Rheological and kinematic conditions for forced subduction initiation at continental passive margins

Antoine Auzemery

**Rheological and kinematic conditions for forced subduction initiation
at continental passive margins**

**Reologische en kinematische vereisten voor initiatie van geforceerde
subductie op continentale passieve marges**
(met een samenvatting in het Nederlands)

**Conditions rhéologiques et cinématiques pour l'initiation de la
subduction forcée aux marges passives continentales**
(avec un résumé en français)

Proefschrift

ter verkrijging van de graad van doctor aan de
Universiteit Utrecht
op gezag van de
rector magnificus, prof.dr. H.R.B.M. Kummeling,
ingevolge het besluit van het college voor promoties
in het openbaar te verdedigen op

vrijdag 3 september 2021 des middags te 2.15 uur

door

Antoine Auzemery

geboren op 12 mei 1991
te Limoges, Frankrijk

Promotoren:

Prof. dr. L.C. Matenco

Prof. dr. D. Sokoutis

Copromotor:

Dr. E. Willingshofer

This research was accomplished with financial support from the European Union's EU Framework Programme for Research and Innovation Horizon 2020 “Subitop” under Grant Agreement No 674899

“Opportunities multiply as they are seized.”

Sun Tzu

Members of the reading committee:

Prof. Dr. Francesca Funiciello

Prof. Dr. Taras Gerya

Prof. Dr. Douwe van Hinsbergen

Prof. Dr. Jeroen van Hunen

Prof. Dr. Stefan Schmalholz

This is Utrecht studies in Earth Sciences volume 239

ISBN 978-90-6266-606-5

© Antoine Auzemery

Printed in the Netherlands by Ipskamp Printing

Cover: Long Exposure Of Tungurahua Volcano Explosion At Night, Ecuador, South America © KalypsoWorldPhotography

All rights reserved. No part of this publication may be reproduced or transmitted in any form or by any means, electronics or mechanical, including photocopying, recording, or by any information storage and retrieval system without prior written permission from the author.

Table of content

Summary	9
Chapter 1 - Introduction	14
1.1 Mechanisms of subduction initiation	14
1.2 Rheology of passive continental margins	15
1.3 Subduction initiation at continental margins	18
1.3.1 <i>Scope of the Thesis</i>	18
1.3.2 <i>Thesis outline</i>	19
Chapter 2 - Methodology	22
2.1 Modelling approach	22
2.2 Numerical modelling	22
2.3 Analogue modelling	25
2.3.1 <i>Experimental procedure</i>	25
2.3.2 <i>Scaling theory</i>	26
2.3.3 <i>Brittle behaviour</i>	28
2.3.4 <i>Ductile behaviour</i>	29
Chapter 3 - Passive margin inversion controlled by stability of the mantle lithosphere	32
Highlights	32
3.1 Introduction	34
3.2 Rheology and geometry of passive continental margins	33
3.3 Experimental approach	34
3.3.1 <i>General concept</i>	34
3.3.2 <i>Materials properties and scaling</i>	36
3.3.3 <i>Analysis and quantification of vertical movements</i>	37
3.3.4 <i>Assumptions and simplifications</i>	38
3.4 Experimental results	39
3.4.1 <i>Weak mantle lithosphere (Set 1)</i>	39
3.4.2 <i>Medium strength mantle lithosphere (Set 2)</i>	42

3.4.3 <i>Strong mantle lithosphere (Set 3)</i>	44
3.5 Discussion.....	46
3.5.1 <i>Controlling parameters for passive margin instability</i>	46
3.5.2 <i>Comparison with numerical modelling</i>	50
3.5.3 <i>Length scales of deformation during passive margin inversion</i>	51
3.6 Conclusions.....	54
Appendix A.....	55
Chapter 4 - Strain localization mechanisms for subduction initiation at passive margins	58
Highlights	58
4.1 Introduction.....	58
4.2 General concept and modelling setup	59
4.2.1 <i>Analogue modelling specificities</i>	61
4.2.2 <i>Numerical modelling specificities</i>	61
4.2.3 <i>Combined approach</i>	65
4.3 Analogue modelling results	68
4.3.1 <i>Effect of lithospheric mantle ductile strength</i>	68
4.3.2 <i>Effect of brittle strength of the oceanic lithosphere</i>	69
4.3.3 <i>Effect of the thermal age of oceanic lithospheres</i>	69
4.4 Numerical modelling results	73
4.4.1 <i>Effect of the thermal age of oceanic lithospheres:</i>	73
4.4.2 <i>Effect of crust-mantle coupling</i>	76
4.4.3 <i>Effect of Peierls creep</i>	76
4.4.4 <i>Effect of shear heating</i>	76
4.5 Discussion.....	77
4.5.1 <i>Mechanisms of strain localization for subduction initiation</i>	78
4.5.2 <i>Role of thermal softening and Peierls creep</i>	80
4.5.3 <i>Timing and locus of subduction initiation</i>	81
4.6 Conclusions.....	82
Appendix B.....	83

Chapter 5 - Influence of magma-poor versus magma-rich passive margins on subduction initiation	86
Highlights	86
5.1 Introduction.....	86
5.2 Modelling approach	88
5.2.1 <i>Methods</i>	88
5.2.2 <i>Modelling strategy</i>	89
5.2.2 <i>Model limitation</i>	94
5.3 Reference model results.....	94
5.3.1 <i>Role of crust-mantle coupling</i>	94
5.3.2 <i>Role of the passive margin length</i>	96
5.3.3 <i>Timing of subduction initiation</i>	96
5.4 Application to magma-poor and magma-rich passive margins.....	97
5.4.1 <i>Subduction initiation at magma-poor passive margins</i>	97
5.4.2 <i>Subduction initiation at magma-rich passive margin</i>	99
5.5 Discussion.....	99
5.5.1 <i>Comparison to previous models</i>	99
5.5.2 <i>Application to observations and reconstructions of subduction initiation near the former Adriatic passive continental margins</i>	102
5.5.2.1 <i>Subduction initiation along the former Adriatic MPPM presently exposed in the Alps</i>	102
5.5.2.2 <i>Subduction initiation along the former Adriatic MRPM presently exposed in the Dinarides-Hellenides</i>	104
5.6 Conclusions.....	106
Appendix C.....	107

