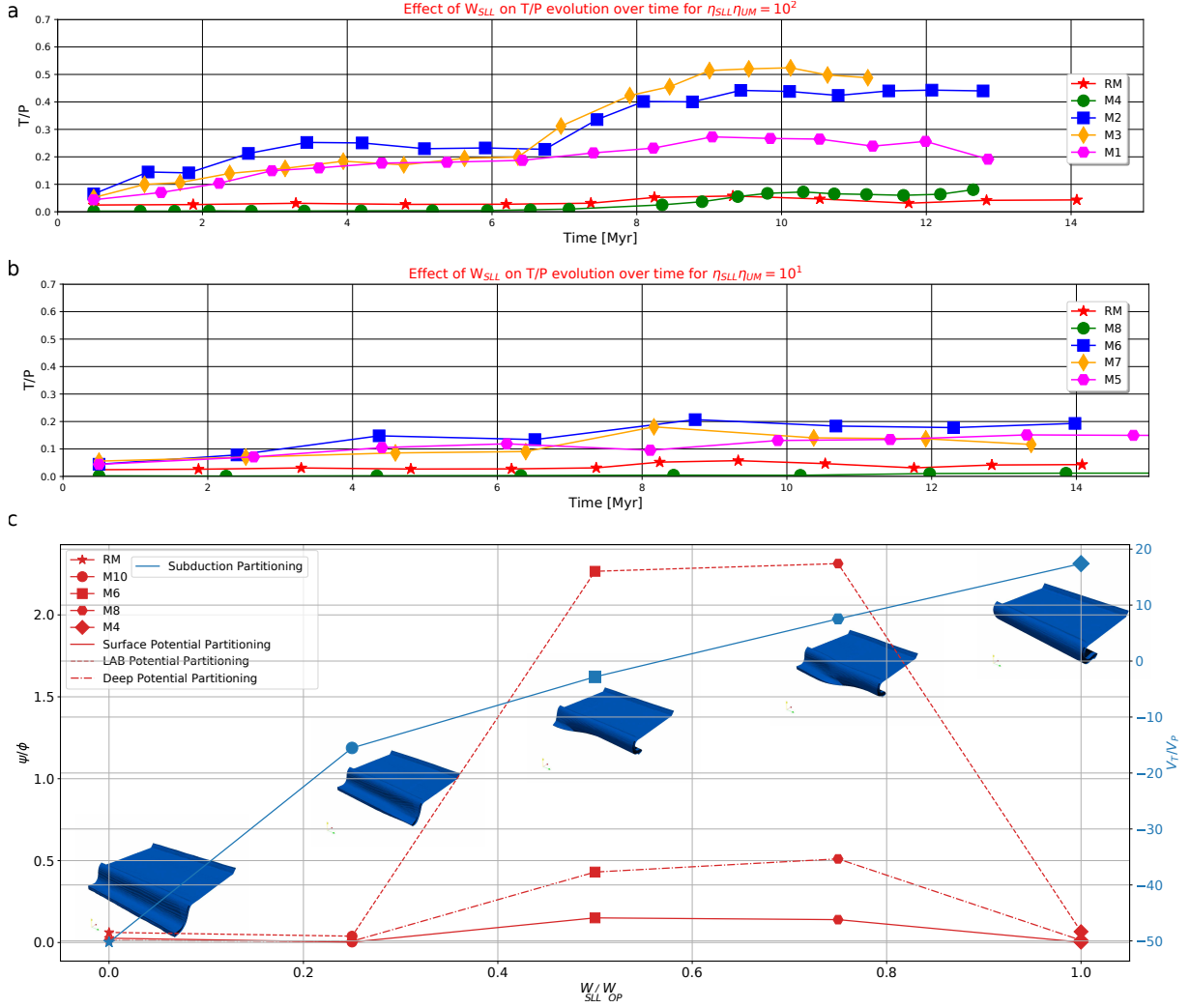


Figure 5.9: A comparison between the time evolution of the partitioning of induced mantle flow (T/P) for models with viscosity contrast mantle-SLL of 10^2 (a), and 10^1 (b). In c, the relationship between T/P, surface motions partitioning (V_T/V_P %) and SLL- plate width ratios ($W^* = W_{OP}/W_{SLL}$) at different XY cross-sections through depth (surface, LAB and depth).

divergence fits well with the average data for the present-day superficial velocity field obtained from the analytical NUVEL-1 model of O'Connell et al. (1991), (figure 5.1). As such these findings aid the interpretation of patterns of deformation flow characterised by localised peak in high and low values at plate margins, such as those observed in Mariana, Izu-Bonin and Nazca natural subduction zones.

5.6 Discussion and concluding remarks

A range of geophysical studies (Kawakatsu et al., 2009; Schmerr, 2012; Naif et al., 2013; Stern et al., 2015; Hawley et al., 2016) has recently brought considerable attention to the detailed structure of the asthenosphere at subduction zones. These report that astheno-

spheric local and regional configurations may lead to up to a ≈ 4 orders of magnitude melt-induced viscosity reduction focussed into one or multiple thin and more buoyant layers at the base of the lithosphere. Such regional and local ultra-viscosity reductions appear compatible with global plates and seismic anisotropy models as long as their lateral extent remains limited compared to the rest of the tectonic plate (Becker, 2017). Yet, poor or no constraints currently exist on the lateral extent of these features and the impact they may have on 3D subduction dynamics.

Our numerical models addressed the impact of combined SSL viscosity and width anomalies on the partitioning of subduction surface motions, induced mantle circulation and slab morphology. We use the plate and trench migration rates, the vertical component of the mantle vorticity and the horizontal divergence vectors, as well as, the poloidal and toroidal potentials and their ratios to characterize the subduction induced dynamics and mantle flow in terms of geometry, strength and degree.

In our study, the primary effect of changing SLL width is to produce variation in the slab subducting speed promoting the development of substantial toroidal flow at the LAB. This influences slab and trench migrations, the T/P amplitude ratio and, in some cases, slab morphology throughout the 3D subduction zone (figure 5.3- 5.9c). Additionally, an SLL promotes plate advance and hampers trench retreat, consistent with the 2D study of Carluccio et al. (2019b).

Our results show that for intermediate to large viscosity contrasts ($\eta_M/\eta_{SLL} \geq 10^2$) a significant change in velocity, trench migration mechanism and morphology develops through the subducting slab (figure 5.4, 5.5a and b), where the impact depends on SLL width. Models with $W_{SLL} > 3/4 W_{OP}$ show the larger plate velocities, trench advance rates and a steady-state subducting-slab continuously buckling at the 660 km discontinuity. Subduction experiments with $3/4 W_{OP} \leq W_{SLL} < 1/4 W_{OP}$ show significant plate speed and change in slab morphology (s-to c shape) and trench migration (retreat to advance) mechanism from one side to the other of the subduction zone (figure 5.3, 5.4). Models with reduced SLL width ($W_{SLL} \leq 1/4 W_{OP}$) and/or viscosity contrast ($\eta_M/\eta_{SLL} < 10^2$) show overall smaller variations in trench migration rates, plate velocities and slab shape throughout the subduction zone, which is generally in retreat. It is worth noticing that the lateral variation in trench migration is constrained to a maximum of $\Delta X t_{z_{surface}} \approx 70$ km and the increase in V_P to up to a factor of ~ 2.7 , independently of the viscosity contrast. Hence showing that there is a limit on the amount of deformation an SLL can provide to the system (figure 5.5b, d and figure 5.9c). Additionally, an SLL produces only a limited effect on the time necessary for the slab tip to reach the bottom of the model; thus, showing once again that an SLL has a greater influence on tangential motions than on vertical movements in agreement with previous work (Carluccio et al., 2019b).

In accordance with previous work (Piromallo et al., 2006; Stegman et al., 2006; Funiello et al., 2008; Capitanio and Faccenda, 2012; Király et al., 2017), we also find that

the subduction-induced mantle flow to be highly three dimensional where poloidal and toroidal flow components coexist since the early stage of subduction (Funiciello et al., 2004; Schellart, 2004b), as shown in figure 5.9. We generally register higher T/P in correspondence of the slab mid-mantle depth, in agreement with the work of Holt et al. (2017); however, we record the highest T/P at the LAB and for the most asymmetric and steeper slabs (figure 5.9). If in the literature slab and mantle properties contrast (Király et al., 2017), ridge position and plate coupling at the trench zone (Gérault et al., 2012) have shown to exert strong controls on trench motions and flow components, yet these factors have not explicated a clear trend in the partitioning of the subduction induced mantle flow and surface motions.

For the very first time, we have explored this impact in light of the recent observation of a decoupling SLL at the base of subducting zone in a 3D fashion. Furthermore, while in many previous studies toroidal flow component emerged from laterally discontinuous plate models with a slab free-edge, we were able to achieve substantial Ψ ($T/P \sim 0.2-2.$) by only including an SLL (figure 5.9) and by employing a continuous slab model. The typical T/P values we obtain at the surface range from 0.2 to 0.5, consistent with what would be expected in a no-net reference frame, and increase up to 2.0 deeper near the base of the subducting plate. Our values of vertical vorticity and divergence also fit well what is expected at natural convergent margins $150-200 \text{ Gyr}^{-1}$ (figure 5.1). The perturbation of the induced flow components around the subducting slab and the trench zone has a significant role on the global geological circulation. It can be responsible for the presence of the off-arc volcanism in subduction zones, (Long and Silver, 2008; Faccenna et al., 2010; Strak and Schellart, 2014) with consequences for magma genesis, volatile transport and thermal structure, as well as substantial slab deformation, including strike-slip motion, and subduction anisotropy (O’Connell et al., 1991; Long, 2013; Coltice et al., 2017).

The strength and intensity of poloidal and toroidal flow components may vary during the subduction process to follow the evolution of the velocity field (figure 5.8). However, all our simulations with an SLL show a more intense poloidal cell behind the subducting slab, which results from the larger velocity of the oceanic plate and viscous shearing at the base of the lithosphere being accommodated within a thin and weak sub-lithospheric interval. Our outcomes show how SLL lateral extent and viscosity influence the development of toroidal flow, which in turn is responsible for the lateral change in slab morphology and subduction surface motions across the subduction region (figure 5.5). We also outline a simple scaling relationship to relate the effective width and viscosity of the lithosphere relative to the underlying mantle to the subduction partitioning (figure 5.5). Moreover, figure 5.9 highlights a striking linear relationship relating V_T/V_P to W_{SLL} , and a quadratic relationship connecting T/P to W_{SLL} , and the subduction styles; hence demonstrating the influence of mechanical processes occurring at the LAB on subduction slab evolution and surface deformation. Our chosen model setup allows us to test and verify that substantial

toroidal flow when is found is not due to an artefact of numerical leakage, but that slab change in morphology, as well as low trench migration and plate speed, can be the result of a simpler force balance around the subducting plate, whilst no extra forces or factors are needed. We do not exclude that other factors which have already shown to reduce trench motions or affect mantle induced flow may affect or coexist with the mechanism illustrated here; however the decoupling effect of SLL shows a consistent signature even with the inclusion of a visco-elastic-plastic temperature-dependent rheology or the upper/lower mantle transition zone (Carluccio et al., 2019b). We also employ a small number of additional models to explore the role of a free slab edge subducting setup, which would naturally lead to vorticity at the slab free edge or the lower mantle (supporting figure S2-S3). While the choice of parameters can be different, an SLL maintains a consistent signature.

The importance of tangential forces (e.g., the toroidal field) to improve our numerical capacity to match modelling outcomes with expected present-day values, and possibly past, of the Earth’s deformation field (figure 5.1 and surface motion 5.5e) has long been proposed (e.g., Bokelmann, 2002), along with the fact that plates and mantle cannot be fully coupled to be able to explain the large toroidal flow component we currently observed at plate margins. However, thin and narrow channels might have remained “hidden” and their dynamic role not thoroughly explored because several difficulties hinder the ability to fully image the extent of these layers and gain constraints about their physical properties (Cathles, 1975; Fischer et al., 2010; Rychert, 2015; Richards and Lenardic, 2018)). These difficulties include the the non-uniqueness relating to viscosity and thickness of deep Earth layer, elusive nature of the LAB and trade-off between power and lateral extension of geophysical methods, among others (Cathles, 1975; Fischer et al., 2010; Rychert, 2015; Richards and Lenardic, 2018)). Yet, thin asthenospheric channels can accommodate horizontal shearing of the lithosphere much better than distributed vertical loading or unloading over regions much larger than the channel depth and could decouple plate motion from mantle tractions thus contributing to the global net rotation of a few cm/yr (Scoppola et al., 2006).

Variations in channel width have shown to induce significant rotation at the base of the lithosphere generating an additional toroidal cell at the mid-plane of symmetry of the plate; hence increasing the amplitude of the toroidal to poloidal ratio. These outcomes provide an alternative explanation and a novel mechanism for the apparent lack of a direct link between the slab pull driving force in buoyant convection (which drives only vertical and divergent motion) and the generation of toroidal energy. Thus, showing how buoyancy-driven poloidal motion can induce substantial toroidal flow through horizontal viscosity variations.

Additionally, these findings outline a means to interpret variations in slab dip, morphology and trench migration throughout a subduction zone as those observed along some

of the subducting margins of the Pacific plate where the presence of an SLL is confirmed by geophysical studies, such as in Mariana, Izu-Bonin and Nazca subduction regions. We suggest that a regional SLL mechanism may have important consequences for global subduction dynamics, and possibly for the better understanding of mantle evolution through geologic time.

5.7 Supporting Information for “On the Influence of Sub-Lithosphere Width on the Partitioning of the Subduction Induced Mantle Flow”

Contents of this file

1. Figures from F10 to F12

In this section additional subduction modelling results are provided. Figure F10 shows the XZ divergence velocity fields at constant Y ($Y=1500$ km, $Y=0$ km and $Y=-1500$ km) for RM ($\eta_M/\eta_{SLL}=1$, $W_{SLL}=0 W_{OP}$), M2 ($\eta_M/\eta_{SLL}=100$, $W_{SLL}=1/2 W_{OP}$) and M4 ($\eta_M/\eta_{SLL}=100$, $W_{SLL}=W_{OP}$) at the steady-state stage of the time evolution.

Figures F11 and F12 show the XY vorticity and divergence field for a slab free-edge models where the plate width is half of the box width, in for M10 ($\eta_M/\eta_{SLL}=1$) and $W_{SLL}=0 W_{OP}$), M14 ($\eta_M/\eta_{SLL}=100$) with $W_{SLL}=1/2 W_{OP}$) and M12 ($\eta_M/\eta_{SLL}=100$).

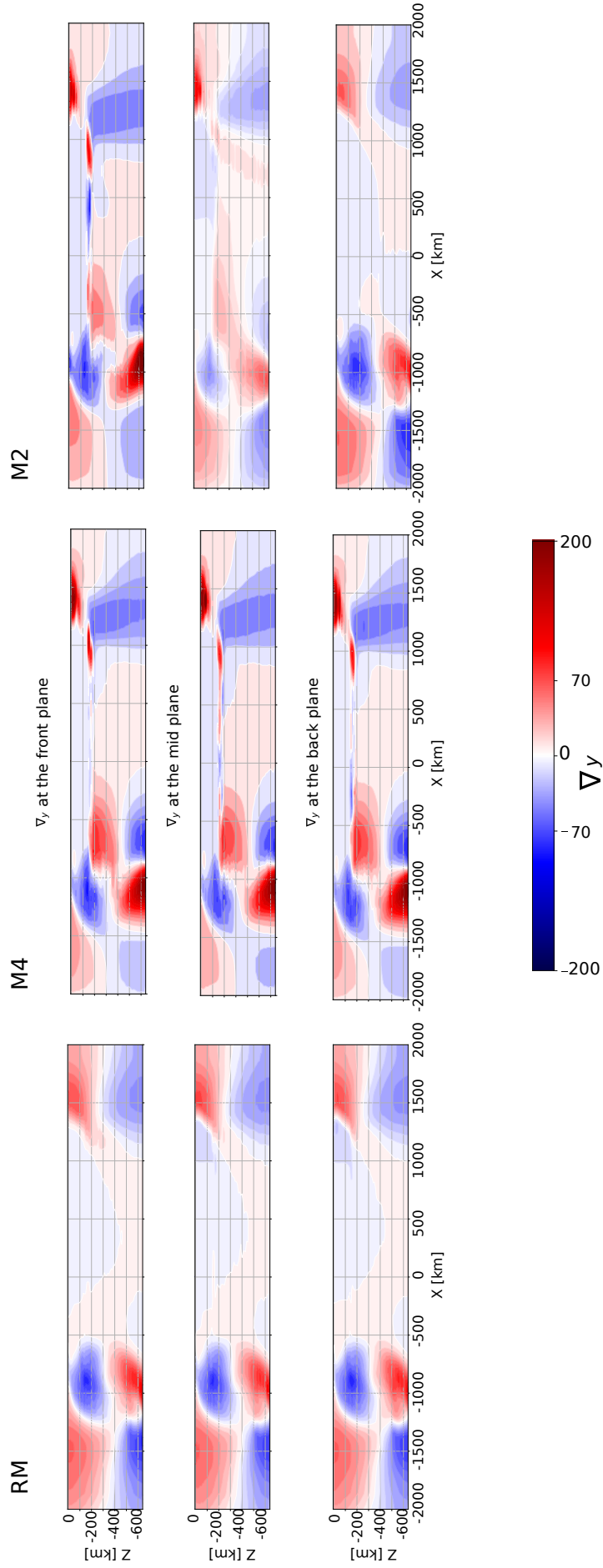


Figure 5.10: Vertical divergence (XZ) for the RM M10 ($\eta_M/\eta_{SLL} = 1$, $W_{SLL}=0$ W_{OP}), M14 ($\eta_M/\eta_{SLL}=100$, $W_{SLL} = 1/2W_{OP}$) and M12 ($\eta_M/\eta_{SLL}=100$, $W_{SLL} = W_{OP}$) cases (left to right) at the steady-state stage of the evolution of the model taken at different XY cross sections towards the front, the centre and at the far lateral end of the model.

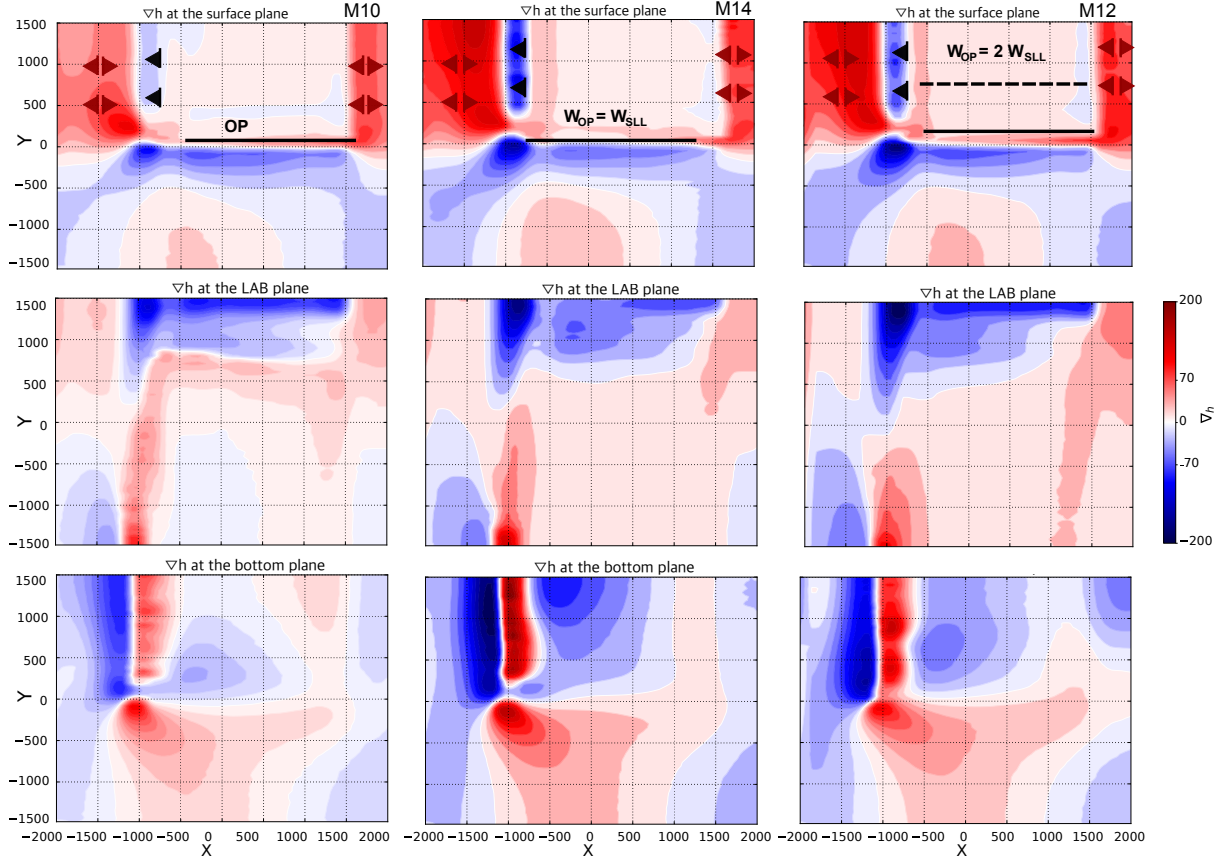


Figure 5.11: Horizontal divergence for a slab free-edge models where the plate width is half of the box width, for M10 ($\eta_M/\eta_{SLL} = 1$, $W_{SLL} = 0 W_{OP}$), M14 ($\eta_M/\eta_{SLL} = 100$, $W_{SLL} = 1/2 W_{OP}$) and M12 ($\eta_M/\eta_{SLL} = 100$, $W_{SLL} = W_{OP}$) cases (left to right) at the steady-state stage of the evolution of the model. These are shown in the proximity of the surface (top), lithosphere-asthenosphere boundary (middle) and 660 km (bottom) boundary of the model.