Chapter 6 - Stress dependence of passive margin inversion: a key factor controlling subduction initiation	112
Highlights	112
6.1 Introduction.....	112
6.2 Stress magnitudes in oceanic basins	113
6.3 Numerical model.....	114
6.3.1 <i>Modelling approach</i>	115

6.3.2 <i>Model geometry and rheology</i>	115
6.3.3 <i>Investigated parameters</i>	117
6.3.4 <i>Boundary and loading conditions</i>	117
6.3.5 <i>Model output and stress analysis</i>	119
6.4 <i>Modelling results</i>	120
6.4.1 <i>Oceanic lithosphere with laterally uniform thickness</i>	120
6.4.2 <i>Mechanisms for subduction initiation</i>	122
6.4.3 <i>Oceanic lithosphere with age dependant thickness</i>	124
6.5 <i>Discussion</i>	127
6.5.1 <i>Favourable conditions for subduction initiation at passive margins</i>	127
6.5.2 <i>Model limitations</i>	129
6.5.3 <i>Subduction initiation in the Wilson cycle</i>	130
6.6 <i>Conclusion</i>	131
Chapter 7 - Concluding remarks	134
7.1 <i>Mechanisms and evolution of induced subduction initiation at passive continental margins</i>	134
7.2 <i>Strain localization and propagation of deformation in the mantle lithosphere</i>	135
7.3 <i>Favourable parameter combination for subduction initiation at a passive margin</i>	136
7.4 <i>Subduction initiation in the Alpine system</i>	138
7.5 <i>Future perspectives</i>	139
Acknowledgements	142
References:	144

Summary

Subduction zones drive plate tectonics and are major sites of natural hazard generation such as volcanism, earthquakes and tsunamis. Therefore, understanding their formation and evolution is of paramount importance for both science and society. To date, it is still unclear which processes are involved in the nucleation of subduction zones, which parameters are controlling, and which tectonic settings are prone to the initiation of a new subduction zone. Uncertainties arise from the scarcity of observational constraints of modern and ancient subduction zones precluding the unambiguous identification of the initiation processes. In this PhD thesis a novel combination of state-of-the-art physical analogue and thermo-mechanical numerical modelling (chapter 2) is used for exploring geometric, kinematic and rheologic parameters that favour the initiation of new subduction zones. Our study focuses on the unresolved question of subduction zone formation at passive continental margins and aims at identifying and quantifying the deformational response of the ocean-continent transition to plate convergence to arrive at diagnostic features that can be tied to the subduction initiation process. The modelling results are validated by comparison with deformation structures from modern inverted passive margins and the geological record of past natural examples, which are now part of orogenic systems within the Tethys realm.

The results of our modelling study reveal that subduction initiation at passive continental margins critically depends on the buoyancy and strength of the oceanic lithosphere and requires failure of the load-bearing crustal and mantle layers. As such, upon orthogonal convergence subduction initiation is controlled by the strength contrast at the ocean-continent transition as well as the strength of the mantle lithosphere at the margin.

In chapter 3, 3d-analogue modelling is used to describe the mode and length-scale of deformation at a passive continental margin and adjacent oceanic and continental lithospheres as a function of lithosphere mantle viscosity and strength. Our experiments portray an evolution in which early-stage deformation at the ocean-continent transition is controlled by the ductile lower crust of the passive margin lithosphere. Associated deformation involves flexure of the oceanic plate and folding of the passive continental margin over a characteristic length scale of 500-600 km when scaled to nature. In the next stage, shear localization through the formation of a decollement within the crust favours underthrusting of the oceanic lithosphere, leading to a reduction of the area affected by deformation to ~ 200 -300 km. Subsequently, underthrusting is followed by the development of a subducting slab. The detailed description of incipient deformation is compared to modern Earth, where contractional deformation of passive margins may resemble early stages of induced subduction initiation.

Building on the results of chapter 3, chapter 4 focuses on mechanical instabilities at both crustal and mantle lithospheric levels that are required for the development of a subduction plate boundary. Different to previous studies, our combined

analogue and numerical modelling approach does not require pre-existing weak zones to localize deformation but deploys shear heating driven thermo-mechanical feedback mechanisms associated with shear zone development within Peierl's domain. The modelling results show that the locus of subduction is a function of a complex interplay between the strength of the of the oceanic and the continental lithospheres, regulated by their thermo-tectonic age, as well as crust-mantle coupling at the passive continental margin. Suitable mechanical conditions for subduction initiation at passive continental margins correspond to an intermediate thermal age (ca. 50-100 Myr) of the oceanic lithosphere. Younger oceanic lithosphere leads to intra-oceanic deformation, whereas old oceanic leads to deformation of the continental lithosphere.

Central to chapter 5 are the rheological stratification of magma-poor and magma-rich passive margins for the initiation process. We show that in case of magma-poor, hyper-extended margins, crust mantle decoupling facilitated by exhumed and serpentized mantle lithosphere is key for localizing deformation at the base of the passive margin crust below the necking zone. Subsequent propagation of shear deformation into the mantle lithosphere is controlled by shear-heating induced softening mechanisms. In contrast, magma-rich passive continental margins are characterized by strong crust-mantle coupling leading to transfer of deformation into the oceanic domain, where a new subduction zone evolves. These modelling results are particularly consistent with the geologic record from the Alps and Dinarides/Hellenides orogenic systems, where contrasting types of rifted margins, (magma-poor in the Alps and magma-rich in the Dinarides-Hellenides), of the same Adriatic micro-continent have been facing the Alpine Tethys and Neotethys oceans, respectively. In the Alps subduction initiation occurred at the passive margin, whereas that of Dinarides-Hellenides was intra-oceanic.

Additionally, in chapter 6, we discuss the relationship between the compressional stresses acting on the lithosphere and the strength of the mid-oceanic ridges that varies through time by cooling. We highlight the importance of convergence rate that controls stress regulation in the oceanic lithosphere. We find that slow convergence favours subduction initiation at passive margins over subduction at mid-oceanic ridges. Such kinematic conditions are consistent with the slow Africa – Europe convergence between 130 and 85 Ma reflecting suitable boundary conditions for subduction initiation in the Alps.

Samenvatting

Subductiezones zijn niet alleen de drijvende krachten van plaattektoniek, maar ook de plekken waar gevaarlijke verschijnselen zoals vulkanisme, aardbevingen en vloedgolven gegenereerd worden. Het is daarom van cruciaal belang om het ontstaan en de ontwikkeling van subductiezones goed te begrijpen. Maar zelfs vandaag de dag is nog onbekend welke processen, parameters en tektonische omstandigheden een rol spelen in de initiatie van nieuwe subductiezones. De vele onzekerheden die een eenduidige beschrijving van de initiatie van subductiezones onmogelijk maken, vloeien voort uit een gebrek aan gegevens en waarnemingen van huidige en vroegere subductiezones.

In dit promotieonderzoek is een nieuwe manier van gecombineerde analoge-numerieke modellering ontwikkeld en gebruikt voor een studie naar de geometrische, kinematische en reologische parameters die initiatie van subductiezones bevorderen. Dit onderzoek richt zich met name op de vorming van subductiezones in passieve continentale marges, en daarbij specifiek op de deformatie van de oceaan-continent overgang ten gevolge van plaatconvergentie. De modelresultaten zijn geverifieerd door ze te vergelijken met de deformatie van zowel hedendaagse geïnverteerde passieve marges als vroegere subductiezones, waarbij deze deel uitmaken van orogenen in het Tethys domein.

De resultaten van de modelstudie laten zien dat initiatie van subductie in passieve marges afhangt van het drijfvermogen en sterkte van de oceanische lithosfeer, en dat breken van de korst en mantellagen vereist is. Bij loodrechte convergentie wordt initiatie van subductie daarbij bepaald door zowel het sterktecontrast van de oceaan-continent overgang als de sterkte van de mantellithosfeer.

Hoofdstuk 3 presenteert analoge modelstudies van het type deformatie en de lengteschaal daarvan in passieve marges en naastgelegen oceanische en continentale lithosfeer als functie van de viscositeit en sterkte van de lithosferische mantel. De experimenten tonen aan dat vroegtijdige deformatie van de ocean-continent overgang gecontroleerd wordt door de ductiele onderkorst van de passieve marge. De bijbehorende deformatie bestaat uit buiging van de oceanische plaat en plooiing van de continentale passieve marge, met een opgeschaalde karakteristieke lengte van 500-600 km. In de volgende fase ontwikkelt zich een decollement in de korst waarlangs de oceanische lithosfeer kan onderschuiven, en waardoor de lengte van de deformatiezone afneemt tot ~200-300 km. De onderschuivende lithosfeer ontwikkelt zich tenslotte tot een subducerende plaat. De gedetailleerde beschrijving van beginnende deformatie in de schaalmodellen is vergeleken met hedendaagse passieve marges op aarde waar deformatie lijkt op beginnende fases van subductie initiatie.

In Hoofdstuk 4 ligt de focus op instabiliteiten in zowel de korst als mantellithosfeer, die nodig zijn voor de ontwikkeling van een subducerende plaatgrens. In tegenstelling tot in andere modelstudies is er in de in de nieuwe gecombineerde analoge-numerieke modelprocedure geen noodzaak voor de aanwezigheid van eerder bestaande zwaktezones om deformatie te concentreren, maar wordt dit gerealiseerd door thermo-mechanische verzwakking ten gevolge van warmteontwikkeling in een zich

vormende schuivingszone in het reologische Peierl's domein. De modelresultaten tonen aan dat de locatie van subductie bepaald wordt door enerzijds de sterktes van de oceanische en continentale lithosfeer, die beide afhangen van hun thermo-tektonische ouderdom, en anderzijds de koppeling tussen korst en mantel van de passieve marge. De thermo-mechanische vereisten nodig voor initiatie van subductie in passieve marges passen het best bij een oceanische lithosfeer met een middelbare thermische ouderdom (ca 50-100 Myr). Verkorting van jongere oceanische lithosfeer zou resulteren in intra-oceanische deformatie, terwijl oudere oceanische lithosfeer leidt tot deformatie van de continentale lithosfeer.

In Hoofdstuk 5 wordt gekeken naar de rol van reologische stratificatie van magma-arme en magma-rijke passieve marges in het subductie initiatieproces. In het geval van magma-arme, hyper-uitgerekte marges blijkt dat door exhumatie en serpentinisatie van de mantel lithosfeer de korst en mantel ontkoppeld zijn, waardoor bij inversie de convergente deformatie zich concentreert onderin de korst onder de zone van maximale uitrekking. Verdere deformatie propageert vervolgens in de mantel lithosfeer die verzwakt is door shear heating processen. Magma-rijke marges worden daarentegen gekarakteriseerd door een sterke koppeling tussen korst en mantel, waardoor convergente deformatie verplaatst wordt naar oceanisch gebied, alwaar zich een nieuwe subductiezone ontwikkelt. De verkregen modelresultaten komen met name goed overeen met geologische gegevens van orogene systemen in de Alpen en de Dinariden/Helleniden, waarbij tegengestelde typen passieve marges (magma-arm in de Alpen en magma-rijk in de Dinariden/Helleniden) gelegen waren op hetzelfde Adriatische micro-continent, grenzend aan respectievelijk de Alpiene Tethys en de Neotethys oceanen. In de Alpen vond subductie plaats in de passieve marge, terwijl in de Dinariden/Helleniden de subductie intra-oceanisch was.