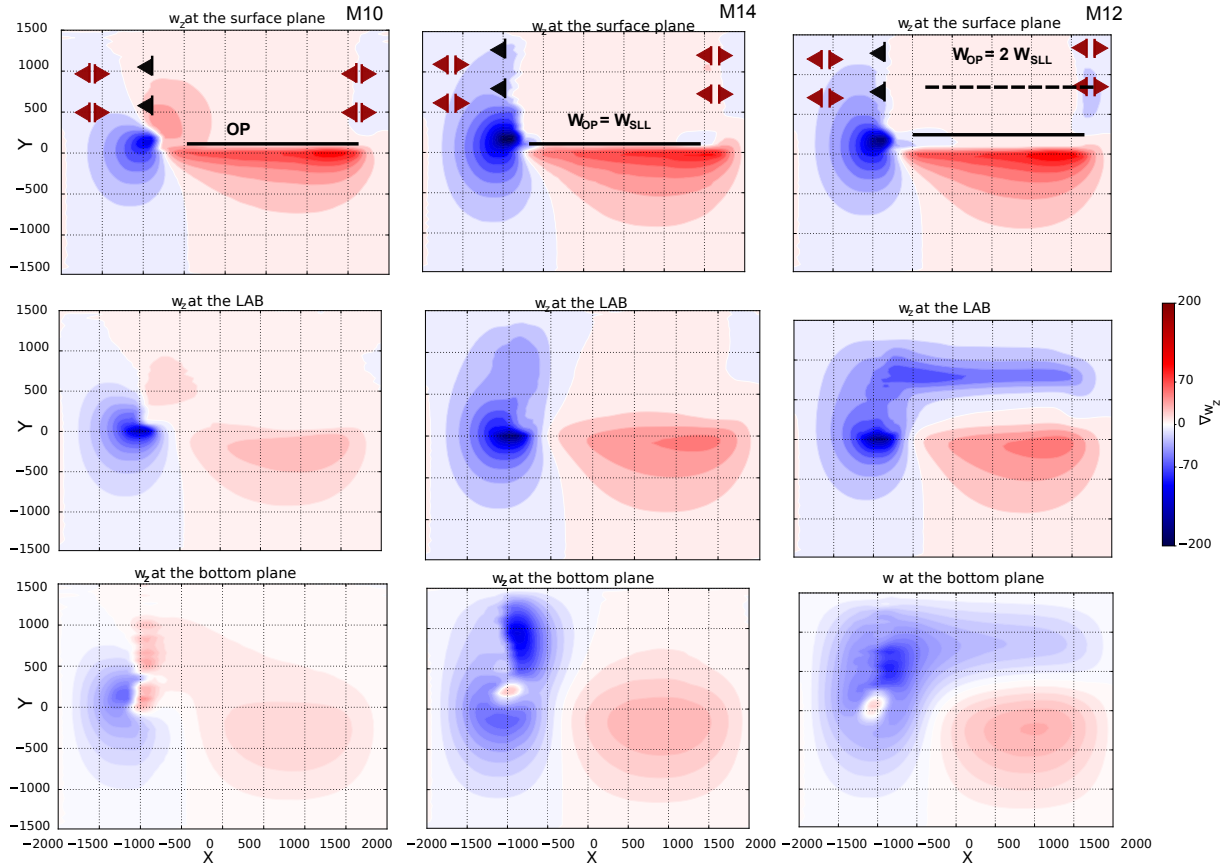


Figure 5.12: Radial vorticity Horizontal divergence for a slab free-edge models where the plate width is half of the box width, for M10, M14 and M12 cases (left to right) at the steady-state stage of the evolution of the model. These are shown in the proximity of the surface (top), lithosphere-asthenosphere boundary (middle) and 660 km (bottom) boundary of the model.

6. Chapter 6: Asthenosphere dynamics during subduction

Asthenosphere dynamics during subduction

1 2

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6.1 Highlights

- By compiling observations reporting an abrupt seismic velocity drop and/or electrical conductivity increase within an ultra-low velocity zone, ULVz, in the upper asthenosphere, we find a spatial relation with the presence of a relict oceanic plateau, OPl, and a ULVz;
- we explore ULVz and OPl individual and combined influences on subduction dynamics using three-dimensional subduction numerical models, in addition to testing the seismic anisotropy hypothesis that trench-parallel flow in the sub-slab could be induced by the presence of a ULVz;
- our results show that mantle flow can be divided in two domains depending on depth, which we refer to as the slab-mantle and sub-slab mantle domains. In the former, the misalignment in the orientation of the maximum finite strain axes, γ , scales negatively with the ratio between trench and plate velocity V_{norm}^* , and in the latter γ scales positively with it;
- A ULVz promotes a trench-perpendicular pattern in the slab-mantle and trench-parallel flow in the sub-slab mantle and wedge domain, which would otherwise be trench-parallel to perpendicular in the sub-mantle and slab-parallel in the sub-slab mantle domain. We propose a seismic anisotropy model where variations in ULVz width promote miscellaneous patterns of mantle flow within the same subducting zone.
- our findings could be of global applicability and particularly important for those subduction zones where plates move faster than their trenches retreat, as well as where miscellaneous patterns of mantle flow exist, with even more important consequences for our own understanding of subducting slabs evolution over geological time.

¹ *Preprint submitted to Geology*

² *Keywords: subduction modelling, seismic anisotropy, mantle dynamics*

6.2 Abstract

The importance of an asthenosphere for the very existence of the Earth plate tectonic regime has long been recognised; however, a puzzling question remains as to how reconcile asthenosphere dynamics with surface observations, especially at subduction zones. Using a global subduction zone dataset and three-dimensional numerical subduction models, we show that a decoupling layer directly beneath subducting plates promotes the development of miscellaneous patterns of surface motions, mantle flow, and seismic anisotropy along different segments of the same subduction zone. We compare our results with data from 19 subduction zones, revealing systematic patterns for oceanic lithosphere and shallow mantle flow scaling negatively with the ratio of trench and plate velocity, V_{norm}^* , and deeper patterns scaling positively it. Our results show that mantle flow patterns reflect in seismic anisotropy signatures that can be used along with the relative magnitude of V_{norm}^* , as a proxy to identify the amount of decoupling at the lithosphere-asthenosphere boundary. We offer a novel outlook for interpreting slab dynamic evolution on a global scale allowing us to reconcile previous findings with applicability across a broad spectrum of the geosciences.

6.3 Background

The Earth’s mantle circulation drives plate-tectonic cycles of continental drift over hundreds of millions of years and the changing pattern of continents profoundly influences the evolution of life on our planet. By lubricating the undersides of the tectonic plates, the presence of an asthenosphere allows them to move freely over the mantle and eventually descend underneath. The boundary between the lithosphere and asthenosphere (LAB) accommodates differential motions between the strong lithospheric lid and weaker layer underneath. The LAB corresponds with a strong vertical drop in seismic waves speed or impedance, including changes in the strength and orientation of both radial and azimuthal anisotropy. This is known as the Gutenberg discontinuity and defines the upper boundary of the low viscosity zone, LVZ, relating to a weak, $\sim 0.5 \times 10^{18} - 10^{21}$ Pa·s (Mitrovica, 1996) and extended 100-300 km thick layer of asthenosphere (Becker, 2017). The LVZ corresponds to a bulk ≈ 2 orders of magnitude viscosity jump compared to the above lithosphere and mainly due to pressure and temperature changes in the medium (Richards and Lenardic, 2018).

More recently a range of geophysical studies (Naif et al., 2013; Stern et al., 2015; Hawley et al., 2016) has brought considerably more attention to the detailed structure of the asthenosphere at subduction zones. These studies report an additional abrupt velocity or electrical conductivity drop (henceforth referred to as an ultra low-velocity zone, ULVz) within the upper asthenosphere compatible with up to four orders of magnitude viscosity

drop compared to the background mantle flow, concentrated in one or multiple channels, with a combined vertical thickness of no more than 50 km (Bagley et al., 2009). Such a sharp drop in seismic properties cannot be easily explained by invoking the role of pressure and temperature alone, as these would require a gradual transition occurring within a thicker layer (such as the LVz). The boundary is instead identified as a sharp transition and other processes such as the degree of melting, water and/or volatile phases availability have been, invoked to justify its origin (Kawakatsu et al., 2009). While the better understanding of how these channels operate necessitates more constraints on where they exist, the non-uniqueness relating viscosity and thickness of deep layers, as well as the trade-off associated with geographical extension and power of geophysical methods can undermine the capacity to detect ULVz. Such that a ULVz could remain “hidden” to seismic waves passing through it. With the intent to find a correlation between subduction zone properties and seismic observables, we gather information about subduction parameters of interest around 19 subduction systems, as well as compiling observations from 12 published studies interpreting the LAB as a sharp boundary. By compiling this information, we find a spatial relationship relating the presence of a relict oceanic plateau, OPl, and a ULVz along some segments of the subducting Pacific plate (figure 6.1).

An oceanic plateau corresponds to an over-thickened (5–30 km, sometimes >30 km) vast area ($>0.1 \times 10^6$ km²) of anomalously buoyant crustal layer generally formed by decompression melting of hot mantle plumes, broad continental crust or micro-continents (Kerr, 2003). While discussion of the origin of ULVz and OPl remains beyond the scope of this work, some lines of evidence for a ULVz currently exist. One consists of the hypothesis of a plume fed asthenosphere for the origin of ULVz (Phipps Morgan et al., 1995; Yamamoto et al., 2007; Morgan et al., 2013), which could justify the co-occurrence of ULVz with OPl. Another hypothesis entails the presence of a fossilised base of the lithosphere resulting from a rise in hydration with depth of lithospheric rocks as they move away from spreading ridges, which would have decreased solidus temperature creating a damp zone directly underneath the lithosphere (Revenaugh and Jordan, 1991; Gaherty et al., 1996; Hirth and Kohlstedt, 1996). LAB inherited weakening gained during periods of large crust formation could perpetuate through the base of the lithosphere due to shear heating production caused by discordant motion between oceanic plate movement and mantle flow underneath (Turcotte and Schubert, 1982). Perpetual shear heating could induce large amounts of strain and contribute to melt generation within a thin sub-lithospheric interval. Experiments on partially molten olivine aggregates and lherzolite rocks show that perpetual shear deformation at the LAB promotes melt ability to segregate into lenses and channels (Kohlstedt et al., 1995; Holtzman et al., 2003; Zimmerman and Kohlstedt, 2004). Melt generation at this interface may allow such heterogeneities to become shear zones, some of which are still present in the geological record (Kelemen and Dick, 1995; Kennedy et al., 2002) and possibly induce giant mega-thrust earthquakes (Fan and Zhao,

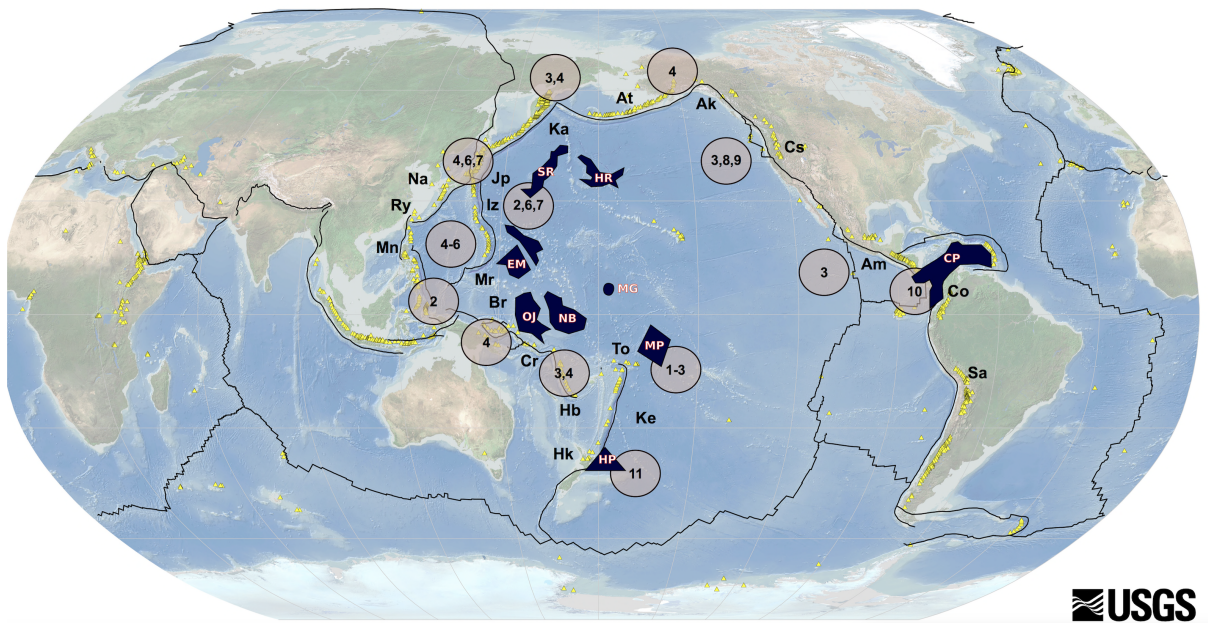


Figure 6.1: A compilation of geophysical studies that interpret the LAB as a sharp boundary along some major subduction zones. 1: Gaherty et al. (1996), 2: Revenaugh and Jordan (1991), 3: Schmerr (2012), 4: Bagley and Revenaugh (2008), 5: Gaherty et al. (1999), 6: Kawakatsu et al. (2009), 7: Bagley et al. (2009), 8: Rychert et al. (2018), 9: Hawley et al. (2016), 10: Naif et al. (2013), 11: Stern et al. (2015) and 12: Wang et al. (2020). Map showing all major present-day oceanic plateaux formed within the last 150 Myr (modified after Kerr (2014)) in dark blue: HP: Hikurangi; MP: Manihiki; OJ- Ontong Java, NB: Nauru Basin; EM- East Mariana Basin; PF: Pigafetta; SR- Shatsky Rise, HR- Hess Rise, MG- Magellan Rise, CP- Caribbean. Subduction zone segments: Hk – Hikurangi; Ke – Kermadec; To- Tonga; Hb- New Hebrides; Cr- San Cristobal; Br- New Britain; Mn- Manila; Ry- Ryukyu; Na- Nankai; Mr- Mariana; Iz- Izu-Bonin; Jp -Japan; Ku- Kuril; Ka- Kamchatka; At- Aleutian, Ak- Alaska; Cs- Cascadia; Am- Central America; Co- Colombia; Sa- South America. Figure edited from NOAA National Geophysical Data Center (NGDC).

2021). In addition to this, it has been shown that oceanic plateaux can substantially alter subduction dynamics by inducing deformation in the subducting plate and substantially affecting trench curvature (Moresi et al., 2014). Given that an OPl might be expected to co-occur with a ULVz, our three-dimensional (3D) numerical models explore both their individual and combined influences on subducting slab evolution and patterns of seismic anisotropy, which are often used to draw conclusions about subduction dynamics.

Seismic anisotropy can provide a direct constraint on mantle dynamics because of the link that exists between deformation and resulting anisotropy. The directional dependence of seismic wave speed results from the development of induced lattice preferred orientation of individual mineral crystals (LPO) or shape-preferred orientation of mantle materials (e.g., melt) under deformation. Seismic anisotropy, therefore, represents a valuable tool for probing mantle dynamics. Around the subduction zone, wave splitting

corresponds to regional anomalies in the direction/geometry of maximum strain, ϕ , (e.g., a composite pattern of trench-parallel, $T_{\parallel}$, oblique, $T_{\backslash}$, and trench-perpendicular, $T_{\perp}$) and provides information about the strength of the anisotropic layer through the delay time, δt . Regional anomalies are generally recorded in four different areas of the subducting system, such as the slab, sub-slab and wedge mantle, as well as the overriding plate. By global analysis of seismic anisotropy signatures, a puzzling trend emerges from observations. This trend is not often compatible with the simplest model for mantle dynamics, which invokes two-dimensional (2D) corner flow caused by overriding plate motion and simple (A-, C-, or E-type) olivine LPO fabric in the subducting plate, as it cannot adequately predict splitting in the observations of many natural subduction settings (Lynner and Long, 2014b). In these natural settings, an alternative explanation is required for the transition from A-type to B-type olivine LPOs and observed $T_{\parallel}$ signatures in the sub-slab and wedge mantle.

Several sub-slab and wedge mantle flow models have been proposed to aid the interpretation of possible correlations between seismological indicators and subduction parameters, including, availability of hydrous and melt phases (Holtzman et al., 2003; Faccenda et al., 2008), subduction zone obliquity (Mehl et al., 2003) and slab morphology (Kneller and Van Keken, 2007; Jadamec and Billen, 2010), among others. While these models may have local viability, they fail to predict global scale dynamics. Among the many hypotheses, the absolute ratio between trench and plate migration rates, V_{norm} ($|V_t|/V_P$), appears to better correlate with anisotropy patterns on a global scale (Long and Silver, 2008). More recently, an alternative model (Lynner and Long, 2014b) relating plate age and seismic anisotropy patterns has offered new insights. This model invokes the role of a thin decoupling layer underneath the old Pacific lithosphere in enhancing 3D returning flow, as opposed to coupling promoting 2D corner flow in the younger subducting lithosphere. Given that negligible to moderate correlation between plate age and occurrence of a ULVz emerges from our dataset, and age-related buoyancy variations have been treated in previous work resulting in little lateral flow with mainly sub-slab parallel flow in the mantle (Capitanio and Faccenda, 2012), we focus our attention to reproducing the model envisioned by Long and Silver (2008). Long and Silver envisioned a mechanism under which a sub-lithospheric thin decoupling “hidden” layer of entrained material beneath the subducting plate, along with a barrier to the flow at depth, were necessary to induce trench-parallel flow in the sub-slab and corner-flow in the wedge mantle. In this model, flow in the wedge is controlled by a competition between 2D flow induced by V_{norm} , whereas the sub-slab mantle regime can be predicted by absolute trench migration only.

Today seismic anisotropy signatures remain sparse and elusive, and the apparent lack of entrained flow requires a multidisciplinary effort to help explaining the nature of the observations. Yet, very little subduction modelling has been conducted on this subject.

6.4 Three-dimensional models of subduction and induced mantle flow

Previous large-scale 3D numerical work (Paczkowski et al., 2014b) on the relationship between prevailing direction of maximum strain and subduction-related parameters has shown that $T \parallel$ strain in the volume of mantle directly beneath the slab can be achieved through significant decoupling between the slab and the sub-slab mantle, consistent with the Long and Silver (2008) decoupling hypothesis. Such decoupling appears more likely to occur in deeper and steeper slabs than in shorter and shallower dipping ones. Another series of studies (Capitanio and Faccenda, 2012; Faccenda and Capitanio, 2012, 2013) has presented the development of a $T \perp$ due to simple shear at shallower depths, and a $T \parallel$ pattern behind the dipping slab controlled by pure shear deformation caused by slab rollback. Alternatively, variations in slab dips and morphologies have been shown to induce stretching in the direction parallel to trench locally in the mantle wedge (Kneller and Van Keken, 2008; Jadamec and Billen, 2010).

While progress has been made, some outcomes of these studies could be conjectures caused by model initial and boundary conditions rather than the result of time-evolving slab dynamics. These limitations could be related to the use of instantaneous-kinetic models with imposed pressure flow gradients, lateral geometry of the subducting oceanic plate introducing slab free-edge toroidal effects, as well as a combined velocity gradients approach to assess flow dynamics, as it is the geometry of finite strain, rather than flow velocity per se, that controls the splitting direction of anisotropy. Here, we re-evaluate the status of current literature and focus on reconciling models of seismic anisotropy with surface motions by using fully dynamic subduction numerical models where the oceanic lithosphere, OL, has the same lateral extent of the modelled box (3,000 km).

In our models, the ULVz is placed directly underneath the OL and is characterised by a viscosity reduction of two orders of magnitude compared to the surrounding mantle within a 30 km thick channel, consistent with observations (Naif et al., 2013; Hawley et al., 2016). As limited constraints currently exist on ULVz width, W_{ULVz} , we study end-member scenarios by varying the ratio between ULVz and OL widths from 1 to 0, in addition to also performing simulations that include an OPl. We calculate the prevailing direction of the maximum principal finite strain, θ_{MPFS} , (0 prolate – 1 oblate ellipsoids, and 0.5 shear deformation) of the flow field on a continuous scale, which can be used as a proxy of the anisotropy for most olivine crystals. We quantify the normalised predominant direction of the flow, θ' , of the absolute angle, θ_{MPFS} , as well as V_{norm} , where we conveniently refer to V_{norm}^* as the ratio between trench and plate migration rates (V_t/V_P) with trench retreat and advance referred to as $-V_t$ and $+V_t$, respectively. We ultimately estimate the strength of the anisotropy by calculating the the overall misalignment in the prevailing direction of maximum finite strain, γ , in the specific areas of interest.

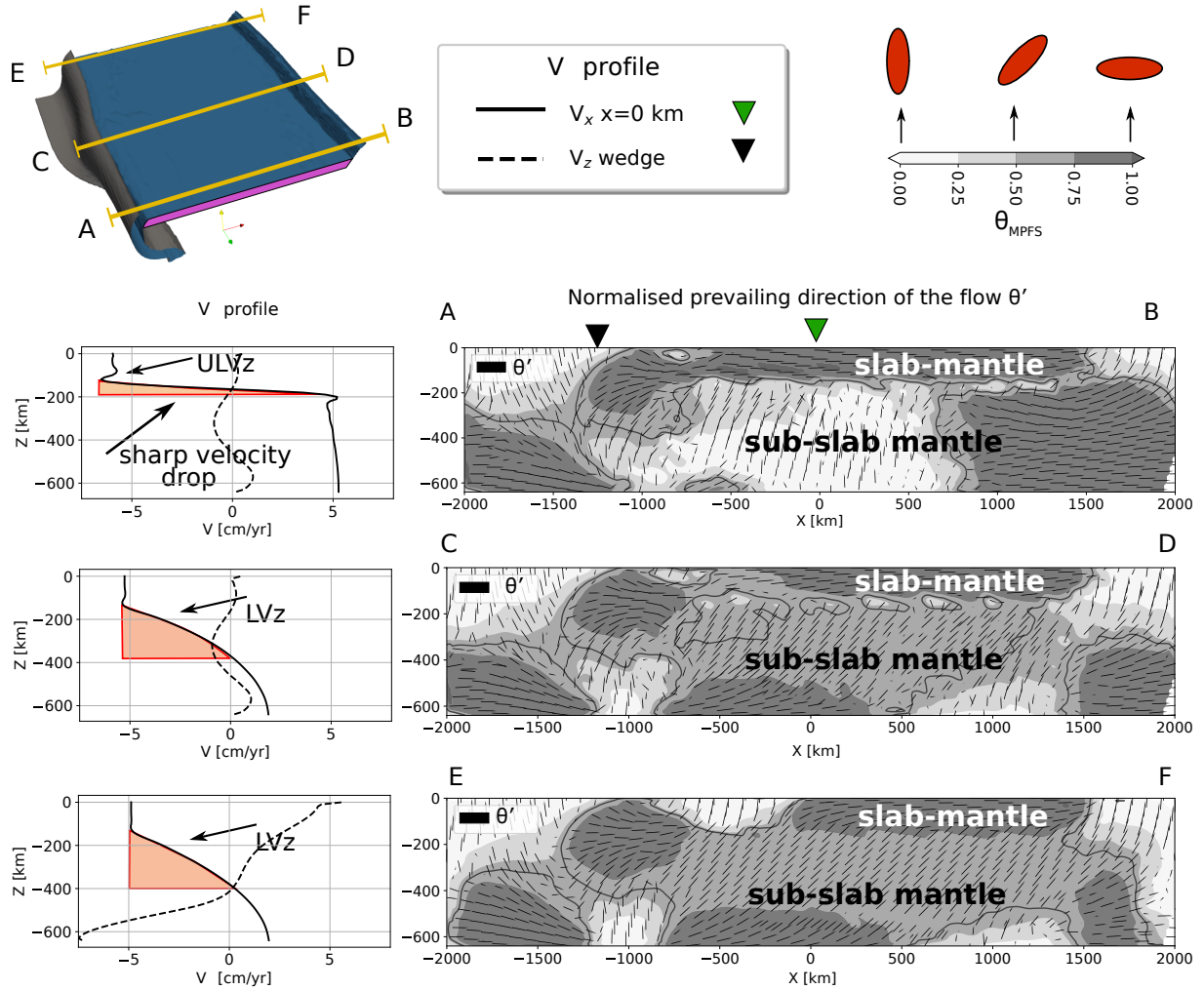


Figure 6.2: Summary of the outcome for i) the subduction style (top left); ii) patterns of prevailing direction of the flow field, θ_{MPFS} through a contour plot and normalised predominant direction of the flow, θ' , for the vector field within the colour map (right) in front (profile AB), mid (CD) and back (EF) vertical profiles; and iii) horizontal (solid) and vertical (dashed) velocity profiles of the centre (green triangle, solid line) and wedge (black triangle, dashed line) sections of the model, respectively (plots, left).