Tenslotte wordt in Hoofdstuk 6 de relatie besproken tussen enerzijds de compressieve spanningen in de lithosfeer en anderzijds de thermo-mechanische sterkte van de oceanische lithosfeer, die door afkoeling toeneemt met de tijd. Van kritisch belang hierbij is ook de convergentiesnelheid, die bepalend is voor de spanningsverdeling in de oceanische lithosfeer. Een analyse van modelresultaten laat zien dat langzame convergentie gunstiger is voor initiatie van subductie in passieve margins dan bij mid-oceanische ruggen. Dergelijke kinematische condities waren aanwezig tijdens de trage Afrika-Europa convergentie tussen 130 en 85 Ma, en vormden dus geschikte randvoorwaarden voor initiatie van subductie in de Alpen.

Chapter 1

Chapter 1

Introduction

1.1 Mechanisms of subduction initiation

Subduction zones are important agents of the dynamic Earth and their analysis provides critical insight in short- and long-term deformation processes including the seismic cycle and plate tectonics, respectively. Although fundamental to geohazard characterization and prediction as well as plate tectonic theory, it is not fully clear how subduction systems start. Previously published subduction initiation mechanisms and their favourable dynamic, kinematic and rheologic conditions remain controversial due to the limited availability of observations in modern and ancient natural examples.

According to plate tectonic theory, the subduction process involves the sinking of oceanic lithosphere driven by buoyancy forces arising from density contrasts (Burke et al., 1977; Vlaar & Wortel, 1976; Wilson, 1968). It follows that passive continental margins, where density contrasts across the ocean-continent transition are large, should be favourable sites for subduction initiation (e.g. Dewey, 1969). However, previous studies (Cloetingh et al., 1982, 1989; McKenzie, 1977; Molnar & Gray, 1979; Oxburgh & Parmentier, 1977) have shown that subduction initiation at passive continental margins is physically difficult to achieve without (external) forcing, because the rapid cooling and associated strengthening of the mantle lithosphere prevents failure (Burov, 2011; Cloetingh et al., 1984; Mueller & Phillips, 1991).

Modern concepts of subduction initiation suggest that subduction zones initiate both at continental margins or in intra-oceanic or intra-continental domains, driven by body forces or by external forcing often referred to as “spontaneous” or “induced” subduction initiation (Stern, 2004). Spontaneous subduction initiation is characterized by gravitational collapse facilitated by weak, pre-existing structures and simultaneous

Figure 1.1: Cartoons showing end-member models of intra-oceanic subduction initiation. (a) spontaneous subduction initiation at transform faults (inspired from Stern, 2004); (b) induced subduction initiation controlled by a pre-existing weak zone (inspired from Guilmette et al., 2018).

extension of the overriding fore-arc plate with intense alkaline magmatism (Fig. 1.1a, Stern et al., 2012). This mechanism of subduction initiation is relevant for oceanic transform / fracture zones or in mantle plume settings, where strong lateral density and strength contrasts exist (Cloetingh et al., 2021). The Izu-Bonin-Mariana system is probably the best documented case of spontaneous subduction initiation along a fracture zone (Shervais et al., 2019 and reference therein).

Induced nucleation of subduction requires continuous plate convergence, presumably accommodated at inherited, favourably oriented weak structures, until subduction eventually becomes self-sustained. Forced convergence results in a compressive stress regime at the evolving subduction interface and early uplift of the forearc with limited magmatism (Fig. 1.1b, Gurnis et al., 2004; Guilmette et al., 2018), with the Puysegur convergent system considered as a modern natural analogue (House et al., 2002). We note that subduction initiation at passive margins includes all stages from their inversion to the development of shear structures along which the underthrusting of the down going plate takes place, and ultimately, the formation of a subduction plate boundary (Nikolaeva et al., 2010).

Although the classification of Stern (2004) is debated (Arcay et al., 2020; Guilmette, 2018), the distinction in fundamental mechanisms facilitates the quantitative description of these processes and allows for comparing the modelling results with natural observations (Gurnis, 2004; Hall, 2019; Stern & Gerya, 2018).

Given existing uncertainties on the mechanical conditions for subduction initiation, several studies have been conducted aiming at getting further insight in the driving forces, processes and parameters that favour the formation of subduction zones. These include sedimentary loading and the lubricating effect of water (Cloetingh et al., 1984, Regenauer-Lieb et al., 2001, Nikolaeva et al., 2010), variations of gravitational potential energy in the adjacent continental topography (Pascal & Cloetingh, 2009; Schmalholz et al., 2019), subduction at fracture zones and pre-existing faults (Arcay et al., 2020; Cramer & Kaus, 2010; Gerya et al., 2008; Toth & Gurnis, 1998), density contrast in the lithospheric mantle (Chemenda et al., 1996; Faccenna et al., 1999; Goren et al., 2008; Mart et al., 2005; Nikolaeva et al., 2010), structural heterogeneity of a continental margin (Lévy & Jaupart, 2012; Rolf & Tackley, 2011; Ulvrova et al., 2019), large mantle downwelling (Baes et al., 2018), grain size reduction (Ricard & Bercovici, 2009), post-glacial rebound stresses (Grollmund & Zoback, 2003), small scale convection (Solomatov, 2004), plume induced subduction-initiation (Burov & Cloetingh, 2010; Cloetingh et al., 2021; Cramer & Tackley, 2016; Gerya et al., 2015; Ueda et al., 2008), subduction zone migration/invasion (Duarte et al., 2013; Munch et al., 2020; Waldron et al., 2014; Zhou et al., 2020), or the inversion of spreading ridges (Beaussier et al., 2019; Gülcher et al., 2019b; Maffione et al., 2015).