We find that two different subduction styles and mantle flow patterns can develop within the same subduction system due to variation in asthenosphere width.

The part of the subduction zone underlined by a decoupling asthenosphere exhibits a subduction style characterised by a slow-moving trench, a larger V_P and a steep and concave slab shape. The remaining part of the slab without a ULVz shows larger rates of trench retreat, lower V_P and a convex-slab shape (figure 6.2a top left). In the system, the nature of the mantle flow can be divided into two regimes with coherent prevailing flow direction depending on depth (figure 6.2). These are: i) the slab-mantle referring to the flow in the slab and the mantle directly underneath it (until approx. 270 km); and ii) the sub-slab mantle relating to deeper mantle flow.

We obtain a variety of mantle flow patterns. In the slab-mantle domain, we explain this variability as the vectorial difference between plate movement and asthenospheric mantle flow underneath it. In the portion of the model including a ULVz, large plate advance motion corresponds to plate-motion parallel flow where efficient compressional regime dominates within the plate (figure 6.2b). The discordant motion of the lithosphere and mantle blocks generates abundant shear stresses in the mantle directly beneath the base of the lithosphere to accommodate a velocity drop of ≈ 10 cm/yr in a 30 km thin interval (figure 6.2b, left). Such velocity drop produces a strain rate of $\approx 5 \times 10^{-14} \text{ s}^{-1}$ equivalent to ≈ 5 MPa for a $2.5 \times 10^{19} \text{ Pa}\cdot\text{s}^{-1}$ weak layer. It should be noted that if the thickness of this layer would be 10 km, it would translate to a strain rate of $\approx 3 \times 10^{-13} \text{ s}^{-1}$, consistent with the estimate of Stern et al. (2015) allowing melt localisation and abrupt seismic anomaly detection within this interval. Conversely, in the region not underlined by a ULVz, the velocity drop beneath the plate is half of that in the counterpart, and it is accommodated within a gradual transition ≈ 220 km resulting in strain rates of $\approx 7 \times 10^{-15} \text{ s}^{-1}$.

In the shallower slab-mantle domain of this portion of the model, the competition between larger trench migration and lower plate motion rates gives rise to the observed $T \parallel$ signature just behind the trench zone, whilst finite strain aligns $T \perp$ further away from the trench it. Being the modelled subducting system in a status of dynamic equilibrium, lower plate velocity corresponds to larger velocity in the wedge which causes an enhanced corner flow signature in the shallower fore-arc mantle (figure 6.2d). In our models, the addition of an OPl contributes to increasing trench curvature and enhancing strength of the corner flow in the fore-arc shallow mantle. A full list of case descriptions is available in §6.7.

We attribute the combined contributions of 3D lateral flow and vectorial difference between asthenosphere flow in the direction of plate motion and returning flow in the opposite direction to the sub-slab mantle case. When a ULVz is included in the models, the flow in the sub-lithospheric mantle is fully decoupled, thus it aligns in the counter direction to plate motion, $T \parallel$ (figure 6.3a). On the other hand, if a ULVz is not included,

then the sub-slab mantle flow acts as a passive agent for plate motion. Limited shear deformation in the direction of plate motion occurs in a thicker interval underneath the plates and returning mantle flow aligns in the opposite direction at further depth. The flow in this region is given by the vectorial difference in the asthenosphere flow and results in a slab dip parallel flow pattern (figure 6.3b).

The development of strong toroidal motion at the base of the lithosphere causes slab and trench asymmetric deformation throughout the subduction zone, which in turn forces 3D mantle flow to escape laterally and reorganise at depth. Lateral flow migration is more competitive where lithosphere and mantle are well coupled at depth as horizontal and vertical velocity gradients are inferior and toroidal motion becomes more effective. This is also the case in the wedge region including a ULVz beneath the slab, as stronger velocity gradients are located away from wedge (figure 6.3). Toroidal flow corresponds to vortex-like motion and is responsible for transform (strike-slip) motion or plate rotation at the surface. Such a strong toroidal component observed in our models could represent an effective mechanism for creating transform faults far away from the trench in the oceanic plate, as well as an efficient mechanism for describing volatiles and other phases transport and melt-mixing processes farther in the wedge region.

6.5 Discussion and conclusion

Consistent with the “hidden” anisotropy model proposed by Long and Silver (2008), we observe trench-parallel flow in the sub-slab mantle induced by a thin decoupling sub-lithospheric layer; however, we find that this mechanism necessitates significant convergence to allow LAB decoupling to remain effective. We also do not observe corner flow in the wedge. While the absence of an overriding plate represents a limitation of this study, this was a necessary choice to isolate ULVz contribution to flow dynamics. This choice has allowed us to predict flow alignment $T \perp$ in the slab-mantle and $T \parallel$ sub-slab mantle domains as effect of ULVz only.

In agreement with previous subduction numerical work (Paczkowski et al., 2014a), we were able to obtain a trench-parallel signature in the sub-slab mantle induced by decoupling at the LAB which is stronger for steeper dipping slabs. In addition to reproducing two distinct shallower and deeper mantle domains for the development of seismic anisotropy patterns (Faccenda and Capitanio, 2013), and slab-parallel signatures behind slabs of retreating trenches, also consistent with previous work (Capitanio and Faccenda, 2012). However, in our models miscellaneous patterns of deformation throughout the same subduction system emerge naturally without imposing external boundary conditions or inducing slab-free edge effects.

In our flow models, it is the amount of asthenosphere decoupling that controls relative plate motions and slab morphology, which in turns affects flow mantle capacity to

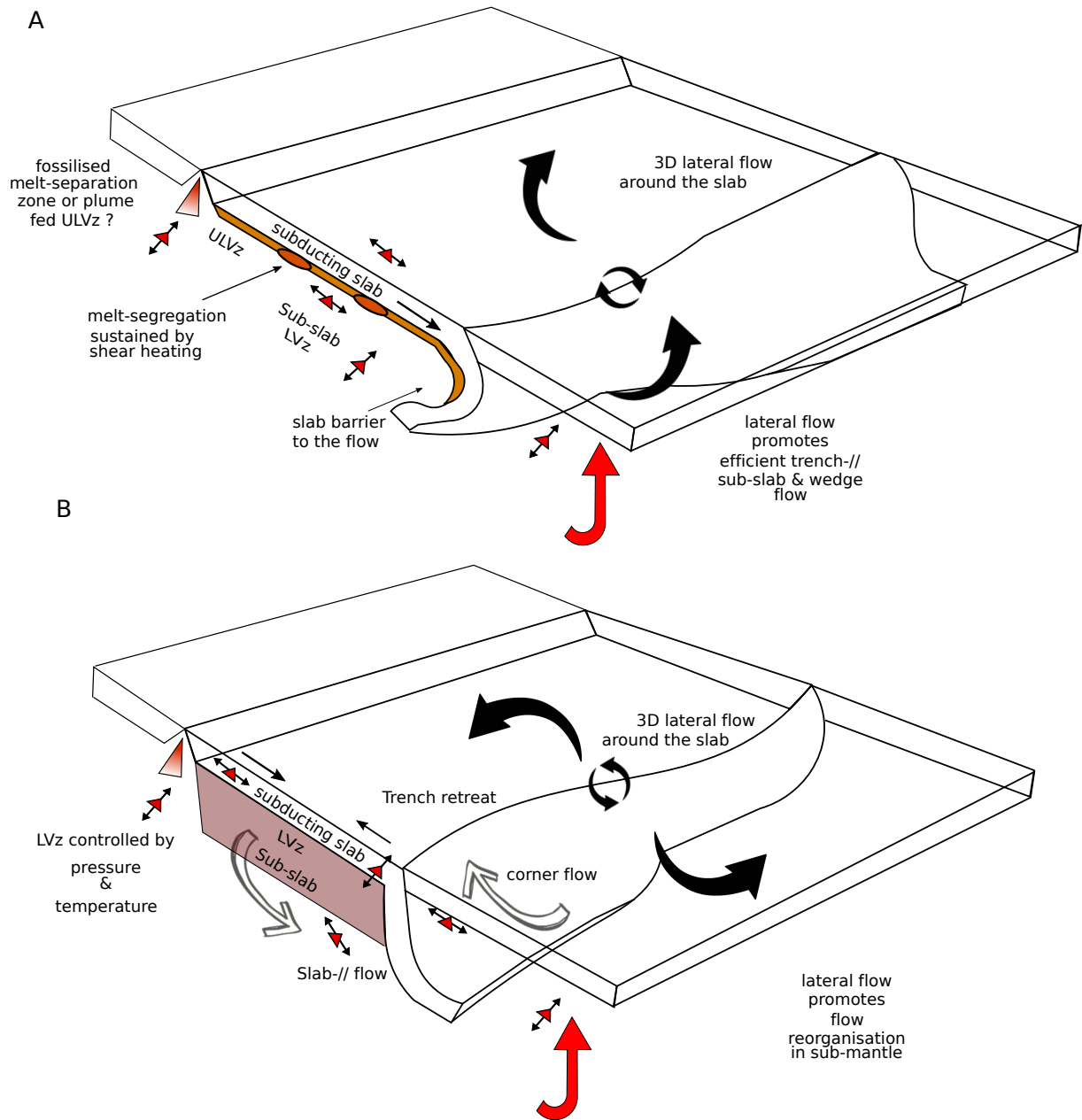


Figure 6.3: Schematic cartoon illustrating the combined influence of 3D lateral flow and shear stresses on patterns of seismic anisotropy (red triangles). In A and B, the 3D mantle flow evolution for the portions of the subduction model with and without an ultra-low-velocity zone (ULVz), respectively.

reorganise at depth; and more importantly, it is the lateral variability in asthenosphere decoupling that enables the generation of a variety of subduction regimes along different segments of the same subduction zone. As such, the combined use of seismic anisotropy patterns and strength of the anisotropy layer, together with V_{norm}^* can be used to predict the amount of lithosphere-mantle decoupling at depth, and vice versa.

Another important caveat of our study is that the slab-mantle flow regime scales negatively with V_{norm}^* . Such as that faster-moving trenches and lower plate velocities relate to larger γ values (figure 6.4a). Conversely, the deeper sub-slab mantle regime scales positively with V_{norm}^* , and lower migrating trenches and faster moving plates relate to a more coherent pattern of seismic anisotropy 6.4c)

The relative influence of both mechanisms is governed by the relative amount of asthenospheric decoupling. These results aim to highlight the critical importance of mantle survey depth for the assessment of mantle flow and seismic anisotropy patterns. Consistent with detailed observation compiled in our dataset (and plot in figure 6.4c), we find that both V_t and V_P exert a primary control on seismic anisotropy signatures. However, we find that combined used of V_{norm}^* and γ (which relates to δt from seismic observations) can be a sufficient proxy to determine amounts of mantle-lithosphere coupling.

There are some locations, such as Tonga, as well as Sumatra, that show very low V_{norm}^* and larger δt values. In these cases, it is plausible that mechanisms of local applicability, such as structural heterogeneities in the slab or overriding plate, among others, could provide additional contributions to the mechanism illustrated here. Yet, a ULVz has shown the overall effect of reducing increasing plate velocity and reducing trench retreat and better representing those subduction zones falling into a $V_T \pm 1/3 V_P$ range (Carluccio et al., 2019b), which includes the majority of subduction zone across different reference frames (Coltice et al., 2017), as shown in figure 6.4c. This mechanism could be particularly relevant in the Alaska, Norther New Zealand and Cascadia regions where miscellaneous patterns of seismic anisotropy, such as those presented in figure 6.2, are observed and where the presence of ULVz and/or OPl are confirmed by geophysical studies. However, this may also be relevant for other subduction zones (e.g., Mariana and South America) where V_{norm}^* and seismic anisotropy patterns predict the presence of a decoupling asthenosphere.

In an attempt to reconcile previous findings, we present a regional framework representing complexity of observations throughout different segments of a natural subduction zone where the mantle flow evolves dynamically underneath. This study deciphers the extent to which mantle-lithosphere processes can influence the structure, evolution and dynamic behaviour of subducting slabs into the deeper mantle on global scale. As such, our findings offer a novel perspective on the asthenosphere role in the evolution of complex convergent margins, for which little predictive previous numerical modelling exists. Thus,

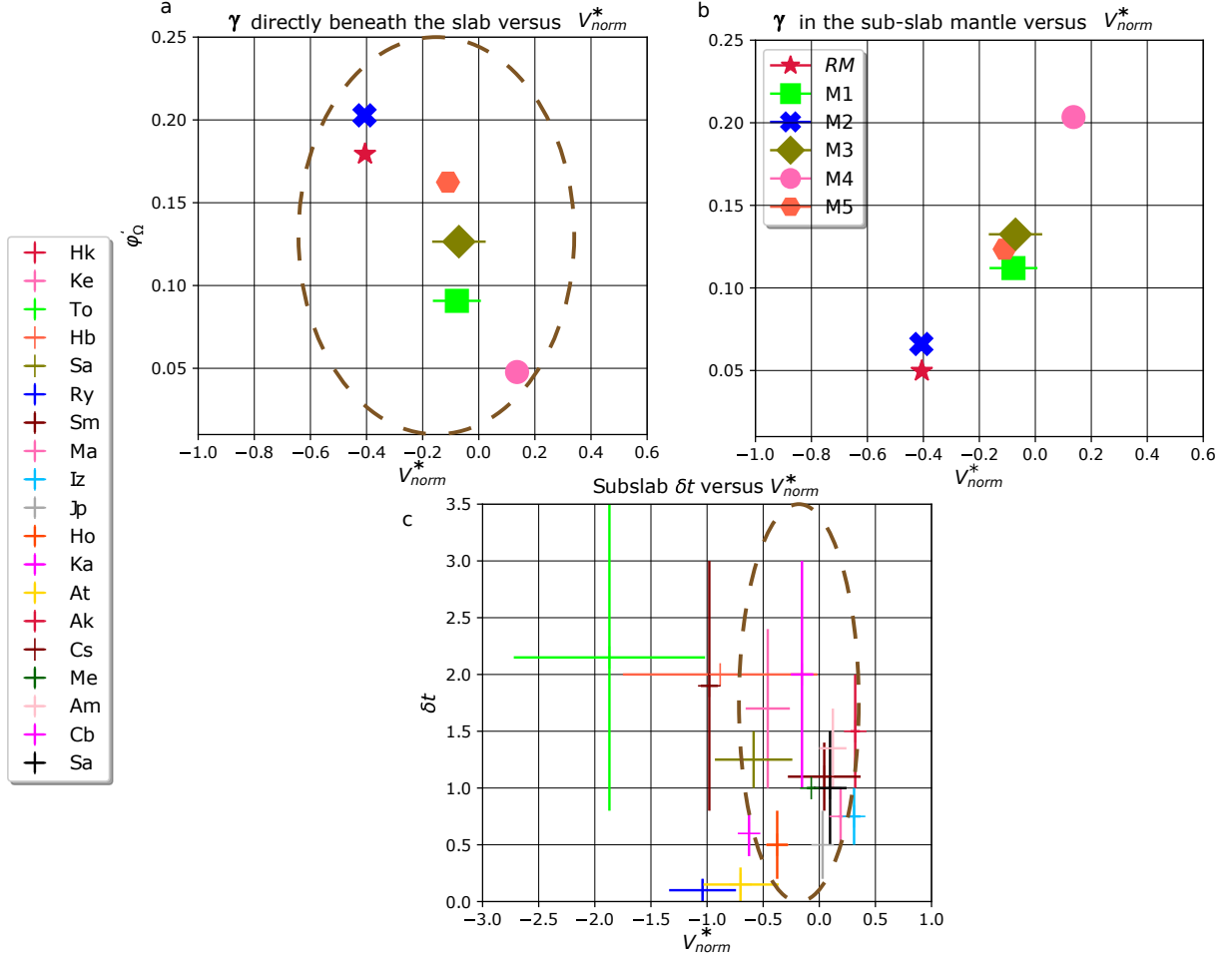


Figure 6.4: A comparison between model results of the overall misalignment in the prevailing direction of maximum finite strain, γ , of θ_{MPFS} , and observations of shear wave delay time for the sub-slab mantle for natural subduction zones. In a, model results show inverse dependence of γ for the ratio between plate and trench migration rates, V_{norm}^* in the slab-mantle regime; whereas, the results show a direct dependence γ - V_{norm}^* in the sub-slab mantle regime. In c, observations of delay time (δt) of shear waves splitting for the sub-slab mantle versus natural V_{norm}^* in natural subduction zone (in no-net reference frame). Our results could apply to those natural subduction zone falling within the dotted area where $V_T \pm 1/3 V_P$.

particularly relevant for the reinterpreting the geological record of many convergent margins, but also for the very understanding of inner workings of plate tectonics and forces behind plate motion.

6.6 Acknowledgements

The authors are grateful to J. Mansour and F. Capitanio for their constructive comments and suggestions. Users can access data from this paper by contacting the authors.

6.7 Supporting Information for “Asthenospheric dynamics during subduction”

Contents of this file

1. Method
2. Compilation of measurements
3. Additional results
4. Figures from F5 to F16

6.7.1 Method

The models are implemented in a three-dimensional Cartesian box with a size of 4'000*3'000*660 km (figure 6.5) and solve the conservation equations of mass and momentum using the numerical framework *underworld* (Moresi et al., 2003; Mansour et al., 2020). The model setup builds upon Carluccio et al. (2019b) to expand to three dimensions. The model setup includes an oceanic crust (OC) and lithosphere (OL) laterally homogeneous that extend across the entire modelled box, and additionally an OPl underneath one half of OL width. All layers are surrounded by mantle material (UM). The crustal thickness is 10 km, and the oceanic plateau and ULVz when included have a thickness of 25 km and 30 km, respectively. Previous numerical work (Carluccio et al., 2019b) has shown that the decoupling effect of a ULVz maintains a constant signature wherever a visco-elasto-plastic rheology age-dependent is implemented. The density contrast between the oceanic lithosphere and the underlying mantle is 80 kg/m³ (Cloos, 1993) and the OL-UM and ULVz-UM viscosity ratios are both 10², consistent with observation (Moresi and Gurnis, 1996; Hawley et al., 2016). Because we focused only on the influence of a ULVz and OPl on the strain accumulation in the mantle and behind the subducting slab, the overriding plate and other factors are not included. We model the evolution of a mature oceanic lithosphere (80 Myr) with a 30 km thick channel directly underneath the base of

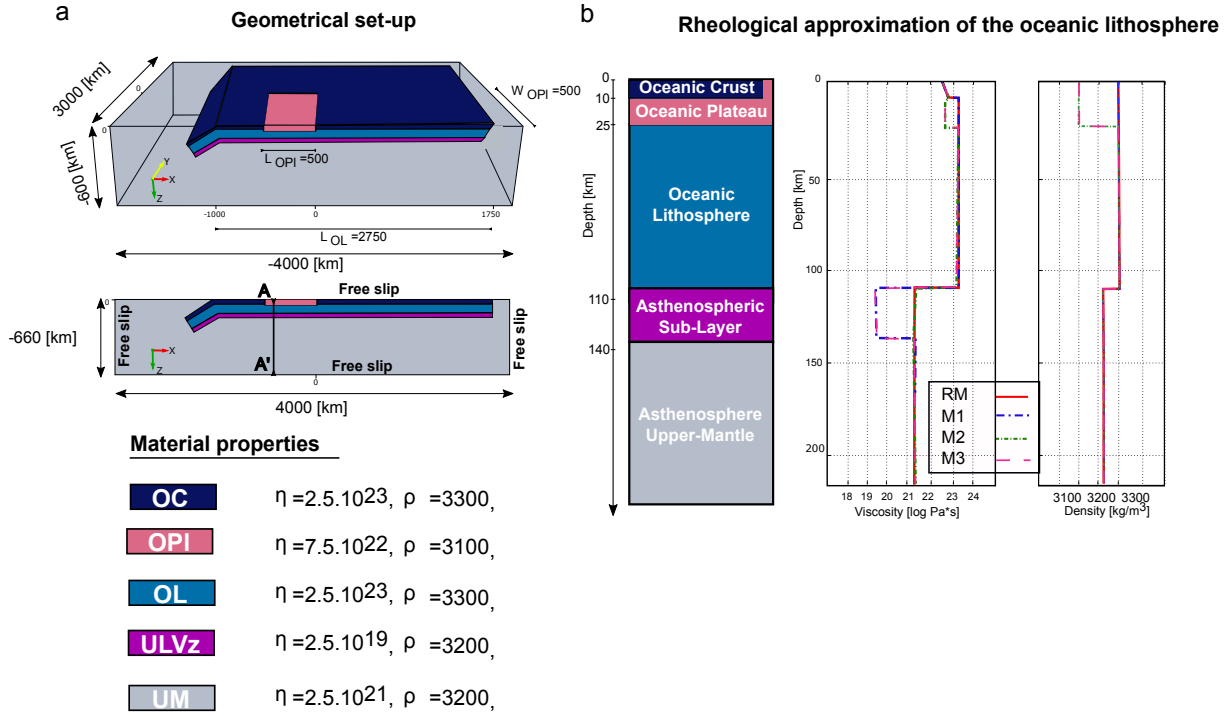


Figure 6.5: A schematic representation of the geometrical setup for the models with a weak layer (WL) and an oceanic plateau. In a, an edited sketch from Long and Silver (2008). In b, our model setup which builds upon the work of Carluccio et al. (2019b) and expands to three-dimensions. In c and d, a summary of the material field and initial rheological profiles through depth.

the plate. All layers modelled adopt Newtonian viscous formulation except for the visco-plastic rheology of the crust. The visco-plastic rheology of the crust allows the surface layer to detach freely from the surface and descend into the mantle underneath, such that subduction is dynamically initiated and self-sustained. Free-slip boundary conditions are enforced everywhere to minimize the influence of the box sidewalls and no other boundary conditions are applied.

As minimal constraints currently exist on ULVz width, W_{ULVz} , we explore different possibilities varying the ratio between ULVz and oceanic plate width from 1 to 0. In addition to also performing simulations that include an OPI.

To quantify the prevailing direction of the mantle flow, Our models evaluate the direction of maximum principal finite strain (MPFS) and compared it with the geometry of finite strain from seismic anisotropy studies. The evaluation of MPFS is carried through 2D surfaces by a post-processing reconstruction of the three-dimensional (3D) finite strain tensors from the velocity field at the desired time through the evolution of the models. The orientation of the max principal strain axes can be used as a proxy of the geometry of the anisotropy for most olivine crystal lattice preferred orientations. While the effects of inherited fabrics are not considered here, our approach represents a good approximation for regions with a large amount of accumulated strain.