1.2 Rheology of passive continental margins

The cumulative length of Earth's passive continental margins, hereafter referred to as "passive margins", is in the order of 105,000 km (Bradley, 2008). Passive margins exhibit fundamental differences in structural style, geometry and subsidence histories.

With reference to magmatism, passive margins have been classified into magma-rich and magma-poor margins (Fig. 1.2a). Magma-rich passive margins are usually thicker, largely intruded by magmatic rocks and are underlain by mafic underplates (Geoffroy, 2005), as observed, for example, in the east Greenland or Uruguayan rifted margins (Fig. 1.2b, Clerc et al., 2018). In contrast, magma-poor passive margins are generally characterised by a steep necking domain, a lack of magmatic intrusives and extrusives and the exhumation of the mantle lithosphere at the base of the continental crust (e.g. Manatschal & Müntener, 2009). The conjugate margins of Newfoundland and Iberia are often used as the typical examples (Roberts & Bally, 2012, among others). However, passive margins that completely lack magmatism are rare (Franke, 2013), and several rifted margins exhibit a transitional stage between these two end-members (Fig. 1.2a).

The rheology of passive margins and neighbouring oceanic domains is complex due to inherent lateral changes in thickness, composition, structure and temperature at the transition between continental and oceanic lithospheres. Further complexities can be inherited from the geometric and thermo-rheological characteristics of the pre-rift lithosphere, but also from deep thermal and surface processes occurring during the syn- and post- rift stages (Buck, 1991; Osmundsen & Redfield, 2011). Regardless their magmatic budget, passive margins continental crust wedge out (Figs. 1.2b, c) due to syn-rift lithospheric thinning. The width of thinning area varies anywhere from 100 to 600 km (e.g. Osmundsen & Redfield, 2011; Reston & Manatschal, 2011) averaging ~200 km (Bauer et al., 2000). Furthermore, passive margins mark the rheological transition from a single-layer oceanic to a multi-layer continental lithosphere. Depending on the type of rheology and thermal structure, mechanical decoupling may occur between the various layers, such as between the crust and mantle, thereby reducing the overall strength at the margin lithosphere (Burov, 2011).

Due to the high strength contrast in the transition between the continental and oceanic lithosphere, higher compressional stresses are usually observed at passive margins where deformation at shallow levels is well constrained by seismicity, geophysical and geological studies. This stress concentration is reflected in the reverse fault-plane solutions of seismicity and tectonic inversion, as for instance in the passive margins of SE Brazil and Argentina (Assumpção, 1998), Iberia (Masson et al., 1994; Alvarez-Marron et al., 1997), the British Isles (Ziegler et al., 1995), Brittany (Bonnet et al., 2000), NW African passive margin (Robertson & Stillman, 1979), Norway (Doré & Lundin, 1996), northern, or southern Australia (Keep & Harrowfield, 2008), Arctic Canada (Stephenson & Cloetingh, 1991), and the north-eastern Virginia escarpment (Mixon & Newell, 1977) (Fig. 1.2).

Although crustal deformation is common at passive margins, it is not clear how the underlying mantle reacts to the applied forces as earthquakes below 40km are absent (Watts & Burov, 2003). Deformation in the lithospheric mantle requires very high stress levels or weak rheologies (Cloetingh et al., 1982; Kruse & Royden, 1994; Watts & Burov, 2003). Therefore, analogue and numerical simulations frequently implement weak zones or sharp boundaries between different model domains to simulate compositional heterogeneities and localize deformation (Mart et al., 2005; Nikolaeva et

Figure 1.2: (a) Map showing the distribution of major passive continental margin types (modified after Armitage et al., 2011 and Clerc et al., 2018), and the locations of tectonically inverted passive margins during the Cenozoic (magenta stars, Johnson et al., 2008; Pedoja et al., 2011; Yamato et al., 2013; Ziegler & Cloetingh, 2004). Continental shelves (dark grey shading) and plate boundaries (thin lines) are from Müller et al. (2008). Black areas correspond to large igneous provinces. (b) Example of a magma-rich passive continental margin, modified after Clerc et al. (2015). (c) Example of a magma-poor passive continental margin, modified after Sutra et al. (2013).

al., 2010; Shemenda, 1992; Zhong & Li, 2020). In the case of passive margins, compositional heterogeneities may be present in the lithospheric mantle due to re-fertilization (Müntener et al., 2004), rejuvenation by sustained plume activity (Guo et al., 2014; Moyen et al., 2017), or delamination of dense deep lithosphere (Hu et al., 2018). Under such conditions, the rheological contrast between the depleted and undepleted mantle lithosphere at the ocean-continent transition may enable the development of large structures facilitating underthrusting of the subducting plate (Gerya et al., 2015; Krystopowicz & Currie, 2013). Although the existence of a weak domain may be justified by assuming such thermo-compositional mechanisms, in nature,

heterogeneities are more distributed throughout the mantle lithosphere (Maystrenko et al., 2013). Consequently, implementing a priori lithospheric-scale faults is probably an oversimplification and more realistic thermo-mechanical driven localisation mechanisms also need to be considered. For instance, the combination of Peierls creep in olivine (Cramer & Kaus, 2010; Drury, 2005; Korenaga & Korenaga, 2016) with shear induced thermal softening (Duretz et al., 2014; Jaquet & Schmalholz, 2018; Kiss et al., 2020; Thielmann & Kaus, 2012) is a better representation of deformation in the mantle lithosphere.