We quantify the normalised predominant direction of the flow, θ' , of the absolute angle, θ_{MPFS} , (0 prolate – 1 oblate ellipsoids, and 0.5 refers to simple shear deformation) in the flow field, as well as V_{norm} , while we conveniently refer to V_{norm}^* as the ratio between trench and plate migration rates (V_t/V_P). We ultimately estimate the strength of the anisotropy calculating the overall misalignment in the specific areas of interest, γ , through a surface integral over the surface of interest. More information about how we calculate kinematics parameters can be found in previous work (Carluccio et al., 2019b).

To facilitate comparison between sub-slab anisotropy observations and numerical model predictions, we model a generic subduction system and, then, we compare it with available high quality uniformly processed shear wave splitting measurements. We compile seismic anisotropy parameters from studies that consistently constrain the anisotropy in the sub-slab and wedge mantle in several regions along the edges of the Pacific plate (e.g., Long and Silver, 2008; Long, 2013; Long and Wirth, 2013; Lynner and Long, 2014b). For each of these subduction zones, we gather information about the orientation of the fast wave, ϕ , known as the fast splitting direction, which contains information about the strain geometry, and the delay time, δt , between the time of arrival of the two waves which measures the strength or the thickness of the anisotropic layer. We collect the estimates on slabs dips and morphologies, wherever possible, from the global compilations of Lallemand et al. (2005) and Goes et al. (2017), respectively, and calculate V_{norm}^* in a no net rotation reference frame (NRM) (Schellart et al., 2008). Where more than one average value for either V_P or V_t is available for the same subduction system, we calculate the mean average including the standard deviation error. Full details on this compilation can be found in figure (figure 6.6).

6.7.2 Compilation of measurements of Shear Wave Splitting, Plate Kinematics, Slab Ages, Dips and Geometries, and LAB inferred parameters.

In this section of the literature review, we compile a series of geophysical observations. These observations include the recording of an abrupt drop in share wave velocity or increase in electrical conductivity underneath the base of the subducting lithosphere, patterns of seismic anisotropy in the sub-slab and wedge mantle, plate and trench kinematics, and slab age, dips and morphologies. We gather this information for 11 subduction systems i) Hikurangi- Kermadec- Tonga; ii) New Hebrides, San Cristobal, New Britain, Sanighe; iii) Nankai-Ryukyu; iv) Sumatra and Sunda; v) Mariana and Izu-Bonin; vi) Hokkaido, Kuril, Kamchatka; vii) Aleutian-Alaska; viii) Cascadia; ix) Mexico; x) Central America, Colombia, Caribbean; and xi) South America.

The kinematics of these systems are described accordingly to the no-net reference (NR) frame (Schellart et al., 2008), where NR is usually more suitable for comparison with numerical results in cartesian coordinates (Tackley, 2000). Nonetheless, in our outcomes

Subduction segments	subduction zone ID	UVZ	Opl	slab sketch	Vt mean	Vc mean	Vnorm	Wedge δt (s) mean	Wedge ϕ	Subslab δt (s) mean	Subslab ϕ
Hikurangi,	Hk	11	HP		1.40	4.70	0.298	0.340.2	Trench- to trench-⊥	~1.5±0.5	Trench-
Kermadec	Ke	1, 2, 3			-3.90	6.10	-0.644	1.0±0.4	Trench-	~1.7±0.7	Trench-
Tonga	To		MP		-13.10	6.10	-2.139	1.2 ± 0.3	Trench- to trench-⊥	~2.15±1.35	Trench-
New Hebrides, San Cristobal, New Britain	Hb, Cr, Br	3, 4	OJ, NB*		-4.53	4.80	-0.943			~2.0	Trench-
Sangihe	Sa	2			-1.70	7.10	-0.484			1.25±0.25	**Trench-
Ryukyu	Ry	4, 6, 7, 12			-4.15	4.70	-0.883	0.8 ± 0.5	Trench-	0.140.1	**Trench-⊥
Sumatra, Sunda	Sm				0.00	6.80	~0.	0.340.1	Trench-	1.140.30	**variable, trench-⊥, Trench-
Marina	Ma	2, 4, 5, 6, 7	EM, PF		1.37	7.10	0.310	0.6±0.3	Trench-	~0.75±0.25	Trench-
Izu-Bonin	Iz				2.20	7.10	0.310	1.4 ± 0.4	Trench-	~0.75±0.25	Trench- , some oblique; splitting patterns complex
Japan	Jp	4, 6, 7	SR		0.20	7.10	0.028	0.2±0.1	Trench- to trench-⊥	~0.540.3	Oblique, some trench- ; splitting patterns complex
Hokkaido, Kuril	Ho				-2.40	6.40	-0.375	0.7±0.3	Variable		Trench-
Kamchatka	Ka	3, 4			-4.00	6.40	0.625	0.4±0.2	Trench-⊥ to trench-	~0.640.2	**Trench-⊥, complex
Aleutian	At				-2.30	4.20	-0.548	1.5±0.4	Trench- or oblique	~0.15±0.15	**Trench-⊥
Alaska	Al	4			-1.70	6.10	-0.279	0.9±0.4	Trench-⊥ to trench-		**trench-⊥, complex
Cascadia	Cs	3, 8, 9			-1.90	6.10	-0.311	*1.540.5	Trench-⊥ to trench- , oblique	~1.9±1.1	Trench-perp (SKS) complex (source-side)
Mexico	Me				-1.00	9.10	-0.110	0.2 ± 0.1	Trench- to trench-⊥	~1.0	Trench-⊥
Central America, Colombia	Am-Co	3, 10	CP		0.950	9.10	0.104	0.3±0.3	Trench-	~1.35±0.35	**Trench-⊥
Caribbean	Cb				-1.40	9.10	-0.154	0.27±0.03	Trench-	~2.0±1.0	Trench-
South America	Sa				0.567	6.45	0.088	0.3±0.2	Trench-	~1.0±0.5	***Trench- to trench-⊥

Figure 6.6: A compilation of measurements of shear wave splitting, plate kinematics, slab age, dip and geometry, and LAB inferred parameters. The first and second columns illustrate names and acronyms of subduction regions of interest in this study: Hk – Hikurangi; Ke – Kermadec; To-Tonga; Hb- New Hebrides; Cr- San Cristobal; Br- New Britain; Mn- Manila; Sa- Sanughe, Sm- Sumatra & Sunda, Ry- Ryukyu; Na- Nankai; Mr- Mariana; Iz- Izu-Bonin; Jp -Japan; Ku-Kuril; Ka-Kamchatka; At- Aleutian, Ak- Alaska; Cs- Cascadia; Am- Central America ;Co-Colombia; Sa- South America. The third column provides reference and location of SLL observation at the LAB. Reference list includes: 1: Gaherty et al. (1996), 2: Revenaugh and Jordan (1991), 3: Schmerr (2012), 4: Bagley and Revenaugh (2008), 5: Gaherty et al. (1999), 6: Kawakatsu et al. (2009), 7: Bagley et al. (2009), 8: Rychert et al. (2018), 9: Hawley et al. (2016), 10: Naif et al. (2013), 11: Stern et al. (2015) and 12: Wang et al. (2020). The fourth column presents locations of current oceanic plateaus found in the vicinity of Pacific subduction zones: HP: Hikurangi; MP: Manihiki; OJ- Ontong Java, NB: Nauru Basin; EM- East Mariana Basin; PF: Pigafetta; SR- Shatsky Rise, HR- Hess Rise, MG-Magellan Rise, CP-Caribbean following the description of Kerr (2014). Column fifth to seven provide information about the mean of trench migration, plate convergence and their ratio averaged across the subduction zone region utilizing data from Schellart et al. (2008) global compilation in the no-net reference frame. The last four column provide details about mean shear wave splitting delay time (δt) and fast splitting direction (ϕ) in the wedge and sub-slab mantle where available. Unless specifically noted otherwise shear wave splitting come from the review paper of Long (2013). These data have been updated to take into account recent development in the observations for Cs* wedge (Long, 2016), Ry, Na, Sm, Ka, At, Ak and Am** Lynner and Long (2014b) and S*** Long et al. (2016) mantle flows.

the average absolute trench velocity is approximately 30% or lower than the total convergence plate velocity (figure 6.4), as found along many of the aforementioned subduction segments and consistent across several plate motion reference frames (e.g., Goscombe and Gray, 2008; Schellart et al., 2008; Becker and Faccenna, 2009).

Average deep slab dips values and slab morphologies have been compiled from the work of Lallemand et al. (2005); Hayes et al. (2018) and Goes et al. (2017), respectively, and we neglect shallow crustal processes. We divide slab dips and age in categories according to the classification of Lallemand et al. (2005) Müller et al. (2019), respectively. Deep slab dips are classified as: flat ($\text{dip} < 30^\circ$), shallow ($30^\circ < \text{dip} < 50^\circ$), moderate ($50^\circ < \text{dip} < 60^\circ$), steep ($60^\circ < \text{dip} < 70^\circ$), and very steep ($\text{dip} > 70^\circ$). Plate and seafloor ages are classified in Millions of years (Myr): i) Quaternary (present -2.58); ii) Neogene (2.59-23.0); iii) Paleogene (23.0-66.0); iv) Cretaceous (66.0-145); v) Jurassic (145-201); vi) Triassic (201-252); and vii) Permian (252-299). While, there are inevitably uncertainties associated with the data presented in figure 6.6 and in the description below in terms of plate velocities, seismic anisotropy, slab morphologies, dips and ages, for the present-day slab observations, these are supposed to be relatively small and local, in general, less than 5–10% (Schellart, 2020), and will therefore have no discernible impact on the global data distribution as shown in 6.1. There may be larger uncertainties in some local and regional trends, however, here we describe global trends and focus on highlighting larger variations in wave splitting, kinetics, slab shapes, dips and ages within the same subduction system, and in comparison to others. There are possibly larger uncertainties associated with subduction zone age. However, discussing in deep detail the topic of plate age is beyond the scope of this work and these values are here reported only for the sake of completeness. Below, each of these subduction zones is briefly described and a justification for the evidence on a ULVz or OPl is provided.

- (i) The Hikurangi- Kermadec-Tonga subduction system is characterised by the subduction of the Pacific plate underneath the Australian plate. The average Pacific plate convergence velocity is 6.1cm/yr and trench migration rates vary from -16.6 (retreat) in north Tonga to 1.4cm/yr (advance) in the Hikurangi region. The subducting plate dip steepens southwards from Tonga-Kermadec ($50\text{--}60^\circ$) to Hikurangi ($60\text{--}70^\circ$), and the average oceanic plate age is cretaceous and increases southward (roughly from 80 to 160 Myr). The minimum age of the subduction zone is generally based on the oldest ages of fore-arc magmatism in the Tonga arc (49– 51 Myr) (Meffre et al., 2012; Schellart, 2020). More recently a younger age has been suggested (~ 30 Myr) (van de Lagemaat et al., 2018); however, other reconstructions show a subduction zone that was already active at 90 Myr (Schellart et al., 2006). Seismic anisotropy studies (e.g., Long, 2009, 2013) report a widespread trench-parallel flow average fast splitting direction beneath the sub-slab of this subduction system. A

trench-perpendicular to parallel anisotropy signature is reported within the wedge mantle of Tonga and Hikurangi segments, whereas a prevailing trench-parallel trend is displayed across the wedge in Kermadec (Long and Wirth, 2013). The stronger shear wave splitting delay time for the sub-slab and wedge mantle across this compilation of studies are registered in the Tonga region. A range of geophysical studies (i.e. Revenaugh and Jordan (1991); Gaherty et al. (1996); Schmerr (2012); Stern et al. (2015) suggests the presence of a thin layer of reduced viscosity, being enriched in melt, or hydrated phases or volatiles and ponding at the base of the subducting lithosphere. Stern et al. (2015) use explosion-generated seismic waves to obtain seismic reflection images of the base of the 120 Myr old Pacific plate and Hikurangi Plateau subducting westward beneath continental New Zealand. They registered a P-wave decrease of at the least $8\pm3\%$ confined into a 10 km channel descending parallel to the base of the plate into the mantle. They predict that this layer can accommodate a strain drop of approximately 10^{-13} and a finite strain of 10 in 1 Myr. Also, Revenaugh and Jordan (1991); Schmerr (2012) and Gaherty et al. (1996) report the presence of a ULVz in several zones in the Pacific including Tonga using seismic surface waves, ScS reverberation and Snt, SS, and SSS body waves techniques, respectively. Schmerr (2012) and Revenaugh and Jordan (1991) detect an average reflection coefficient drop of about 3.5%, including values up to 10.3% and negative SS precursors ($>\sim 6\%$) compatible with the presence of a layer up to 30 km thick underlying the base of the Pacific lithosphere where it subducts underneath Tonga. Revenaugh and Jordan (1991) maintain that the G discontinuity could be best explained as the seismic expression of the spinel – > garnet facies boundary. Additionally, strong corner flow in the subducting slab (e.g., Ribe, 1989) could affect the upward migration of melt and resultant anisotropy due to compositional layering and further accentuate the strength of the reflector. Moreover, both Revenaugh and Jordan (1991) and Gaherty et al. (1996) report evidence of a Tonga-Hawaii channelized corridor bounded at 68 km by a sharp G discontinuity. They invoke the role of superficial melt generation during crustal formation as being responsible for the rapid increase in hydrated phases content marking a fossilised base of the LAB at spreading centres. In the proximity of north Tonga and the North Island of New Zealand subduction zones are located the remains of the Manihiki and Hikurangi plateaux, respectively. A series of studies (e.g., Gladczenko et al., 1997; Taylor, 2006; Davy et al., 2008) suggest that the emplacement of the Manihiki–Hikurangi–Ontong Java plateaux occurred during a large igneous province (LIP) event in the Cretaceous (120-124 Myr), which later rifted apart at a triple junction. These studies relate the origin of these plateaux to the large Greater Ontong Java event, during which $\sim 1\%$ of the Earth’s surface was covered with volcanism, possibly due to a thermo-chemical superplume/dome that stalled at

the transition zone. The subsequent alkalic volcanism on the Hikurangi Plateau and the Zealandia micro-continent has been proposed to derive from a second large-scale volcanic event which may have triggered the final breakup stage of Gondwana, and thus the separation of Zealandia fragments from West Antarctica (Hoernle et al., 2010).

- (ii) The New Hebrides, San Cristobal, New Britain, Sanighe system is characterised by the interaction between the Australian and the Pacific tectonic plates. The Pacific and Australian plates move at an average speed of 7.1 and 6.4 cm/yr, respectively, and produce steep dipping slabs in the New Hebrides. Trench retreat migration rates range from -0.1 to 8.4cm/yr and exhibit significant variability in speed within the same subduction region. This region is generally considered young and only formed in the Neogene $\sim$ 10-15 Myr (Hall, 2002; Schellart et al., 2006; Richards et al., 2011; Schellart, 2020). Seafloor ages span from Cretaceous to Jurassic, on average, for the old Pacific plate, however, younger ages, Paleogene to Quaternary, are also seen close to spreading centres (e.g. twin active spreading ridges in the Northern Fiji Basin), which are likely former triple junctions (Auzende et al., 1994). A range of geophysical observations (e.g., Revenaugh and Jordan, 1991; Bagley and Revenaugh, 2008; Schmerr, 2012) suggests the presence of a channelized or ponded LAB underneath some of the segments of this subduction system. For example, Bagley and Revenaugh (2008) registered shear wave impedance discontinuities for eight paths across the Circum-Pacific subduction zones including this region. Their results reveal an impedance drop of approximately 8-10% within a 30 km interval at the base of the Pacific plate.

This subduction system is also well known for hosting a high level of tectonic fragmentation, the remains of many allochthonous terrains, as well as some of the strongest and most frequent earthquakes we register on Earth. Among the terrains present in this region, there is the Ontong-Java plateau (OJP) which is currently the largest plateau on Earth. It is located in the vicinity of New Britain island, and given the proximity and similarity to the Nauru flood basalts, it is proposed to share a similar origin and belong to the same LIP. Mochizuki et al. (2005) suggest that the Mesozoic Nauru Basin in the Western Pacific was created by seafloor spreading in Late Jurassic and Early Cretaceous and, subsequently, affected by Early Cretaceous flood volcanism approximately contemporaneous with emplacement of the nearby 120 Myr Ontong-Java Plateau (OJP). The current estimates of average crustal thickness for the OJP range from 39-32 km to 25 km (Furumoto et al., 1976; Sandwell and Renkin, 1988; Kerr, 2014).

- (iii) The Nankai-Ryukyu system extends from Taiwan to SW Japan, including north-east China. In this region, the slow recurrence of strong earthquakes, the frequency

of slow slip and small repeating earthquakes suggest that overall plate and mantle should be well-coupled at depth. In this region, the Philippine Sea plate is subducting beneath the Eurasian plate at an average rate of 4.7cm/yr with trench migration rates of the order ~ 6.3 -1.5cm/yr decreasing northward. The Philippine Sea plate age spans from Cretaceous to Quaternary and it becomes older moving eastwards away from the spreading centres. Deep slab dips range from shallow to moderate. This region is proposed to have been formed in the Miocene (~ 18 Myr) through subduction of the Philippine Sea plate below Eurasia (Lallemand et al., 2001). Seismic anisotropy studies (Long, 2013; Lynner and Long, 2014b) report a dominant trend of trench-perpendicular and trench-parallel flows for the sub-slab and the wedge respectively. In this subduction zone evidence of channelized LAB arise from the studies Gaherty et al. (1999); Bagley and Revenaugh (2008) and Wang et al. (2020). Gaherty et al. (1999) explore the characteristics of the G discontinuity by combining two data sets. The first data set includes the travel times and impedance contrasts across mantle discontinuities using ScS-reflectivity profiles. The second compiles observations of frequency-dependent travel times of three-component (S, sS, SS, sSS, SSS, Sa) and surface (R1, G1) waves, which constrain the anisotropic velocity structure between discontinuities. Bagley and Revenaugh (2008) and Gaherty et al. (1999) describe the G as a sharp discontinuity significantly deeper beneath the western younger Philippine Sea plate than in the older counterpart (located at 89 and 68 km), respectively, implying that the thermal cooling model alone does not control the thickness of the lid as it would require a thinner plate with decreasing age. Wang et al. (2020) use the P-wave reflection function technique and detects a layer of entrained material beneath the subducting plate. Wang et al. (2020) suggest that this material could be due to partial melting of the sub-lithospheric material with additional melt-induced and high water contents forming at a depth between 410 and 660 km.

- (iv) The Sumatra-Sunda system is very large (~ 7850 km) and is characterized by the subduction of the Australian plate beneath the Sundaland plate (Schellart et al., 2007; Schellart, 2020). The Sundaland plate includes Burma- Andaman segment, the Sumatra-Java-Bali-Nusa segment and the Timor-Banda segments. The Australian plate convergent velocity is currently 6.8cm/yr. The extreme ends of the subduction zones are in retreat while the inner part is in advance. The trench shows high curvature and slow migration rates ranging from -1.9 to 2.5 cm/yr. Beneath the Java arc, the subduction zone is steeply dipping ($\sim 60^\circ$) whereas underneath the Sumatra arc it exhibits shallower dips (30 - 45°), (Kundu and Gahalaut, 2011). Seafloor ages on average span from Jurassic to Paleogene, but younger ages can be found in the central area of the subduction zone (Heuret and Lallemand,

2005). Subduction has been proposed to have first initiated in the North Andaman segment roughly 94 Myr ago (Sarma et al., 2010; Advokaat et al., 2018), followed by the central part of the Sunda subduction zone (Sumatra-Java-Bali-Nusa) likely at $\sim 50\text{--}45$ Myr (Hall, 2012), and ultimately in the easternmost segment (Banda segment) ~ 15 Myr (Spakman and Hall, 2010). Seismic waves splitting studies (Lynner and Long, 2014b,a) have recently shown more complex anisotropy patterns for the sub-slab in this region. The sub-slab mantle flow appears trench-parallel in the southeast, plate-motion-parallel in the centre and ambiguous and more scattered in the northern part. A trench-parallel anisotropy signature is reported for the wedge throughout this system (Long, 2009).

- (v) The Mariana-Izu-Bonin system extends from near Guam to Tokyo, Japan and is an outstanding example of an intra-oceanic convergent margin (IOCM). The average convergence and trench advance rates throughout this subduction system are $\sim 7.1\text{cm/yr}$ and 2.5cm/yr , respectively. The trench shows a high curvature over a smaller length than in the Sm-Sn subduction system. Slab dip varies along strike, as well as the mode of interaction with the transition zone - from moderately dipping and failing to penetrate the 660 km discontinuity in the north to plunging very steeply and almost vertically into the deep mantle in the south (Klemperer, 2003). Pacific plate age estimates vary from Jurassic to Cretaceous and are generally younger in the Izu-Bonin than in the Marian segments. The Mariana -Izu-Bonin system was likely formed in the Paleogene, about 52 Myr (Ishizuka et al., 2011; Schellart, 2020) and subducts the oldest seafloor on the planet. Seismic anisotropy shows a trench-parallel trend for the sub-slab and wedge (Long, 2013; Lynner and Long, 2014b). In the vicinity of this subduction system are the remains of two oceanic plateaux: the Pigafetta and East Mariana basins. Data from basalts recovered in Pigafetta and East Mariana basins differs by less than 10 Myr (Castillo et al., 1994) and given the proximity and similarity to the Nauru basalts, it is suggested an origin related to the same LIP event (e.g., Castillo et al., 1994). In this system, evidence of a ULVz comes from multiple sources (Revenaugh and Jordan, 1991; Bagley et al., 2009; Kawakatsu et al., 2009). Kawakatsu et al. (2009) performed receiver function analysis of borehole broadband ocean bottom seismic data recording a 7-8% shear wave velocity drop, which translates into a compatible melt fraction of about ~ 3.5 to 4.0% that allows for a viscosity jump even greater than 10^3 at the base of both the Pacific and Philippines Sea plates. The layered model for the asthenosphere is the authors' best-fit explanation for the large viscosity drop inferred at the LAB. Similar outcomes were found by Bagley et al. (2009). Bagley et al. (2009) through re-examining the regions sampled by (Obayashi et al., 2006) with multiple ScS reverberations and SH-polarized shear phases. They report a

velocity drop at an interval up to 50 km of, possibly enriched in partial melt at the LAB. They attributed the melt-rich layer to the combined effects of water and temperature, and estimate a maximum temperature anomaly of 155°C with a resultant maximum olivine water content after melting of 0.100wt.%, which is thought to have lowered viscosity near the base of the slab.

- (vi) The Hokkaido-Kuril-Kamchatka system extends from the Japanese trench in the southwest to the triple junction near the Commander Islands in the northwest. The Kuril arc and trench links the Russian Kamchatka peninsula with Hokkaido, Japan and connects the Aleutian with the Japanese arc-trench. The Pacific plate subducts westwards beneath the Okhotsk microplate at an average rate of 6.4cm/yr and trench migration rates increase northward approximately from 0.2 to -4.0cm/yr. Slab deep dips span from shallow to moderate from North Kuril, South Kuril and Japan, and steepen southwards. This system was formed by the collision of the Okhotsk terrain to Eurasia, which is thought to have occurred in late cretaceous-early Paleogene at approximately 60 Myr, and it likely joined at a later stage (~ 25 Myr) with the Izu-Bonin-Mariana region (Schellart et al., 2003; De Grave et al., 2016). Sub-slab and wedge seismic anisotropy studies report variable trends for the Hokkaido, Kuril and Kamchatka subduction segments. Sub-slab anisotropy signatures range from trench-parallel in the southeast, plate-motion-parallel in the north to highly scattered and ambiguous in Kamchatka (Lynner and Long, 2014b). Wedge mantle trend spans from trench-parallel to trench-perpendicular (from the fore-arc to the back-arc) in the southeast, complex and variable in North Kuril and Okhotsk plate to trench-perpendicular to trench-parallel in the northwest of the subduction system (Long, 2013). Evidence of a ULVz in this area comes from the geophysical studies of Bagley and Revenaugh (2008); Bagley et al. (2009); Kawakatsu et al. (2009), whose techniques and results have already been described above. The Shatsky Rise and Hess Rise are two oceanic plateaux located in the vicinity of this subduction system. They formed during the spreading between the Pacific and Farallon plates between 140 and 100 Myr (Verzhbitskii et al., 2006) and appear related to plume tectonics. Geological and petrological studies report evidence of accreted fragments of oceanic plateaux currently present within the Japanese islands (Isozaki et al., 1990; Ishiwatari, 1991; Ichiyama et al., 2008, 2014).
- (vii) The volcanic arcs associated with the subduction of the Pacific plate beneath the North American plate in the Aleutian-Alaska region defines the most tectonically active region in North America. The Pacific North Artic plates move at different rates, 6.1cm/yr and 2.3cm/yr, respectively. The Alaska region is characterised by flat subduction and slab dips that steepen towards the westernmost portion of the

Aleutians ($\sim 40^{circ}$). Seafloor ages also increase westwards from Paleogenic to Cretaceous. Beneath the Alaskan margin subduction has been active since the Triassic, however the current Aleutian-Alaska subduction zone was formed more recently in the Early Cenozoic. The Aleutians segment likely formed in an intra-oceanic setting and at that time it was limited in the west by the Shirshov Ridge, after which the subduction zone grew westward to form a cusp with the Kuril-Kamchatka arc (Scholl et al., 1986; Schellart et al., 2003; Heuret and Lallemand, 2005; Jicha et al., 2006). Seafloor ages are from the Paleogene period on average. Seismic anisotropy studies report evidence of average fast shear wave splitting direction parallel to plate motion in the sub-slab Lynner and Long (2014b), however, the data are commonly noisy and sparse for the Alaska region Long and Wirth (2013). Mantle wedge anisotropy patterns are even more variable and complex in the Aleutian region, and show a trench-parallel to trench-perpendicular trend, from the fore-arc to the back-arc, of the Alaska segment. Schmerr (2012) and Bagley and Revenaugh (2008) report the presence of a layered and sharp G discontinuity through the Aleutian-Alaska subduction system.

- (viii) The Cascadia subduction zone is characterised by the subduction of the Juan the Fuca plate at an average speed of 6.1cm/yr underneath the North American plate. The North American plate velocity diminishes southward from 2.1cm/yr to 1.8cm/yr, whilst the trench retreat with an average speed of about 1.9cm/yr. The subduction angle also steepens southward from shallow to moderate values. The distribution of earthquakes, volcanism and regional horizontal stress also vary significantly throughout the subduction system. The average seafloor ages are Neogenic with a Quaternary ridge located in the northern part of the oceanic plate. During part of the Mesozoic and the Early Cenozoic is generally agreed that the Juan the Fuca plate was once part of the large Farallon plate (Burchfiel and Davis, 1975; Gordon and Jurdy, 1986; DeCelles, 2004) later subducted. However, other plate reconstruction studies favour older ages, such as Triassic or Jurassic (160–210 Myr) for the formation of this zone (e.g., Seton et al., 2012). These report that the Farallon slab underwent fragmentation only significantly after the inception of Farallon flat slab formation, ~ 45 -30 Myr, and due to the formation of slab tears, hence resulting in the creation of the small Farallon/Juan de Fuca and the Nazca-Cocos plates (e.g., Henderson et al., 1984; Gordon and Jurdy, 1986; Schellart et al., 2010; Liu and Stegman, 2012; Copeland et al., 2017). Seismic anisotropy studies record curved to trench-perpendicular fast waves splitting direction in the sub-slab (Long, 2016). Geophysical studies (e.g. Schmerr, 2012; Hawley et al., 2016; Rychert et al., 2018) among others, report the presence of a sharp base of the Juan de Fuca plate. These studies suggest viscosity reduction up to 10^4 compared to the neighbouring mantle

within an interval of ~ 10 km thickness. Hawley et al. (2016) and Rychert et al. (2018) use tomography and scattered wave (S-to-P receiver functions) techniques, respectively, and image a buoyant asthenosphere accumulating beneath the subducting Juan de Fuca plate and a sharp LAB discontinuity beneath newly formed lithosphere at the Juan de Fuca and Gorda Ridges. Both models require partial to explain the recorded velocity drop, which would decrease mantle viscosity and define the base of the young oceanic plate.

- (ix) The Mexico subduction zone is characterized by the flat subduction of the Cocos plate beneath the Caribbean plate in the North American plate. The Cocos plate and the trench move at average speeds of 9.1cm/yr and 1.0cm/yr, respectively. The seafloor age of the Cocos plate in the Mexican northern area is Neogenic on average, whereas seafloor ages in the Gulf of Mexico are older and mainly Jurassic. The Mexican flat slab formed at approximately 30–25 Myr (Ferrari et al., 1999; Morán-Zenteno et al., 1999; Kim et al., 2010), and the Cocos spreading ridge separating the Cocos plate from the Nazca plate, formed only at ~ 23 Myr (Lonsdale, 2005). Seismic anisotropy studies report a trench-perpendicular trend in the sub-slab, and a trench-parallel to trench-perpendicular to signature in the wedge, from the fore-arc to the back-arc (Long, 2013). In the vicinity of the Cocos ridge the G appears as a sharp discontinuity (Schmerr, 2012), yet the evidence of a strong G reflector located more closely to this subduction system has not yet been reported.
- (x) The Central America-Colombia subduction system is characterized by the subduction of the Cocos plate at an average speed of 9.1cm/yr beneath the Caribbean plate. In this region, the trench migrates slowly. Specifically, trench retreat rates span from -0.3cm/yr in the north, 0.0 cm/yr in the south segment, and advance in the middle part ~ 2.2 cm/yr of this region. Observations report slab fragmentation (Protti et al., 1994; Pedraza Garcia et al., 2007) and deep dips that range from very steep ($\sim 80^\circ$) in the northern western part to shallow in Colombia. Seafloor ages for the Cocos plate vary from Neogenic to Paleogenic, whereas in the Caribbean plate ages are on older and Cretaceous on average; however, younger ages (Paleogenic to Neogenic) are also present in the Cayman Islands. Recent work (Boschman et al., 2014) indicates that the Central America subduction zone has been active for a very long time $\sim$ Triassic and report a continuous slab from the surface to the proximity of the mantle-core boundary. However, other tectonic reconstructions exist that provide different outcomes, for instance Müller et al. (2016) whose reconstruction favour a subduction flip and a west-dipping subduction at ~ 130 –90 Myr for the Mexico-Central America-Caribbean region. However, several plate reconstruction studies (Ferrari et al., 1999; Lonsdale, 2005; Seton et al., 2012; Müller et al., 2016) convey on the more recent geodynamics development of this region, which has seen

a large Cocos-Nazca being subducted likely along an individual subduction segment for the last ≈ 30 -25 Myr concurrently with the formation of the Mexican slab (≈ 30 -25 Myr), at which followed the opening of the Cocos spreading ridge (≈ 23) currently separating the two plates (Schellart, 2020). Seismic anisotropy studies indicate trench-perpendicular and trench-parallel average fast direction in the sub-slab and wedge, respectively. A magneto-telluric study from Naif et al. (2013) reports the presence of a melt-rich ~ 30 km thick channel beneath the edge of the Cocos plate at the Middle America trench offshore of Nicaragua. Underneath the resistive oceanic lithosphere, the magneto-telluric data of Naif et al. (2013) reveals a high-conductivity layer confined to depths of 45 to 70 kilometres. Because partial melts are stable at these depths in a warm damp mantle, they interpreted the conductor to be a partially molten layer capped by an impermeable frozen lid that is the base of the lithosphere. This subduction zones also hosts various remains of oceanic plateaux, which we refer to Caribbean-Colombian plateau. The Cretaceous Caribbean plateau derives from a markedly heterogeneous mantle source region high high-MgO lavas are relatively common in this region (Kerr et al., 2002).

- (xi) The South America system is associated with the subduction of the Nazca plate along the western edge of South America. The Nazca plate velocity varies from 5.9cm/yr in the northern part to 7.0cm/yr in the remaining region, whilst trenches migrate at an overall steady-rate. Trenches, specifically, retreat at ~ -0.3 cm/yr in the extreme south and north ends of the system and, advance in the central region up to 1.7cm/yr. Flat subduction likely occurs in Peru and Central Chile, and around the Chilen Ridge (Lallemand et al., 2005; Syracuse and Abers, 2006), whereas the Nazca slab dips to the east with a moderate angle from its southern edge 46° to 34° S, (Lallemand et al., 2005; Syracuse and Abers, 2006). Although the causes of slab flattening are still debated, one of the hypotheses is that slab flattening was initiated due to the subduction of allochthonous terranes, and possibly an anomalously buoyant and thick oceanic plateau (Flueh et al., 1998; Gutscher et al., 2000; Yáñez et al., 2002). Average seafloor ages of the subduction zone are Paleogenic. Many studies interpret this subduction system has very old and of prolonged existence dating from the Late Triassic/Early Jurassic (e.g., Coira et al., 1982; Vasquez et al., 2011; Maloney et al., 2013; Schellart, 2020), which is in accordance with reported old ages being found in the geological record of the southern Central Andes (e.g., Mackaman-Lofland et al., 2019). However, other studies claim younger Cretaceous ages in the north, becoming progressively younger southward (e.g., Pardo-Casas and Molnar, 1987; Chen et al., 2019). Chen et al. (2019), rock samples and tomographic studies report a continuous magmatic-arc record and suggest a continuous slab from surface down to 2400 ± 200 km depth

underneath the Peru segment of the South American subduction zone. MacDougall et al. (2012); Long and Wirth (2013) report a variable but mostly trench-parallel trend for the seismic anisotropy in the mantle wedge. Models for sub-slab anisotropy show that significant misalignment is present below the slab and that the orientation of sub-slab fast shear direction sometimes differs from anisotropy above the slab. This in agreement with the more recent work of Long et al. (2016), reporting a distinctive slab-parallel fast orientation within the Nazca slab and mostly plate motion-parallel to oblique signatures in the sub-slab region. A very interesting feature of this system is the varying geometry of the sub-slab anisotropy relative to the strike of the slab and to the direction of the Nazca plate motion. The best fitting model for the southern region ($\sim 40^\circ\text{S}$) shows that the sub-slab a-axes are aligned sub-parallel and are considered parallel to Nazca plate motion, (Hicks et al., 2012; MacDougall et al., 2012); on the contrary, localities closer to flat slabs ($\sim 35\text{--}36^\circ\text{S}$), reveal sub-slab a-axes oriented roughly parallel to slab strike, while at $\sim 31\text{--}33^\circ\text{S}$, sub-slab a-axes follow the curve of slab strike, but also align sub-parallel to Nazca plate motion and the slab’s down-dip trajectory.

6.7.3 Additional results

In this study, we first address the impact of a ULVz and OPl on slab and trench deformation style, and later the inferences this yields on patterns of mantle flow. We group the outcomes of our models in 4 different classes with different initial conditions: i) a reference model (RM) which does not contain an OPl and ULVz; ii) M1 which includes a ULVz of half of the plate width ($W_{OL} = 1/2W_{ULVz}$); iii) M2 which includes an OPl of half of the plate width ($W_{OL} = 1/2W_{OPl}$); iv) M3 combined model, which includes an OPl and ULVz both widths half of the plate width; and v) M4 including a ULVz with the same lateral extension of the subducting plate ($W_{OL} = W_{ULVz}$) (see figure 6.5 and table 6.7). All models share in common the typical three stages of subduction evolution consistent with previous work (Bellahsen et al., 2005; Stegman et al., 2006; Yamato et al., 2009). The evolution of the different classes of models (RM-M3) is presented through a comparison between end-members taken during the final steady-state stage of subduction (figure 6.7).

The reference model does not contain a ULVz and OPl, and as such, all the layered modelled are laterally homogenous (figure 6.5 and 6.8). This class of models is needed to ensure that if there are no lateral viscosity variations, no vortex-like features emerge in the flow. Hence, being ideally zero toroidal flow (see §5), we can isolate and assess the impact of each factor of interest on the outcome. The time evolution of the RM is characterized by trench retreat and slab roll back over time, with a steady-state stage reached in approximately 8 Myr (figure 6.7). Slab rollback movement allows for the

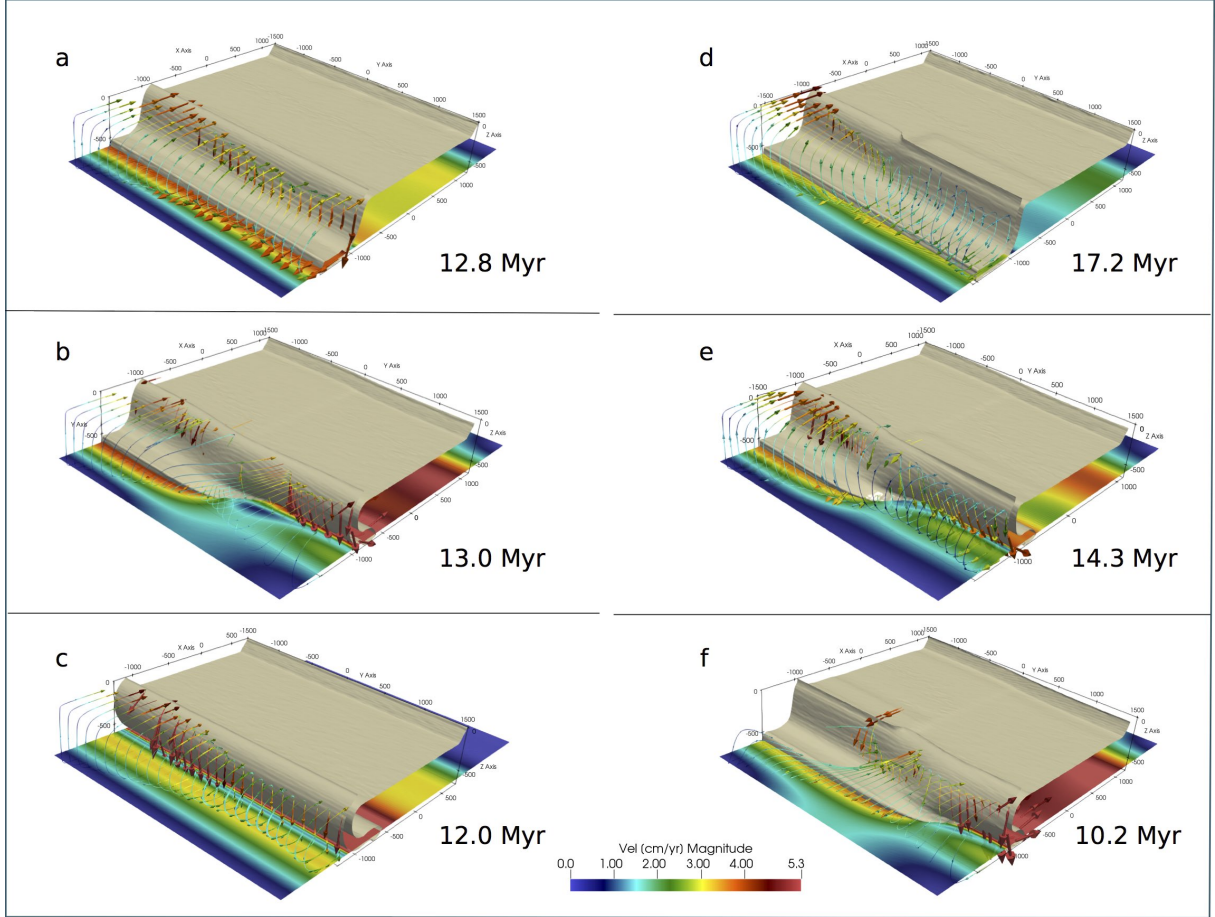


Figure 6.7: A comparison between the steady-state time evolution of RM (a) , M1 (b), M4 (c), M2 (d), M3 (e) and M5 (f) modelled cases. RM ($W_{OPl}, W_{ULVz} = \text{none}$), M1 ($W_{ULVz} = 1/2 W_{OL}$), M4 ($W_{ULVz} = W_{OL}$), M2 ($W_{OPl} = 1/2 W_{OL}$), and M3 and M5 ($W_{OPl}, W_{ULVz} = 1/2 W_{OL}$). The color-bar indicates the velocity magnitude which also applies to glyph vectors and streamlines.

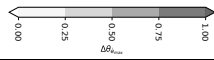
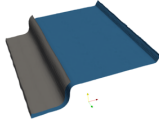
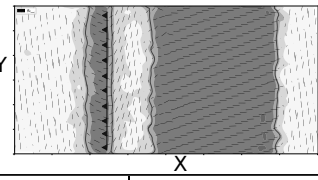
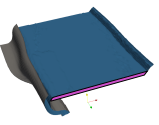
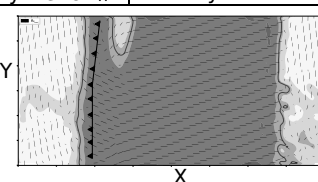
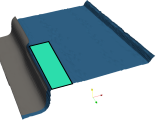
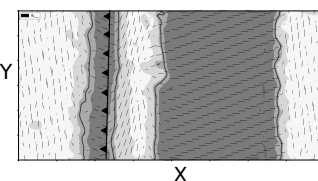
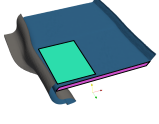
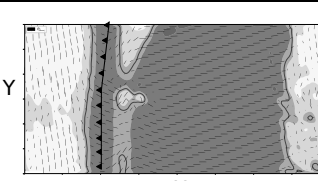
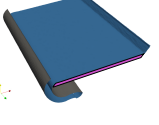
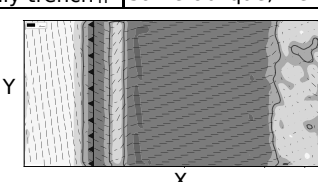
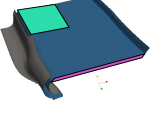
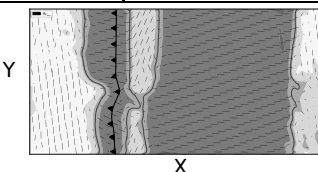
Model results						
model category #ID	Slab morphology		V_{norm}	mantle flow 		model application
	model output	slab 2D sketches		wedge	slab	
RM			- 0.406 + 0.0074	mainly trench	trench to trench ⊥	Ho, Ku, Ke
						
M1			- 0.0780 + 0.0856	mainly trench	mainly trench ⊥	Iz, Sa, At, Cs
						
M2			- 0.407+ 0.00747	mainly trench	trench to trench ⊥	Hb, Cr
						
M3			-0.0708 + 0.0954	mainly trench	some oblique, trench ⊥	Sa, Mr, Am Hk, Ka
						
M4			0.136 + 0.000599	mainly trench	some oblique, trench ⊥	Ho
						
M5			- 0.110+ 0.0216	Trench to ⊥	some oblique, trench ⊥	Am, Ka
						

Figure 6.8: An extended view of the modelling outcomes for slab morphology, trench migration and plate velocity ratio, v_{norm}^* , absolute normalised angle of the flow field θ' and corresponding natural application.

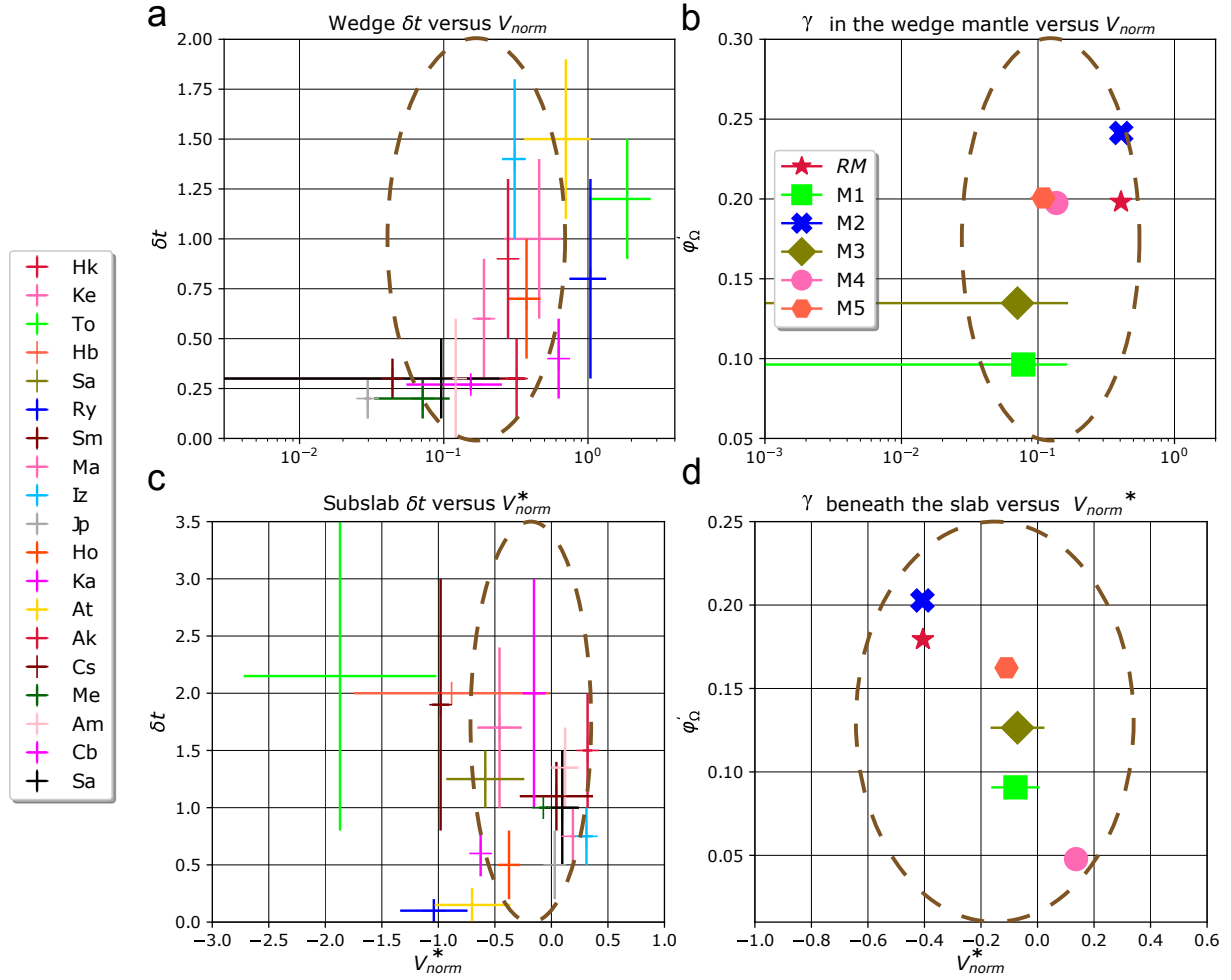


Figure 6.9:

A comparison between model results of the overall misalignment in the prevailing direction of maximum finite strain, γ , of θ_{MPFS} , (b and c) and observations of shear wave delay time for natural subduction zones (a and d). In a, observations of delay time, δt , versus V_{norm} in the wedge. In b, model results of γ versus V_{norm} in the wedge. In c, δt versus V_{norm}^* from observations in the sub-slab domain; and in d, γ versus V_{norm}^* from model results also in the slab-mantle regime. Our results could apply to those natural subduction zone falling within the dotted area where $V_T \pm 1/3V_P$.