1.3 Subduction initiation at continental margins

Despite remarkable progress in fundamental physical understanding of subduction processes, the mechanisms of subduction initiation at passive margins remains a recurrent and long-standing unresolved question. Although observational evidences from modern examples constraining the subduction nucleation process are inconclusive, inferences from the geological record like the Alps (Manzotti et al., 2014a), Ryukyu (Hall, 2019) and the eastern Pacific (McCarthy et al., 2018) suggest that subduction-initiation must have occurred at passive margins in the plate tectonics history.

Despite the lack of actual examples for subduction initiation at passive margins (Stern, 2004), 80% of the worldwide trenches are located at less than 200 km from continental margins (e.g. *East Pacific*, Ducea et al., 2015; *Ryukyu*, Yan & Shi, 2014; *Philippines*, Yumul Jr et al., 2003; *Zealandia*, Sutherland et al., 2001, Fig. 1.2a), while very few exist in open oceanic domain (*Izu-Bonin-Mariana*; Klemperer, 2003). Furthermore, the lack of subduction zones along Atlantic and Indian ocean margins underlines their stability over geological time-scales and suggests that externally driven processes such as subduction zone invasion are needed for the genesis of subduction zones at continental margins (Duarte et al., 2013; Zhou et al., 2020). Although this mechanism could explain natural examples (e.g. *western edge of the Pacific plate*, Hall, 2019; *South Sandwich*, van de Lagemaat et al., 2021), its spontaneous retreating mode of subduction (Zhou et al., 2020) contrasts with the geological record of subduction initiation at magma-poor continental margins such as the Pyrenees or the European Alps orogenic systems (McCarthy et al., 2018).

1.3.1 Scope of the Thesis

The overarching aim of this thesis is to identify and quantify favourable mechanical conditions for induced subduction initiation at passive continental margins. Key questions we address in this study are:

- What are the characteristic length-scales of shortening induced deformation at passive margins and how do these length scales change through time during the various deformation phases of the plate tectonics cycle from passive margin inversion to subduction?

- How are the style, sequence and locus of deformation related to the age and rheological stratification of the passive margin and the oceanic lithosphere and how does this influence the duration of the initiation process?
- How does deformation propagate from the margin crust into the mantle lithosphere and what are the associated localizing mechanism to form a continuous subduction plate boundary?

These fundamental questions are addressed in the present thesis through a novel combination of physical analogue and thermo-mechanical numerical modelling allowing for mapping the parameters involved in subduction initiation at passive margins. The key variables in this study are the initial rheological stratification of the oceanic and continental lithosphere, with an emphasis on the rheology of the mantle lithosphere, and the strength of magma-poor and magma-rich passive margins, their geometry, the rate of shortening and the sensitivity of the modelling results to weakening mechanisms in the mantle lithosphere. Our modelling results are compared to length scales of deformation at inverted passive margins and the geological record of past subduction zones, which are now part of orogenic systems such as the Alps or the Dinarides-Hellenides.

1.3.2 Thesis outline

The thesis consists of 7 chapters. Following the introduction, *Chapter 2* details the relevant numerical and analogue modelling methods and scaling approaches. Chapters 3 to 6 present novel combinations of physical analogue and numerical modelling of passive margins that analyse key parameter combinations relevant for characterizing deformation, including the localization mechanisms from the inversion of the margin to the formation of a self-sustaining subduction zone.

Chapter 3 describes the results of a series of analogue experiments that focus on the stability of passive margins as a function of the strength of the mantle lithosphere. A range of mantle viscosities subject to different convergence rates serve as a proxy to simulate contraction in passive margins of different ages, with the aim of describing and quantifying deformation and mantle stability using the buoyancy number. The presented experiments illustrate distinct differences in the mode and length-scale of deformation as a function of lithospheric mantle strength. The experimental results are furthermore compared with estimates of folding wavelengths from inverted margins with different thermo-mechanical ages observed in natural situations (Cloetingh & Burov, 2011).

Chapter 4 aims to explore the strain localization mechanism for subduction initiation at passive margins. Both, analogue and numerical modelling has been carried out in order to constrain the required first order thermo-mechanical conditions. We analyse the sensitivity of the locus of subduction nucleation in respect to the thermal age of the oceanic lithosphere and thermo-mechanical feedback mechanisms. The analysed strain localization mechanisms include the development of a decollement at the base of the continental crust, which permits a subsequent propagation of deformation into the mantle lithosphere through shear-heating. Domain diagrams allows for mapping the locus of subduction as a function of oceanic lithosphere age at the onset of shortening.

Chapter 5 focuses on the structure and rheological stratification of the passive margin and illuminates the role of crust-mantle coupling/ decoupling for the subduction initiation process. We integrate models of magma-poor and magma-rich rifted margins to accounts for the variability of crustal compositions. In particular we investigate the role of serpentinized mantle lithosphere, which is characteristic for magma-poor passive margins, for the subduction initiation process and compare these results with models simulating magma-rich passive margins. The relevance of the modelling results is then demonstrated for the Alps and Dinarides-Hellenides orogenic systems, where we show that contrasting types of rifted margins control the mode of subduction initiation.

Chapter 6 addresses the kinematic and stress conditions for subduction initiation at passive margins. We focus on the role of stress regulation and dissipation on subduction initiation by testing the stress stress–strain-rate response to convergence rate, and thermo-tectonic age of oceanic and continental lithospheres. The results highlight the importance of slow convergence rate that favours subduction initiation at passive margins over subduction at mid-oceanic ridges. The results allow a discrimination of geodynamic processes that fit conditions for subduction nucleation at passive margins, which is, yet again, relevant for the unique case of the Alps, where the kinematic evolution is well understood.