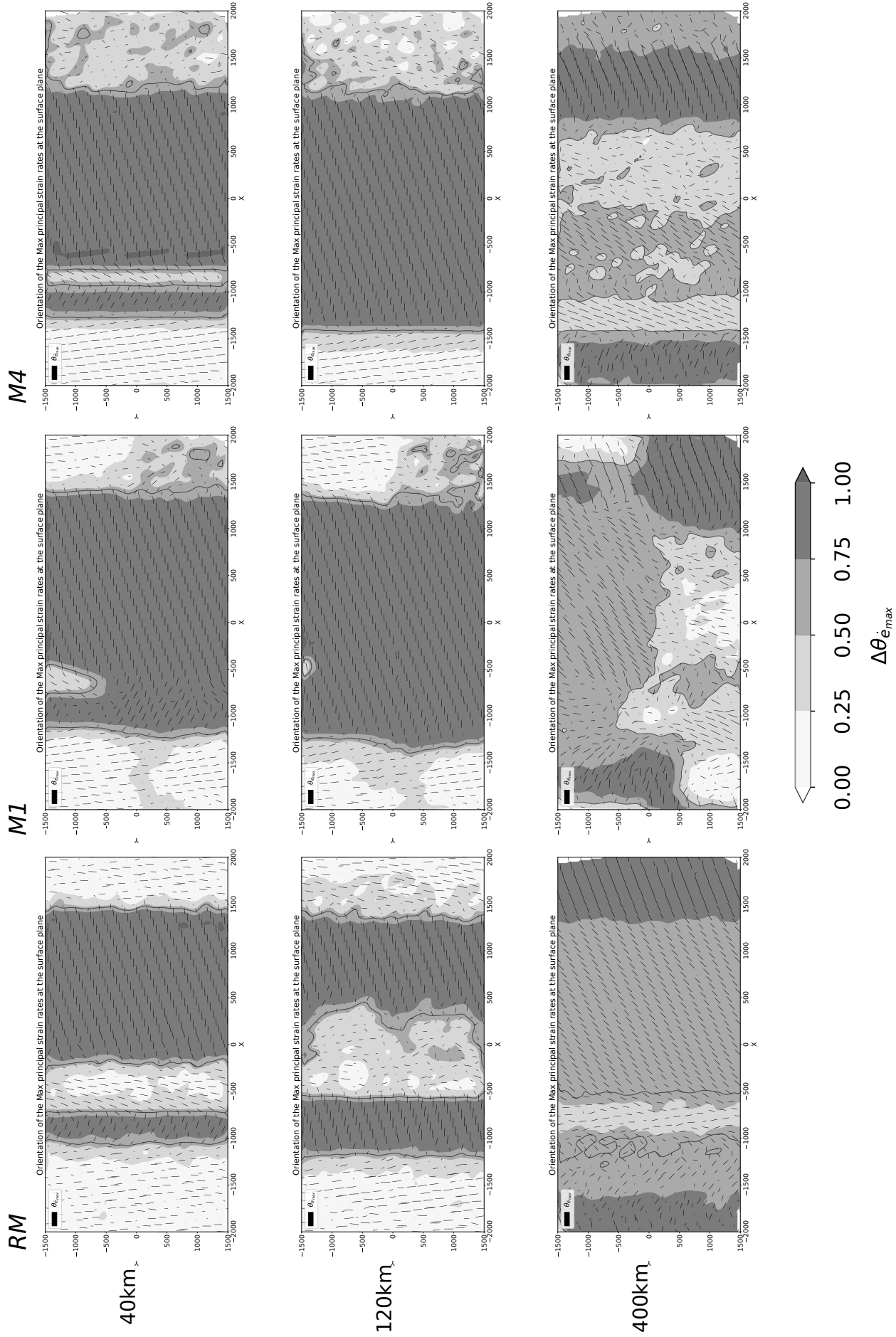


Figure 6.10: A comparison between normalised prevailing direction of finite strain in slab-mantle cross sections at different depths for RM, M1 and M4 (left to right).

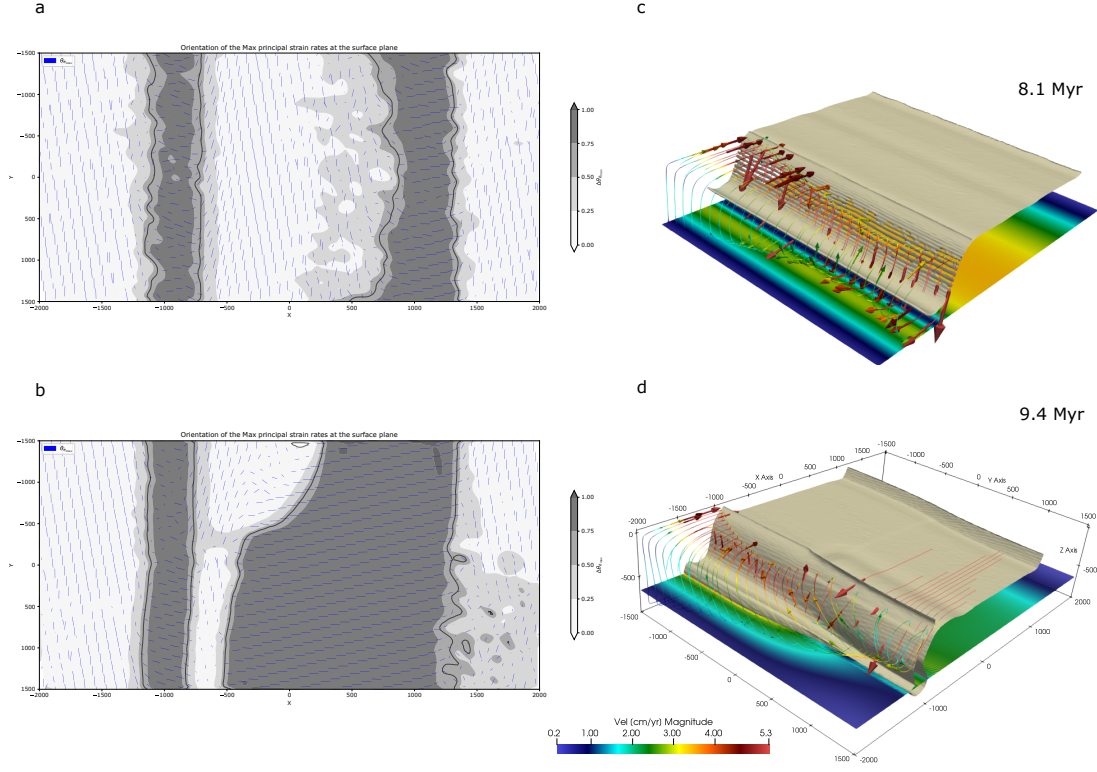


Figure 6.11: A comparison between the steady-state stage of the subduction evolution of RM and M1, in a and b, respectively, both including the lower mantle. In a and b, normalised prevailing direction of finite strain in slab-mantle cross sections, θ' . In c and d, slab contour and velocity magnitude including glyph vectors and streamlines of RM and M1, respectively. The upper-lower mantle viscosity contrast is 30, and the lower mantle extends until 800 km under the surface. The presence of a ULVz hinders the development of a strong trench-parallel flow anisotropy behind the subducting slab, which otherwise would be well extended (a and b).

entrainment of mantle material beneath the subducting slab, which gives rise to the observed trench-parallel fast direction within the slab-mantle and sub-slab parallel flow at further depth reflecting mild decoupling between the lithosphere and mantle (figure 6.12). The subduction time evolution of the RM is consistent with previous studies (Bellahsen et al., 2005; Stegman et al., 2006; Schellart, 2008b), and proposed seismic anisotropy models for entrained 2D mantle flow (e.g., McKenzie, 1979; Turcotte and Schubert, 1982; Long, 2013).

On the other hand, M1 develops variations in slab shape, trench migration and mantle flow across the subduction zone. In particular, the area of the oceanic lithosphere underlined by a ULVz exhibits a quasi-stationary trench, faster plate velocity and buckled slab at the 660 km discontinuity. Conversely, the remaining half shows a retreating trench with a slab that lays flat at the upper mantle transition zone (figure 6.2). The development of a strong toroidal cell at the base of the lithosphere generates a change in slab shape from convex to concave (figure 6.2b and d, and §5 for more details). The steeper slab morphology and trench migration motion act as an effective barrier to horizontal mantle flow beneath the slab in the trench-normal direction, hence forcing mantle material to escape laterally in response to the asymmetric motion of the slab and trench. As a result, the flow migrates predominantly perpendicular to the trench zone until it curves to align parallel to the trench zone in the deeper sub-slab mantle, in the portion of the sub-slab with a ULVz. At shallower depths, the flow aligns parallel to plate motion. The development of a vortex-like feature could also indicate the development of a strike-slip tectonic regime in the central part of the mantle wedge (figure 6.3).

The final stage of the temporal evolution of M2 is characterized by an intermediate V_P between the previous two cases. The half part of the slab with an OPl exhibits a slowly moving trench and a slightly steeper slab (figure 6.7) compared to the other half. At the boundary between the two areas, a small tear can develop in the subducting slab as a consequence of the stretching and rotation motion occurring along the trench zone. The ingestion of lighter material at the trench zone causes the development of a shallow toroidal cell, however, this is insufficient to significantly hamper the flow circulation behind the sub-slab deeper mantle where ϕ' aligns slab-parallel similarly to for the RM case. Trench-strike ϕ' variations reflect changes in trench migration rates (figure 6.8) developing a stronger component of shallow corner flow in the portion of the model containing an OPl. However, the strong trench-parallel flow signature just behind the trench, consistent with the RM, suggests that plate and mantle are still reasonably well coupled at depth (figure 6.15).

M3 includes both an embedded ULVZ and OPl at the base and surface of the subducting oceanic lithosphere, respectively (figure 6.8). This class of models shows V_P larger than in M2 but lower than in M1, with an emerged combined slab morphology between the two. The area of the slab including ULVz and OPl shows a concave slab shape with an

advancing trench whereas the other half displays a convex morphology with a retreating trench. The addition of an OPl contributes to the superficial toroidal motion, while at depth a ULVz generates strong toroidal flow, hence enhancing horizontal flow in the slab-mantle, and trench parallel flow in the deeper mantle underneath. As a result, the flow is deflected behind the subducting slab and transitions from perpendicular to curved, possibly representing the development of strike-slip regime in the slab and wedge. Stronger ϕ' variation along trench-strike variations are present and reflect the asymmetry in trench motion and slab shape, as well as flow signature (figure 6.14).

M4 is a laterally homogeneous model including a ULVz. This case is characterised by the largest rates of trench advance and plate horizontal velocity, as well as slab dip throughout the whole lateral extent of the model. The absence of toroidal flow in this class of experiments shows that trench-parallel flow in the deep sub-slab mantle can be due to the presence of a strong decoupling asthenosphere only (figure 6.13).

The subduction evolution of M5 which include a ULVz and OPl at the opposite side of the subduction zone is similar to M3. Differently from other models, this class of numerical experiments develops larger variability in trench strike parallel mantle patterns (figure 6.8). This variability is due to the development of two toroidal vortex-like structures flowing in opposite directions, from which emerges transform motion behind the trench zone and larger corner-flow signature in the wedge mantle farther away from an OPl (figure 6.7 and 6.16).

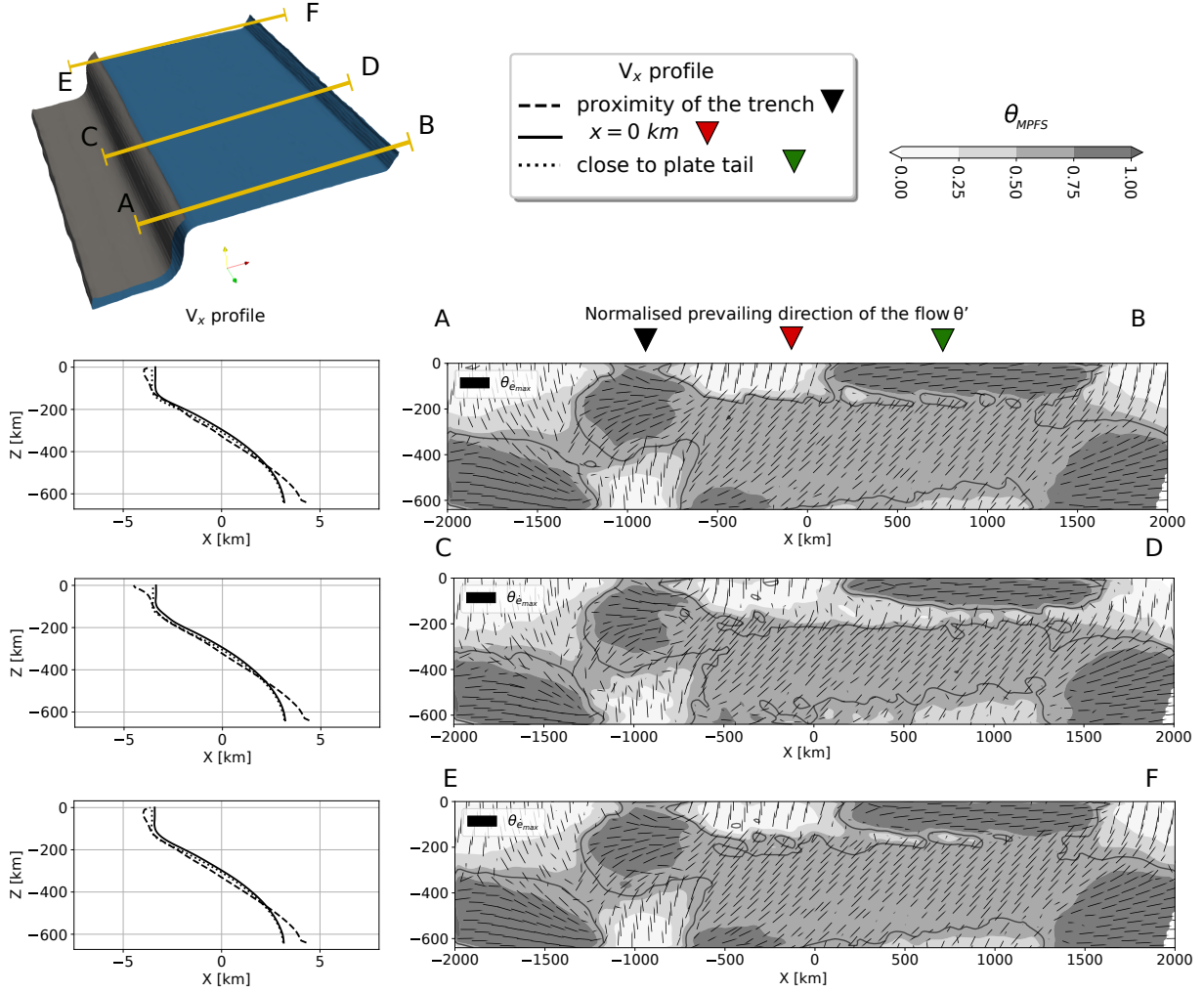


Figure 6.12: Summary of the RM outcomes for i) the subduction style (top left); ii) patterns of prevailing direction of the flow field, θ_{MPFS} through front (profile AB), mid (CD) and back vertical planes (right top to bottom; and iii) horizontal and vertical velocities profiles in the centre and in the wedge sections of the model, respectively (left).

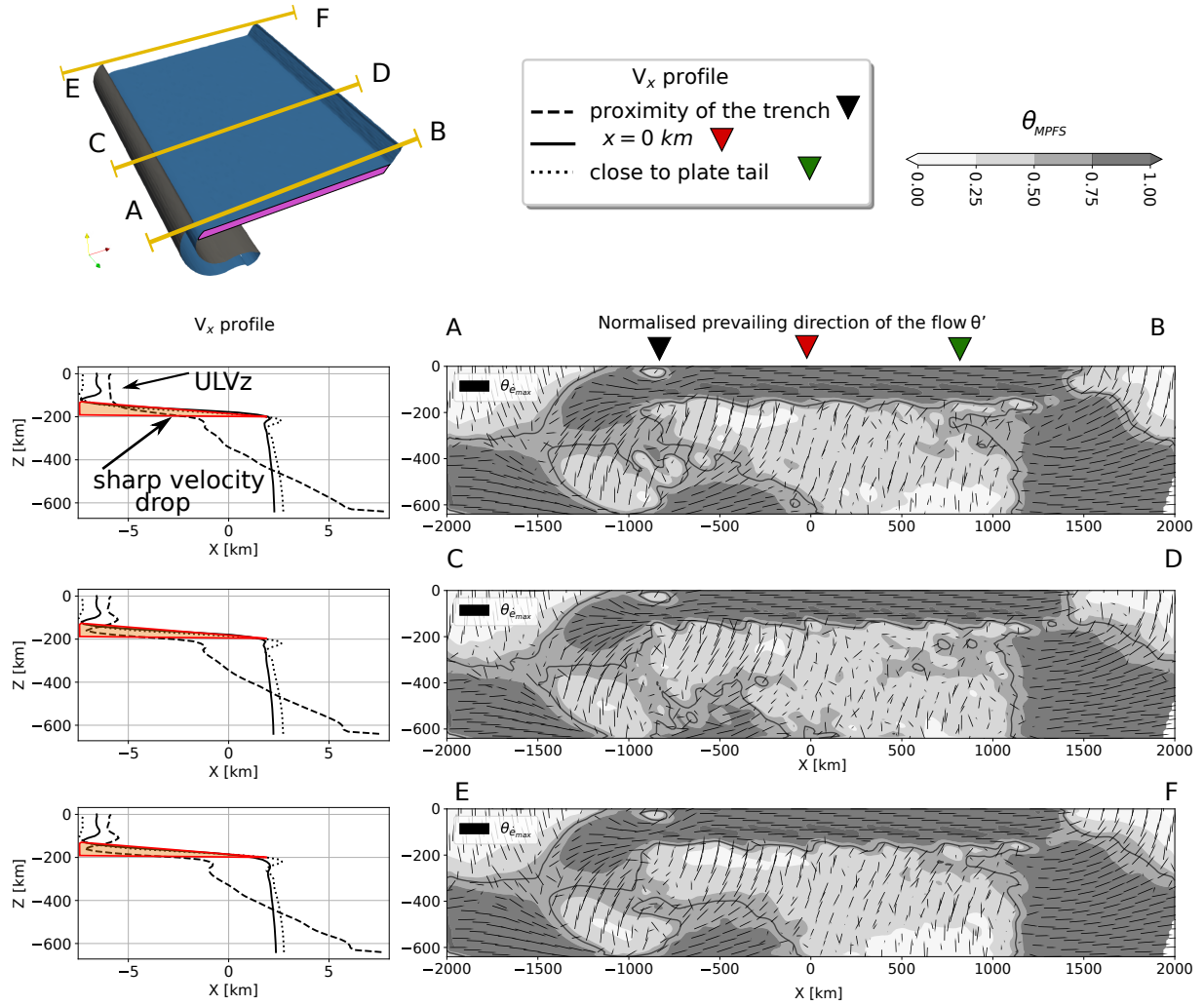


Figure 6.13: Summary of the M4 outcomes for i) the subduction style (top left); ii) patterns of prevailing direction of the flow field, θ_{MPFS} through front (profile AB), mid (CD) and back vertical planes (right top to bottom; and iii) horizontal and vertical velocities profiles in the centre and in the wedge sections of the model, respectively (left).

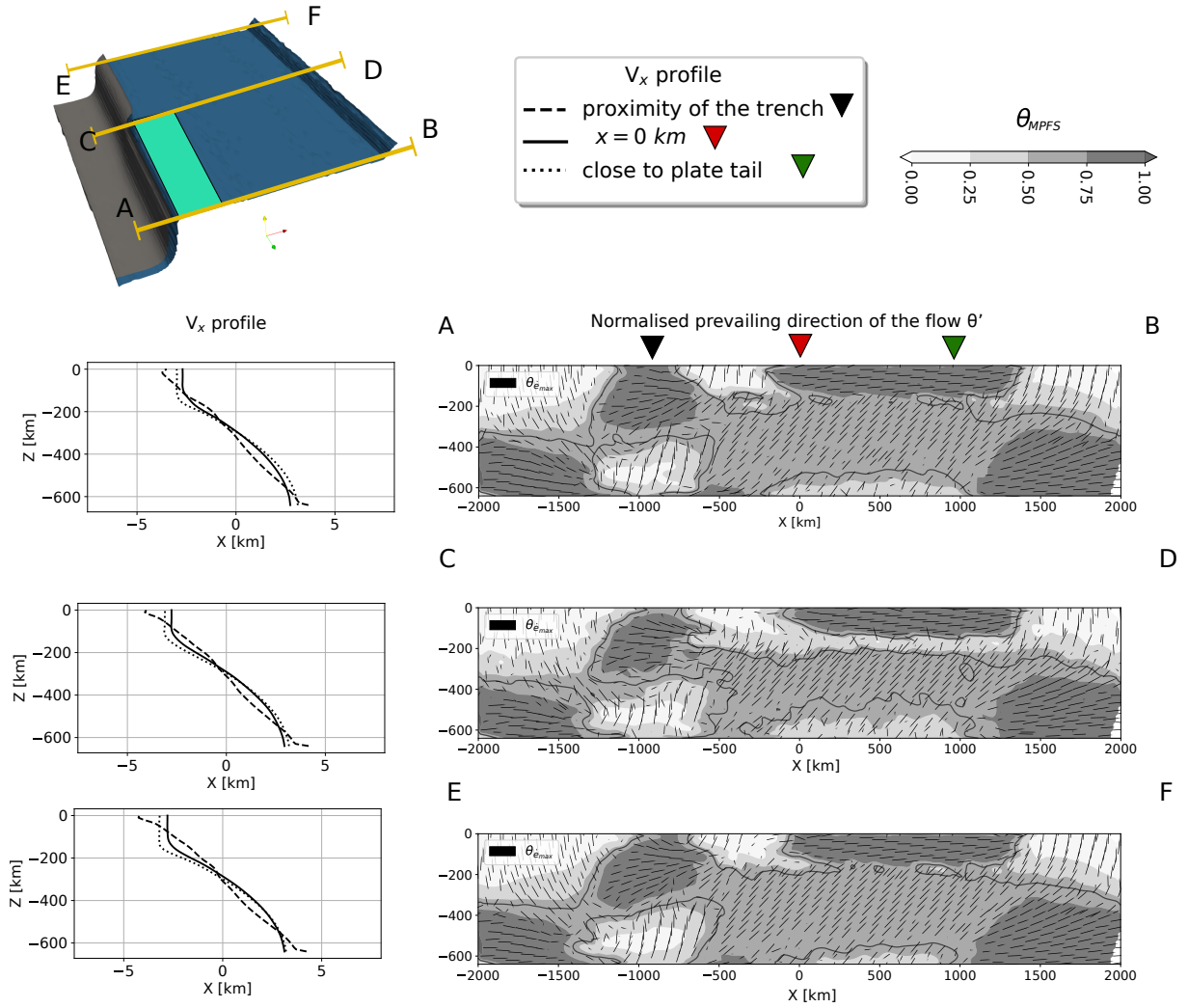


Figure 6.14: Summary of the M2 outcomes for i) the subduction style (top left); ii) patterns of prevailing direction of the flow field, θ_{MPFS} through front (profile AB), mid (CD) and back vertical planes (right top to bottom; and iii) horizontal and vertical velocities profiles in the centre and in the wedge sections of the model, respectively (left).

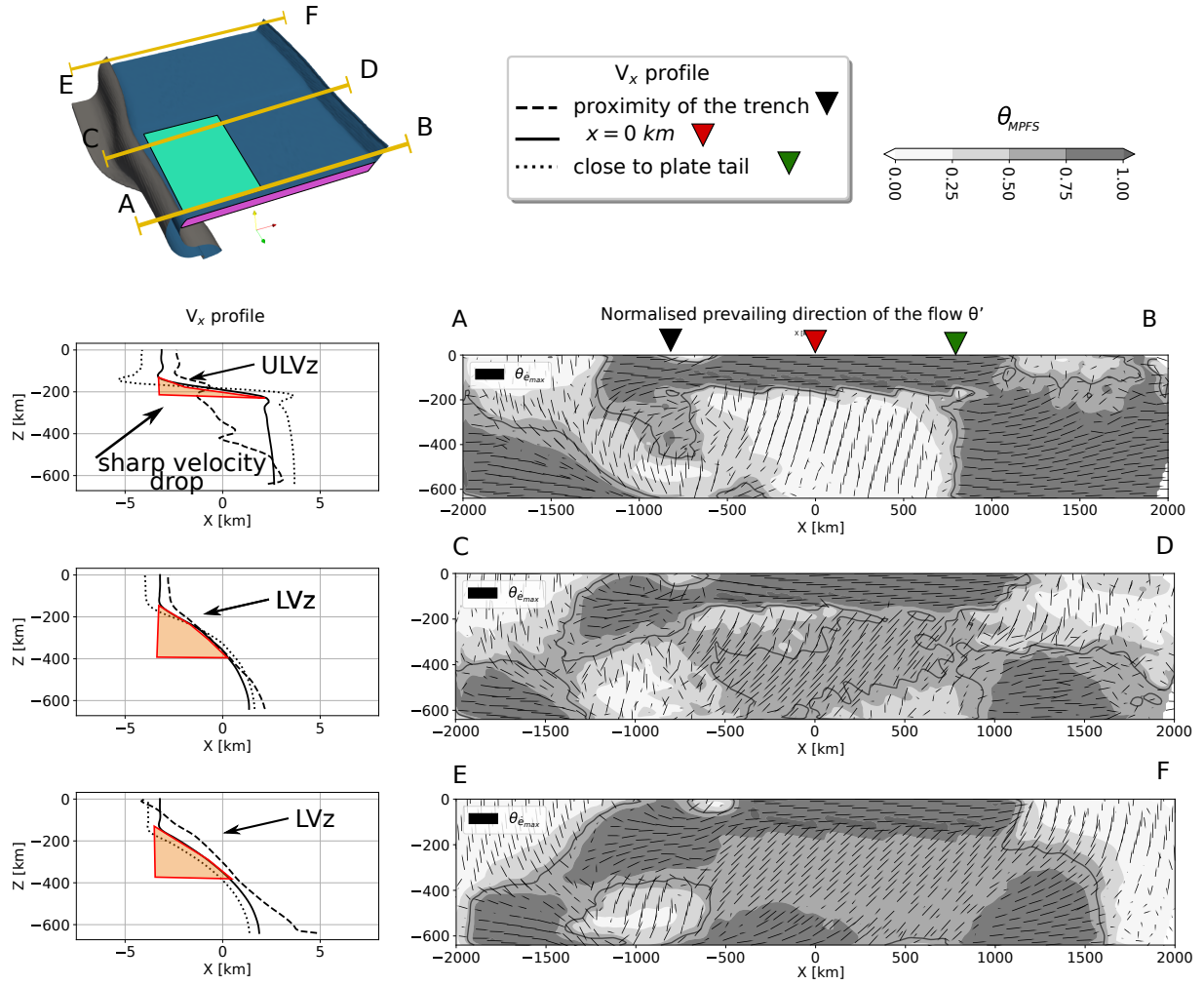


Figure 6.15: Summary of the M3 outcomes for i) the subduction style (top left); ii) patterns of prevailing direction of the flow field, θ_{MPFS} through front (profile AB), mid (CD) and back vertical planes (right top to bottom; and iii) horizontal and vertical velocities profiles in the centre and in the wedge sections of the model, respectively (left).

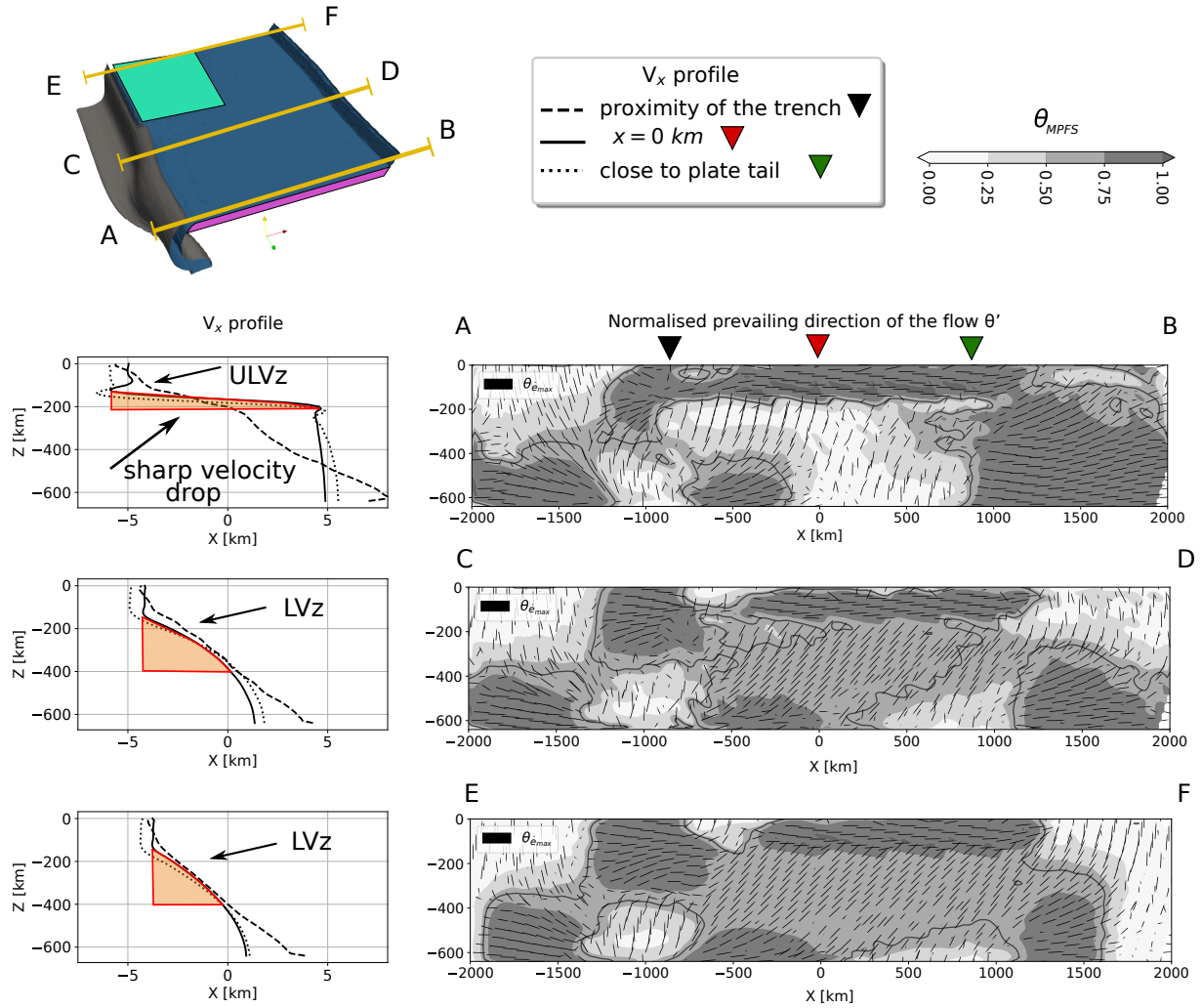


Figure 6.16: Summary of the M5 outcomes for i) the subduction style (top left); ii) patterns of prevailing direction of the flow field, θ_{MPFS} through front (profile AB), mid (CD) and back vertical planes (right top to bottom; and iii) horizontal and vertical velocities profiles in the centre and in the wedge sections of the model, respectively (left).

7. Chapter 7: General discussion

7.1 Introduction

The overarching aim of the research work presented in this thesis was to provide a novel conceptual framework for the interpretation of the geological record of complex regional subduction systems, where the presence of a very weak and thin decoupling sublithospheric layer (SLL) has been inferred by geophysical studies and for which very little or poor holistic modelling exist. Thus the scarcity of previous numerical subduction modelling work on the same topic highlights the novelty of the research. Furthermore, the accurate quantification of the SLL impact on subduction dynamics and induced mantle flow provided through this thesis work reveals the critical importance of these findings for the interpretation of plate margins (figure 6.1). Ultimately, the compiled dataset of inferred SLL weaknesses throughout the Pacific basin attempts to gather information about where SLL exist and explore potential correlations with patterns of subduction anisotropy, slab migration style and surface motions, for the first time.

In 3D large-scale simulations, it is numerically challenging and computationally expensive to incorporate very thin layers with large viscosity reduction relative to the surrounding materials. While this thesis necessarily presents its own set of numerical limitations, I was able to overcome such difficulties through dedication, systematic data analysis and teamwork. In addition, the possibility to constrain the influence of SLL alone, and related parameters, contribute to a resolution that when a particular effect is found, this is only the result of an SLL and is not due to other parameters. This is of particular relevance as very little is known about SLL implications on subduction dynamics; thus, it was crucial to constrain the effect of an SLL alone.

These research findings, therefore, deepen our understanding of slab and mantle dynamics, seismic anisotropy, with implication for the better comprehension of driving forces of plate tectonics. Hence, providing a robust foundation for the reinterpretation of some of the geological record along some segments of the Circum Pacific belt (figure 6.1) and for the future modelling of more complex geological settings that further investigate the mechanisms of an SLL in combinations with other factors.

In this chapter, I first outline how the overall research work compares to previous work on a very similar topic (§7.2). Then, I summarise the results of each research module and discuss in greater detail the scientific contribution and relevance of each of them compared to previous work on related topics (§7.2.1-7.2.3). The numerical challenges and future research outlook are outlined in §7.3, whilst conclusions upon the relevance of the research findings and broad implications for the modelling of slabs' and mantle dynamics during subduction are drawn in chapter 8.

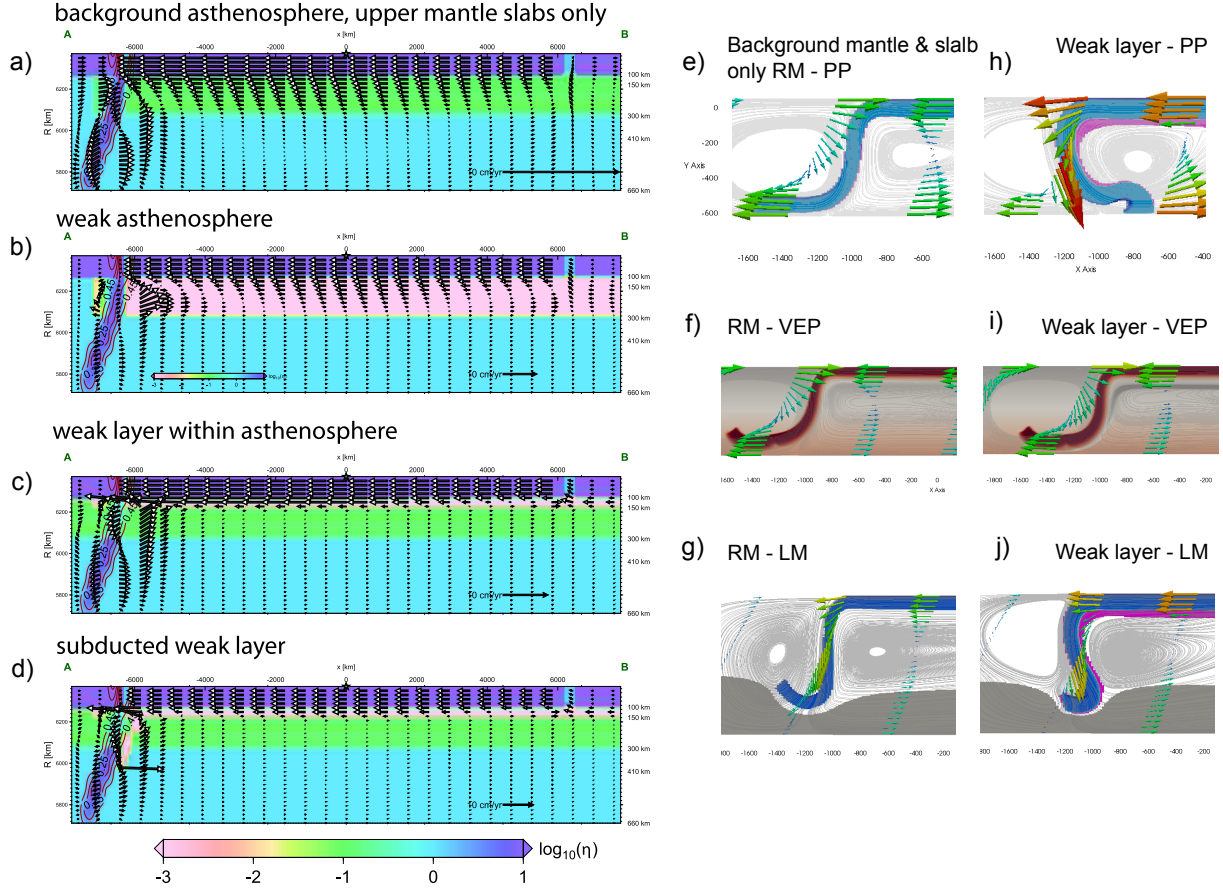


Figure 7.1: A comparison between the study of Becker (2017) and Carluccio et al. (2019b). Becker (2017) shows the models for a) slab and background mantle only; b) extended weak asthenosphere; c) weak layer within the asthenosphere (i.e., SLL); and d) a subducted weak layer. Carluccio et al. (2019b) show the results for the background reference model (e, f and g) and the weak layer within the asthenosphere (h, i and j) for different configurations: pseudo-plastic (PP) rheology in e and h; visco-elasto-plastic (VEP) rheology, in f and i; and PP with the inclusion of the lower mantle in g and j. The glyph vectors and streamlines refer to the velocity field. In both studies, the decoupling effect of a weak layer can be observed by looking at the abrupt velocity drop occurring within a thin interval in the sub-lithosphere, which is further indicated by the lengths of the velocity vectors. When only background mantle or thickened asthenosphere are implemented, the velocity decreases more gradually in a thicker interval, and plate and mantle are reasonably well coupled at depth.

7.2 Main findings

Comparing this thesis outcome with previous numerical studies analysing the consequences of sub-lithospheric layers in subduction geodynamics shows that SLL has a significantly wider range of effects than previously assumed. In this study, it has been shown that the partitioning of both the subduction surface motion and the three-dimensional deformation field, as well as patterns of subduction anisotropy and slab deformation style, were substantially affected by an SLL. None of the previous studies has analysed all the consequences investigated in this work and the only fully dynamic modelling study (Becker, 2017) that did so, had only reported limited second-order effects on anisotropy or plate velocity.

Figure 7.1 shows the comparison between Becker (2017) and my numerical outcomes. Becker modelled a number of scenarios, including background asthenosphere (i.e., reference model), thickened weak asthenosphere, weak thin layer within the asthenosphere and subducted weak layer, among others, as shown in figure 7.1a-d. He employs Newtonian rheology and reports that the cases with background mantle and thickened weak asthenosphere mainly show Poiseuille (i.e., pressure-driven) and Couette (i.e., simple-shear) flows. The weak layer (i.e., an SLL) case displays some channelised Poiseuille flow compared to the reference model but appeared to have only a modest impact on deeper shear. The subducted weak layer case does not modify plate velocities or mantle flow significantly further, according to the choice of parameters of Becker (2017). While he reports that overall azimuthal anisotropy misfit minima are predominantly sensitive to the bulk effect of a layer thickness and viscosity, he also suggests that for the thinner layers cases, it is mainly the modification of the partitioning between Poiseuille and Couette flows affecting the match with anisotropy, rather than the vertical deflections of flow (figure 7.2). On a similar note, the numerical 2D study of Liu and Zhou (2015) explored the role of deep >100 km thick recycled oceanic sub-slab material during subduction. Their findings support the development of trench-parallel sub-slab seismic anisotropy caused by a thick asthenospheric layer and provide a new mechanism for reducing slab dip angle; however, a 2D setup does not allow the development of toroidal flow, which generally leads to more complex patterns of mantle flow and anisotropy.

In comparison to Becker’s background asthenosphere case (figure 7.1a, b) my reference model (RM) shows a similar gradual velocity decrease beneath the lithosphere (figure 7.1e, f and g). In my models, when a weak layer is included, the velocity vectors display a sharpened transition from flow acting in the direction of plate motion to the counter-flow direction in the asthenosphere underneath, which is indeed the decoupling effect of the SLL (figure 7.1h, i and j). The effect on the velocity gradients can also be observed in figure 7.2 through a plot of the horizontal velocity field through depth in a, and a schematic comparison of their velocity gradients in b and c. The decoupling layer accommodates

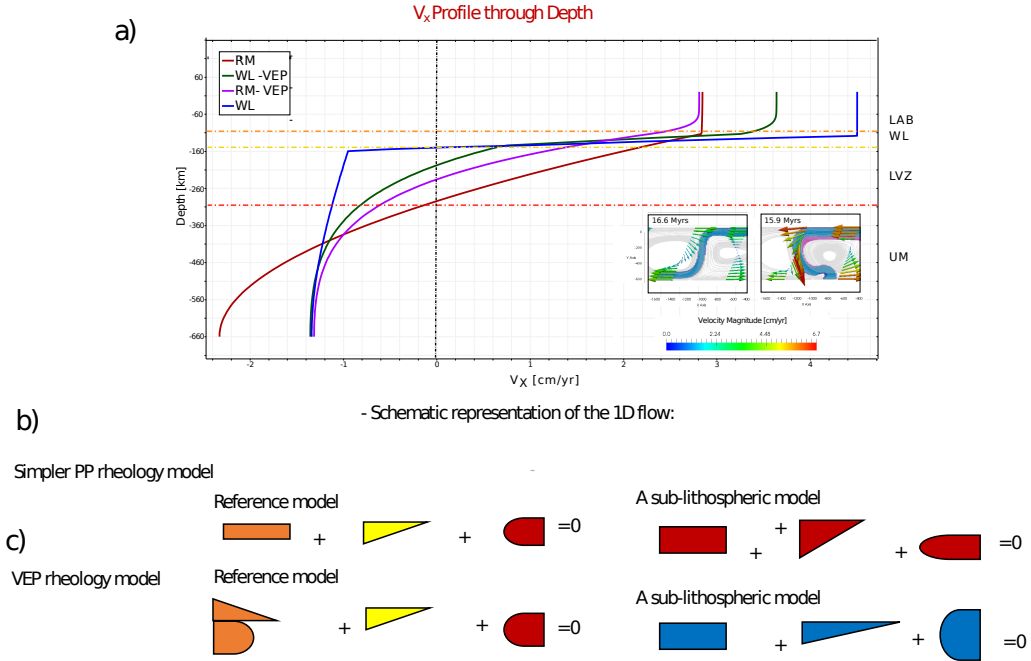


Figure 7.2: In a) a comparison between the horizontal plate velocity, V_x , profile through depth for the reference model (RM) and sub-lithospheric layer (SLL) cases for PP and VEP rheologies derived from the model implemented in chapter 4. Below, in b), a schematic representation of the velocity gradients in the lithosphere and mantle for the general PP RM and SLL cases (left to right). In c), RM and SLL cases (left to right) representation in case of VEP rheology. An SLL act as a slippery base for the motion of the lithosphere, by enhancing the vigour of the Couette flow and suddenly diminishing the velocity within the sub-slab; hence promoting the development of counter-flow in the asthenosphere immediately underneath the lithosphere.

Pseudo- Plastic (PP) rheology models

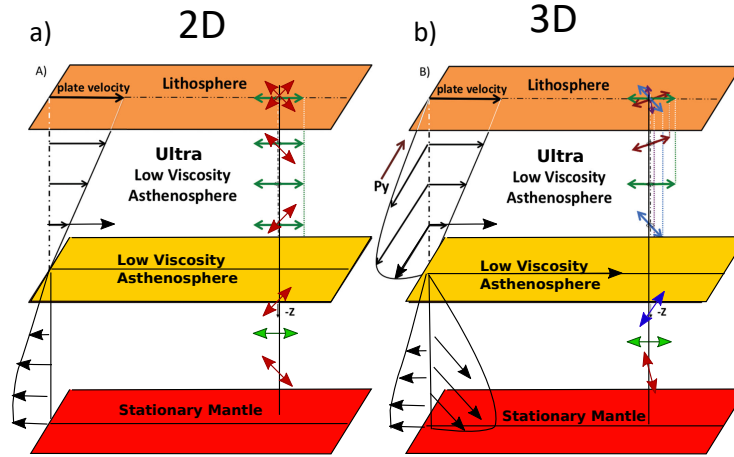


Figure 7.3: A schematic representation of the deformation field through a comparison between 2D (a) versus 3D (b) for PP simulations both including an SLL. An SLL sharpens the transition between Couette flow in the uppermost asthenosphere to Poiseuille counter-flow in the underlying asthenosphere. In addition to this effect, the 3D experiments also show a combination of Couette-Poiseuille flow perpendicular to plate motion, which may lead to LPO rotation. The figure is edited from Natarov and Conrad (2012) and it is used to outline the effect of an SLL.

intensive shear between the lithosphere and mantle blocks within a 10-30 km interval, rather than in a 100-300 km interval, and has a significant effect on mantle dynamics (chapters 5-6). From the analysis of the horizontal velocity profile through depth in figure 7.2a relative to the 2D models, it can be seen a velocity drop of ≈ 5.5 cm/yr within a ≈ 30 km SLL thickness in the direction opposite to plate motion, whereas if there is no SLL the velocity drops of ≈ 3.6 cm/yr in a ≈ 200 km thick layer to align in the counter-flow direction. These values corresponds to a difference of one order of magnitude in strain drop between the former and the latter cases, which could be as great as 2 orders of magnitude if the same velocity drop is recorded in a layer thickness of 10 km. In our models, the type of rheology that controls the type of flow in the lithosphere, given a constant temperature gradient in the mantle. Such as that when a Newtonian rheology is implemented, the Couette flow is dominant in the lithosphere and; on the contrary, a non-Newtonian approach leads to a combination of Couette-Poiseuille flow; however, irrespective of the rheological approach, an SLL maintains a constant decoupling signature. This decoupling signature is further illustrated through two comparisons of 2D versus 3D effects for pseudo-plastic (PP) and visco-elastic-plastic T-dependent rheology approaches in figures 7.3 and 7.4, respectively.