Finally, *chapter 7* integrates and summarizes the modelling results presented in chapters 3-6 and provides a synoptic view on key parameter combinations for subduction initiation (1) at passive margins and (2) in oceanic domain. We infer possible implications for past and present subduction systems and provide an outlook for further research.

Chapter 2

Methodology

2.1 Modelling approach

In this thesis, we deployed a novel approach combining physical analogue and numerical modelling techniques for the study of subduction initiation processes. The analogue modelling provides high-resolution 3D outputs of crust and lithosphere deformation geometries and associated vertical motions, which facilitates the interpretation of complex crust and lithosphere structures and their lateral variability. Additionally, 2d-numerical, thermo-mechanical modelling allows for exploring temperature driven feedback mechanisms. To ensure a meaningful complementary use of both methods we scaled the experiments to obtain, within limits, similar geometrical and mechanical conditions at the onset of shortening. Through both approaches, the models simulate contraction of an ocean-continent transition as a consequence of far-field forces. Therefore, our models are relevant for induced subduction initiation (Stern, 2004). Although, the models described in the various chapters of the thesis serve a different purpose and thus differ in terms of initial geometric, rheologic and kinematic setups, they all involve three different domains: an oceanic, a continental and a wedge shape transitional lithosphere representing the passive continental margin (Fig. 2.1a). In the following sections we describe the basic characteristics of the analogue and numerical modelling setups and provide in case of analogue modelling an overview of the material properties. Details on the rheological and kinematic parameters used in this thesis are listed in the pertinent chapters of the thesis as these are not the same for all models.

2.2 Numerical modelling

The numerical modelling was performed to simulate the shortening of a visco-elastic-plastic lithosphere. We have employed the two-dimensional (2D) thermo-mechanical finite-difference, marker-in-cell code (MDoodz; Duretz et al., 2016a). The presented models were obtained by solving the conservation equation for a steady state momentum, temperature evolution and incompressible mass conservation equations:

$$\frac{\partial \tau_{ij}}{\partial x_j} - \frac{\partial P}{\partial x_i} = -\rho g_i \quad (2.1)$$

$$\rho c_p \frac{DT}{Dt} = \frac{\partial}{\partial x_i} \left(k \frac{\partial T}{\partial x_i} \right) + Q_r + Q_{SH} \quad (2.2)$$

$$\frac{\partial v_i}{\partial x_i} = 0 \quad (2.3)$$

where v is the velocity vector, T is the temperature, k is the thermal conductivity, ρ is the density, c_p is the heat capacity, Q_r is the radiogenic heat production, τ is the deviatoric stress tensor, $\dot{\epsilon}$ is the deviatoric strain rate tensor, P is the pressure and g is the gravity

Figure 2.1: Model configuration showing first-order rheological layering and boundary conditions at $t = 0$. The left and right panels depict the corresponding strength and temperature profiles for the continental (red) and oceanic (blue) lithospheres, respectively. The model incorporates velocity boundary conditions at the left/right boundary of the models ($+V_{in}$) and an outflow ($-V_{out}$) distributed at the base of the model and on the sides from 200 km depth to the base of the box.

acceleration vector. The term $Q_{SH} = \tau_{ij} \dot{\epsilon}_{ij}$ describes the production of heat by viscoplastic dissipation (shear heating).

The density field evolves according to the following equation of state:

$$\rho = \rho_0 (1 - \alpha(T - T_0))(1 + \beta(P - P_0)) \quad (2.4)$$

where ρ_0 is the reference density, α is the thermal expansivity, β is the compressibility, T_0 and P_0 are the reference temperature and pressure which were respectively set to 0 °C and 10^5 Pa.

We employ an additive decomposition of the deviatoric strain rate and a visco-elasto-plastic rheological model such that:

$$\dot{\epsilon}_{ij} = \dot{\epsilon}_{ij}^v + \dot{\epsilon}_{ij}^e + \dot{\epsilon}_{ij}^p \quad \text{where} \quad \dot{\epsilon}_{ij}^v = \dot{\epsilon}_{ij}^{dis} + \dot{\epsilon}_{ij}^{Peierls} \quad (2.5)$$

where the v , e , and p superscripts correspond to viscous, elastic and plastic portions and the superscripts dis and $Peierls$ refers to the dislocation and Peierls creep mechanisms.

The dislocation creep strain rate is computed as:

$$\dot{\epsilon}_{ij}^{dis} = \dot{\epsilon}_{II}^{dis} \frac{\tau_{ij}}{\tau_{II}} = A \left(2f e^{\frac{Q}{nRT}} \right)^{-n} \tau^n \frac{\tau_{ij}}{\tau_{II}} \quad (2.6)$$

where A is a pre-factor, Q is the activation energy, n is the stress exponent, R is the universal gas constant and f is a correction factor (Schmalholz & Fletcher, 2011). The subscripts II stand for the square root of the second tensor invariant. The elastic strain rate is written as:

$$\dot{\epsilon}_{ij}^e = \frac{\dot{\tau}_{ij}}{2G} \frac{\tau_{ij}}{\tau_{II}} \quad (2.7)$$

where G is the shear modulus (set to $3.5 \cdot 10^{10}$ Pa).

The plastic strain rate takes the form of:

$$\dot{\epsilon}_{ij}^p = \dot{\epsilon}_{II}^p \frac{\tau_{ij}}{\tau_{II}} \text{ for } \dot{\epsilon}_{II}^p = \dot{\epsilon}_{II} - \frac{\dot{\tau}_{II}}{2G} - \frac{\tau_y}{2\eta^v} \text{ and } \tau_y = C \cos \varphi + P \sin \varphi \quad (2.8)$$

where φ is the friction angle and C is the cohesion. The friction angle is set to 30° for the mantle lithosphere and asthenosphere. The cohesion is set to 20 MPa for all lithologies. We do not apply any plastic strain softening.