SLL influences on the deeper velocity and pressure fields can substantially affect slab morphology, plate velocities, trench migrations rates and mantle flow, as seen in figures 7.1

and 7.2. Hence, my potential explanation for Becker’s lack of SLL substantial effects relies on either his choice of parameters for the continental layer or the vicinity of the trench to the modelled wall and/or other boundary conditions, which might have precluded the development of appropriate subduction dynamics. Additionally, my current outcomes on slab deformation style have also been previously reproduced by using two additional numerical codes, such as MVEP and LaMEM, and further information is available through my master thesis (Carluccio, 2016). Eventually, a 2D model setup including a passive over-riding plate or the lower mantle was also tested with no substantial influence on the outcomes (see figure 4.10.3).

Another series of instantaneous kinetic-dynamic subduction models exist (Paczkowski et al., 2014a,b) reporting the role of mantle-lithosphere decoupling in enhancing trench-parallel flow in the sub-slab mantle domain, consistent with the “hidden” layer seismic anisotropy hypothesis of Long and Silver (2008). These models also show that greater amounts of decoupling are more likely to develop in longer and steeper slabs than in younger and more gently dipping ones. The work illustrated in chapter 6 generally agrees with the aforementioned outcomes on seismic anisotropy patterns in the sub-slab mantle. However, in this study, we were able to reproduce a large variety of seismic anisotropy signatures, mantle flow and surface motion patterns, as well as slab subduction styles within the same subduction zone by developing a fully dynamic numeric model (chapter 6).

While overall channels provide a better explanation for the elusive nature of the LAB Rychert (2015), these layers are likely of regional occurrence. Melt-induced viscosity reductions appear to be compatible with global plates and seismic anisotropy models, as long as their lateral extent remains limited compared to the rest of the plate Becker (2017). However, the non-uniqueness relating viscosity and thickness of deep layers, as well as the trade-off linking power and lateral extents of seismic methods can undermine the capacity to detect SLL. Additionally, a maximum increase in convergence velocity of a factor of ~ 2 due to the sub-lithospheric layer as reported by our outcome is likely to be neglected in global seismic anisotropy models; yet, its role can become relevant at a regional scale, influencing subduction motions and mantle flow dynamics.

In this study, the systematical analysis of our results has clearly revealed the effect of an SLL on the subduction partitioning, the deformation field partitioning and sub-slab subduction anisotropy. An SLL ultimately improves the ability to match natural observations with simpler numerical models and provide a novel conceptual framework to reinterpret some of the geological records of complex plate-margins; thus an important contribution has been made by providing a more complete picture of SLL role on subduction dynamics.

Visco-Elastic-Plastic rheology (VEP) models

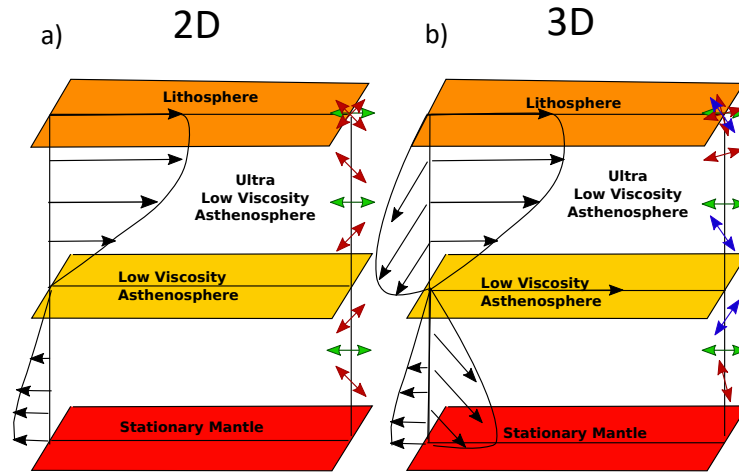


Figure 7.4: A schematic representation of the pressure field through a comparison between 2D (a) versus 3D (b) VEP simulations both including an SLL. Different from the PP case in figure 7.3, here we observe a combined Couette-Poiseuille flow in the lithosphere and uppermost asthenosphere in both 2D and 3D calculations. Then, the overall effect of an SLL is maintained by sharpening the transition between combined Couette-Poiseuille in the uppermost asthenosphere to Poiseuille counter flow in the underlying asthenosphere. The 3D experiments also show a combination of Couette-Poiseuille flow perpendicular to plate motion that has the potential to generate LPO rotation. The figure is edited from Natarov and Conrad (2012) and used to describe the influence of an SLL. When there is no SLL, i.e., only background mantle flow, the vertical variation in the orientation of the velocity vectors from parallel to opposite to plate motion is more gradual and does not necessarily lead to LPO 90° rotation.

7.2.1 A sub-lithospheric layer enhances plate motions and hampers trench retreat

The study presented in chapter 4 reveals the effect of a very weak layer in the top upper asthenosphere on the partitioning of subduction surface motions (Carluccio et al., 2019b). Such layer critically impacts subduction dynamics, particularly enhancing the plate horizontal velocity and hampering trench migration rates and, in some cases, controlling the morphology assumed by a subducting slab. Our results show that for large viscosity contrasts ($\eta_M/\eta_{WL} \geq 10^2$) and an SLL thickness greater than 20 km, the morphology of the subducting slab can be affected. In such cases, the subduction system is likely to develop a quasi-stationary or an advancing moving trench with a moderate to fast convergence rate for the plate and a slab that buckles at the 660 km discontinuity. For moderate viscosity contrasts between the mantle and the SLL ($10^1 < \eta_M/\eta_{WL} < 10^2$), the subduction velocity increases and trench retreat rate decreases, without significantly changing the shape of the slab. The overall effect is reduced where thinner and more buoyant layers exist (figures 4.2c, d and e, and 4.3b). The impact of these effects depends on the effective contrast between the modelled properties of the slab plus the upper asthenosphere layer, and mantle systems. In the research paper, I have shown and summarised this impact, consistent with the recent seismological constraints and observations, for example, a pile-up of weak material in the bending zone of the subducting plate (Hawley et al., 2016).

The simulations performed varied parameters over a range that is Earth-like, as shown in figure 4.1b and supporting table T1. This is relevant because previous numerical and analogical work with slab and mantle only (e.g., Schellart, 2004b; Bellahsen et al., 2005; Di Giuseppe et al., 2008; Funiciello et al., 2008) could not reproduce the dynamics of fast advancing moving plates with slow-moving trenches. Such as that when some trench advance was found, this was due to viscosity contrasts slab-mantle greater than 500, which are not generally observed in nature (Moresi and Gurnis, 1996; Mitrovica and Forte, 1997; Zhong and Davies, 1999) and resulted in slow plate convergence. More recent studies have shown the effect of an asthenosphere and mantle transition zone in reducing trench retreat (e.g. Holt and Becker, 2016; Billen and Arredondo, 2018), ridge push in enhancing plate motion (e.g. Capitanio et al., 2007), as well as an overriding plate or complex mantle rheology for reducing $|V_T/V_P|$ found in numerical modelling (Holt et al., 2015a,b; Holt and Becker, 2016); however, the literature has not explicated a clear trend for patterns of slab deformation style including slow values of subduction partitioning and a range of slab morphologies and dips. The setup chosen has contributed to the resolution of low trench retreat and fast plate speed being the result of a weak slippery base of the tectonic lithosphere only, whilst no extra forces or factors are needed (e.g. an overriding plate and an upper-lower mantle discontinuity); however, other factors already showing to reduce

trench motions may additionally coexist with the SLL or be of major relevance in other subduction settings where, for instance, trenches migrate faster than plates converge.

Here, I suggest that an extremely weak regional top asthenosphere mechanism is already of importance in decreasing the high absolute values of the subduction partitioning observed in many precedent models and to level more in line with the average values presently seen on Earth. Specifically, in our models the average absolute trench velocity is approximately 30 per cent or lower of the total convergence plate velocity (figure 4.3d), as found in many plate motion reference frames, (e.g. Goscombe and Gray, 2008; Schellart et al., 2008; Becker and Faccenna, 2009), and where the presence of a weak layer is confirmed by geophysical studies (e.g. Kawakatsu et al., 2009; Schmerr, 2012; Naif et al., 2013; Stern et al., 2015; Hawley et al., 2016).

7.2.2 A sub-lithospheric layer as an important source of substantial toroidal motion

The study carried out in chapter 5 uses 3D numerical models and examines the influence of a sub-lithospheric layer (SLL) width on the partitioning of the induced mantle flow into its poloidal and toroidal components (T/P). These findings outline how a non-uniform SLL alone can induce a consistent toroidal component underneath the base of the subducting plate, which propagates across the whole subduction region. The development of toroidal flow already occurs in models where the slab is continuous, hence has no free edge, and where the velocity field would otherwise be poloidal. Our results show that for intermediate to large viscosity contrasts ($\eta_M/\eta_{SLL} \geq 10^2$) and intermediate sub-lithospheric lateral extensions ($W_{SLL} > 1/4W_{OP}$ and $W_{SLL} < W_{OP}$), T/P substantially increases up to 0.5 at the surface and 2.0 further at depth. Furthermore, slab morphology changes from convex (S-shape) to concave (C-shape) from one side to the other of the subduction zone. Additionally, trench migration varies laterally from retreat to advance (figures 5.2-5.9). An SLL also promotes plate advance and hampers trench retreat, consistent with the 2D study of Carluccio et al. (2019b). The overall effect is reduced for thinner or stronger layers. This work eventually highlights the relationship relating SLL width to the partitioning of both subduction surface motions and induced mantle flow (figure 5.9); hence demonstrating the influence of mechanical processes occurring at the LAB on both subduction dynamics and surface motions.

Previous work has shown that buoyancy-driven fluids with only vertically layered viscosity do not generate any toroidal motion and lateral variable viscosity convection (e. g., with constant, depth-dependent or temperature-dependent viscosity) calculations typically produce only a limited amount of T/P (Christensen and Harder, 1991; Tackley, 2000). In this frame, variation in SLL width have shown to induce significant rotation at the base of the lithosphere, generating an additional toroidal cell at the mid-plane of symmetry of the plate; hence increasing the amplitude of the toroidal to poloidal ratio. We

present a novel regional mechanism, which is able to self-sustain the production of toroidal energy and, therefore, shows how buoyancy-driven poloidal motion can induce substantial toroidal flow through horizontal viscosity variations. As such, these findings provide an alternative explanation for the apparent lack of a direct link between the slab pull driving force in buoyant convection (which drives only vertical and divergent motion) and the generation of toroidal motion. Additionally, these findings outline a means to interpret slab variations in dip, morphology and trench migration throughout a subduction zone, as those observed along some of the subducting margins of the Pacific plate, such as the Mariana and Izu-Bonin subduction regions, where the presence of an SLL is confirmed by geophysical studies. Our chosen model setup allows us to test and verify that when substantial toroidal flow is found, this is not due to an artefact of numerical leakage, but to the SLL effect, whilst no extra forces or factors are needed. While other factors cannot be ruled out, the decoupling effect of SLL shows a consistent signature even with the inclusion of a visco-elastic-plastic temperature-dependent rheology or the upper-lower mantle transition zone (Carluccio et al., 2019b).

7.2.3 Asthenosphere dynamics during subduction

The research presented in chapter 6 outlines the potential of a sub-lithospheric layer (SLL) to deflect mantle flow around the subduction zone influencing the dynamic behaviour of subducting slab, the character of mantle flow and patterns of seismic anisotropy.

The first carveout of this research chapter is that the nature of the flow in the models can be divided into two domains depending on depth. These domains include: i) a shallower surface to approximately 270 km slab-mantle domain, and ii) a deeper sub-slab mantle domain. Patterns of seismic anisotropy can change dramatically depending on where the anisotropy is measured. In the slab-mantle domain dominated by an SLL, trench-perpendicular flow is recorded in the slab-mantle domain; whereas trench-parallel flow is prevalent in the sub-slab mantle. If plate and mantle are well-coupled, a transition from trench-parallel flow to trench-perpendicular flow is detected in the sub-mantle, in addition to slab-dip parallel flow in the sub-slab domain.

In my models, the estimate on the lateral strength of an anisotropy (such as the measure of the misorientation in the prevailing direction of the maximum finite strain throughout a plane) scales negatively with the ratio between trench and plate velocity in the slab-mantle and positively in the sub-slab domain. Thus, the second carveout of this study is that the ratio between trench and plate velocities along with the estimate on the lateral strength of the anisotropy can be used as a proxy to infer the amount of plate-mantle coupling at depth.

Moreover, in our results, the development of either trench-parallel or sub-slab parallel mantle flow regimes in the sub-slab mantle do not necessitate toroidal motion, as they can emerge spontaneously from a fully poloidal mantle circulation, where no lateral

viscosity variations are introduced into the models. The presence of a laterally varying asthenosphere generates toroidal motion allowing miscellaneous patterns of mantle flow and seismic anisotropy to co-exist within the same subduction system.

My finding confirms the seismic anisotropy hypothesis that a very weak “hidden” layer of embedded asthenospheric material underneath the subducting lithosphere promotes trench-parallel flow sub-slab anisotropy patterns, if these patterns are evaluated in the “deep” sub-slab mantle domain.

In addition to these outcomes, I also compiled observations suggesting the presence of a very weak sub-lithospheric layer beneath the base of some major subduction zones. By compiling, for the first time, these observations with other subduction zone parameters, I report a spatial relationship with the presence of a relict oceanic plateau (OPl) and observations of SLL along some segments of the Circum Pacific Belt. I, therefore, investigate the combined effects of OPl and SLL on subduction dynamics, also for the first time. The results show that an uneven SLL promotes plate-motion-parallel flow in the sub-slab and asymmetric patterns of slab and trench deformation, whilst an OPl aids further superficial complexity and stronger corner flow in the shallow wedge mantle.

While previous numerical work has deepened our understanding of subducting slab dynamic evolution and their relationship with other seismic observables, it could not represent such a large lateral variability for slab subducting style, surface motions, mantle flow and seismic anisotropy patterns within the same subduction zone as it was possible to represent in this research work. The modelling of a subduction zone has also not yet explicated direct correlations among patterns of slab deformation style, seismic anisotropy and subduction surface motions, especially in the case of slow retreating trenches and fast-moving plates (Lallemand et al., 2005; King, 2007; Coltice et al., 2017).

In addition to this, some previous numerical outcomes on the relationship between seismic anisotropy and subduction parameters could have been influenced by the choice of plate initial geometry (e.g., slab-free edge toroidal motion), boundary conditions (e.g., external pressure gradient or fixed trench) or approaches to evaluate the anisotropy itself (e.g., use velocity gradients versus strain geometry) and do not adequately represent a fully dynamic evolving system. This research work employed a fully dynamic approach where the slab subducts into the mantle under the effect of its own buoyancy and no further conditions are applied.

While other factors could contribute to anisotropy patterns and/or co-exist with the mechanism illustrated here, the aim of this chapter was to isolate the contribution of SLL only on patterns of seismic anisotropy. Yet, an SLL has already shown to maintain a constant signature on patterns of subduction zone anisotropy, surface motions and slab deformation style when the lower mantle is modelled in 3D (figure 6.11), or when a non-linear rheology or an overriding plate are introduced in the 2D simulations (Carluccio et al., 2019b).

These findings provide a novel conceptual framework for reinterpreting the geological record of current and, possibly ancient convergent margins, such as Cocos-Nazca and Hikurangi- Kermadec among others, where plates move faster than their trenches retreat, in accordance with observations of SLL and OPl, for which previous numerical modelling has been proven difficult. Finally, these outcomes could have broad implications for better understanding the role of the asthenosphere on slab dynamic evolution and forces behind plate motions.

7.3 Next steps

A natural follow up paper will focus on exploring the effect of an over-riding plate in combination with SLL in a 3D system. This will provide me with the opportunity to test the seismic anisotropy hypothesis of trench-parallel flow in the mantle wedge induced by the presence of an SLL. In addition, in this paper I will provide an updated literature review on asthenosphere configurations, surface motions and subduction geodynamics constrain on patterns of seismic anisotropy. My preliminary 2D results with an over-riding plate have already confirmed that an SLL maintains a consistent signature on slab deformation style.

Another natural follow up will be the testing of the trade-off that exists among non-linear rheology effect, upper and lower mantle transition zone and an SLL on pressure and velocity gradients in the mantle. The supporting information in chapter 4 has already shown that an SLL keeps a constant signature irrespective of a non-linear rheology approximation or the presence of a lower mantle. However, the extent of the impact to which each of all the aforementioned factors could contribute to patterns of seismic anisotropy and slab deformation would be particularly relevant for the modelling of specific regional settings, such as the Mariana and South America trenches, among others, and this is yet to be investigated.

An appropriate scaling paper could address the behaviour of the sub-lithospheric boundary layer relative to the motion of the two adjacent blocks on the partitioning of the Couette and Poiseuille flows in the lithosphere and mantle. This could provide a broader clarification of how our findings fit in the large mantle convection picture and the importance of decoupling layers for the better understanding of how mantle convection drives plate tectonics.

Whether melt ponding layers can be generated and to what extent that can be self-sustained by hotspots or shear heating could also be a great scenario to test with computer simulations. However, the very existence of these decoupling channels itself represents more than an enigma. It would require broad community and interdisciplinary efforts to reveal where and why these layers exist. This thesis work attempts to drive more

attention to the topic, as we show that these channels have very relevant implications for subduction geodynamics.

Concluding now on a more technical note, I am very interested in creating a numerical tool to benchmark different codes for the modelling of a subduction zone in 3D. This could help to encompass numerical limitations and boost replicability, reproducibility and feedback across the geodynamics community. This could be achieved by combining the use of Jupyter Notebooks, Docker containers and Binder (JNDB) (Carluccio et al., 2019a), among other tools. Docker containers and services are powerful tools, as they can be connected together or connected to non-Docker workloads. They also allow the creation, storage and sharing of our own piece of code through a docker interface. The Binder Project makes it possible and easy to create shareable, interactive, reproducible environments of our own work on the clouds. Jupyter notebook allows creating coding presentations of our results in an editable and fashionable manner. Then, the code is made accessible, interactive and editable with whom the content is shared with. This combined (JNDB) approach allows students to easily set up their interfaces and blinder clouds of Jupyter notebooks to both promote feedback within the learning environment and share published data in form of tutorials within the research group and at the community level. Finally, if more computational codes can be made available through the docker interface without the hurdle to compile or learn each code architecture, this could be used to test different codes efficiency and computational costs for a given geodynamic problem; thus having particular implications for managing research costs, deadlines and aims in terms of feasibility.

8. Chapter 8: Conclusion

8.1 Conclusion

This thesis outlines the basis for the better understanding of the role of a very thin and weak sub-lithospheric layer (SLL) on subduction dynamics and large-scale mantle flow, in the light of the geophysical observations reporting a slippery tectonic base for the subducting lithosphere along some of the segments of Pacific ocean. As very little is currently known about the geodynamics impacts of this feature, we address this impact through the implementation of 2D and 3D numerical subduction models of slabs dynamics as they descend into the mantle.

In chapter 4, this thesis starts by exploring the effect of an SLL viscosity and thickness on plate and trench motions, as well as slab subduction style. The research outcomes show that an SLL affects the partitioning of the subduction surface motions by decreasing trench migration rate and increasing plate velocity. Specifically, in the models with an SLL, the average absolute trench velocity is approximately 30% or lower of the total convergence plate velocity, as found in many plate motion reference frames. These research findings provide a novel regional mechanism to reduce the values of subduction partitioning found in many precedent analogues and numerical models and level more in line with what is currently observed in several subduction zones on Earth.

In chapter 5, this thesis subsequently investigates the influence of SLL width on the partitioning of the mantle flow as it splits into its two poloidal and toroidal components. This research work finds that a non-uniform SLL can, by itself, introduce a consistent toroidal component underneath the base of the subducting plate already in models where the slab is continuous, and the velocity field otherwise would be poloidal. In particular, our results show a toroidal to poloidal average ratio that approaches 0.45, which fits well with the values expected for a no-net reference frame. This outcome provides an alternative explanation to the apparent lack of a direct link between the slab pull driving force in buoyant convection (which drives only vertical and divergent motion) and the generation of toroidal energy. Hence, providing a novel mechanism to explain how buoyancy-driven poloidal motion can induce consistent toroidal flow through horizontal viscosity variations and, thus, an answer to the so-called "poloidal-toroidal coupling problem". These outcomes suggest that regional variations in SLL width substantially influence the nature of the mantle flow around the subduction zone with a potential to affect subduction dynamics on a global scale.

In chapter 6, this thesis ultimately gathers and outlines information about where sub-lithospheric layers have been detected to exist along convergent plate margins. By compiling this information, I report a striking spatial coincidence with the presence of a relict oceanic plateau (OPl) and observations of an SLL underneath some segments of the subducting Pacific plate for the first time. Thus, the numerical models performed in this study explore the individual and combined role of an OPl and SLL on patterns of

subduction anisotropy, surface motions and slab deformation style. Additionally, this work tests the seismic anisotropy hypothesis under which observations of trench-parallel flow in the sub-slab could be induced by an SLL. The numerical outcomes of this thesis confirm this hypothesis in the sub-slab domain and further show how asymmetric patterns of deformation for the trench motion, slab deformation and mantle flow, as well as for seismic anisotropy signatures can be obtained through combined SLL and OPl influences which can generate and localise strike-slip motion. These outcomes could be particularly relevant for the Cocos-Nazca and Kermadec-South Tonga subduction regions, among others, where the presence of a thin and weak decoupling sub-lithospheric layer and oceanic plateaus has been previously inferred by geophysical and geological studies.

In conclusion, this thesis work has revealed how thin and very weak layers embedded in the Earth's deep interior influence some of the complexities we currently observe in natural systems, hence demonstrating the importance of Earth's interior tangential forces in shaping patterns of surface and mantle deformations. Our numerical results show for the very first time that a very weak and thin decoupling layer in the uppermost asthenosphere affects the dynamics of sinking plates. By acting as a slippery base for the motion of the subducting lithosphere, it sharpens the decoupling between the strong tectonic plates and the weaker mantle underneath. This has significant consequences on plates velocities and deformation, as well as on the mantle flow nature and evolution around the subduction zone. The relevance of this study lies in its ability to provide a new mechanism to fit simpler numerical models with natural observations. As such, my research findings open the way to a novel conceptual framework to interpret some of the regional observations of slow rates of trench migration and fast-moving plate of current and, possibly ancient complex convergent margins, for which little predictive and holistic modelling exists.

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