In the mantle lithosphere, the Peierls mechanism is also activated and its strain rate is computed as:

$$\dot{\epsilon}_{ij}^{Peierls} = \dot{\epsilon}_{II}^{Peierls} \frac{\tau_{ij}}{\tau_{II}} \quad (2.9)$$

where the effective strain rate is spelled as:

$$\left\{ \begin{array}{l} \dot{\epsilon}_{II}^{Peierls} = (2A^{Peierls})^{-s} \frac{\tau_{II}}{\tau_{II}} \\ A^{Peierls} = f\gamma\sigma^{Peierls} \left(E^{Peierls} e^{-\frac{(1-\gamma)^2 Q^{Peierls}}{RT}} \right)^{\frac{-1}{s}} \\ s = \frac{Q^{Peierls}}{RT} (1-\gamma)^{(q-1)q\gamma} \end{array} \right. \quad (2.10)$$

where the parameters s is the effective stress exponent (T -dependent), $Q^{Peierls}$ is the activation energy ($=540$ J/mol), $\sigma^{Peierls}$ is the Peierls stress ($=8.5 \cdot 10^9$ Pa), $E^{Peierls}$ ($=5.7 \cdot 10^{11}$ s $^{-1}$), q ($=2.0$), and γ ($=0.1$) (Evans & Goetze, 1979). Peierls creep stress is computed using a regularised formulation (Kameyama et al., 1999).

The temperature is kept constant at both the upper (0°C) and lower boundaries (1330°C) and the heat flow is set to 0 across the right and left boundaries. A constant inward normal velocity V_{in} is applied to the left/ right model boundaries of the model to simulate horizontal tectonic loading (Fig. 2.1). In order to satisfy mass conservation, an outflow velocity V_{out} is proportionally distributed at the base of the model and on the sides from 200 km deep to the base of the box. The temperature is kept constant at both the upper (0°C) and lower boundaries (1330°C) and the heat flow is set to 0 across the right and left boundaries. The shear stress is set to zero along the left, right and lower boundaries. The upper boundary is a true free surface that dynamically evolves with time and allows for a direct treatment of surface processes (Duretz et al., 2016). The equation governing the free surface evolution is written as:

$$\frac{Dh}{Dt} = \frac{\partial}{\partial x_i} \left(D \frac{\partial h}{\partial x_i} \right) + \dot{s} \quad (2.11)$$

where h is the height of the free surface, D is the topographic diffusivity (set 10^{-6} m 2 s $^{-1}$) and $\dot{s}$ is the sedimentation rate (set to 0.5 mm y^{-1} for $h < 0.0$). The initial temperature field is obtained by solving the steady state heat equation (neglecting shear heating) using reference thermal parameters excepted below the lithospheric mantle (at the base of thermal lithosphere h_L , Fig.2.1) where the conductivity was set artificially high in order to produce a quasi-adiabatic asthenosphere. The initial topography is set to 0 km.

The conservation equations are discretized using a finite difference/marker-in-cell technique (Gerya & Yuen, 2003). The global linearized systems of equations are

solved using a direct-iterative method (Räss et al., 2017). Non-linear iterations are used at both local and global levels. At the local level, Newton iterations ensure exact partitioning of strain rates and correct evaluation of effective viscosity (Popov & Sobolev, 2008; Schmalholz & Duretz, 2017). At the global level, Picard iterations are employed to best-satisfy mechanical equilibrium equations (to an absolute tolerance of 10^{-6} and within a maximum of 20 iterations).

2.3 Analogue modelling

2.3.1 Experimental procedure

Lithospheric-scale analogue models are used to examine large scale tectonic deformation processes. They are prepared inside a 36 cm wide, 45 cm long and 15 cm deep plexiglass box and consist of a stack of several layers made of sand and/or silicone putty, representing brittle and ductile layers of oceanic and continental lithosphere in nature (Fig. 2.2). The brittle (sand) and ductile (silicon putty) layers are floating on a low-viscosity fluid (Polytungstate +glycerol) that enables isostatic equilibrium throughout the experiments (e.g. Brun, 1999, 2002; Davy & Cobbold, 1991; Dombrádi et al., 2010; Sokoutis et al., 2000; Sokoutis et al., 2005). All layers are sequentially placed on top of the model asthenosphere. Ductile layers are prepared outside the box and then laid down on their substrate in one piece, whereas brittle layers are added by sifting from a height of ca. 30 cm.

All experiments consist of an “oceanic”, a “continental” and a “passive margin” domain (Fig. 2.2) representing a two-layer oceanic and a four-layer continental lithosphere, respectively. Characteristic to all models is the gradual ocean-ward thinning of the brittle and ductile continental crust defining the area of the passive continental margin. This geometry approximates the geometry of rifted margins with a thinned continental crust (Péron-Pinvidic & Manatschal, 2009) at the onset of convergence. During an experiment, the model layers are shortened in one direction through an advancing wall. All other sides are confined by fixed walls (Fig. 2.2a). The moving wall is driven by an electronic engine advances at a constant rate, which is scaled to slowly converging natural systems (~ 0.5 cm/yr).

Experiments are monitored using digital photography and high-resolution laser scanning of the top surface at regular time intervals. Subsequently, laser scanning data are used to calculate digital elevation models of the surface utilizing the “Surfer, version 16” software. Additionally, digital photography is used for particle image velocimetry (Thielicke & Stamhuis, 2014) utilising the Pivlab module within the “MATLAB, version R2018b” computer programme. At the end, each experiment has been sprinkled with water and has been frozen for a period of at least 12 hours to allow for cutting the model parallel to the transport direction and retrieve the structure of the model with depth.

2.3.2 Scaling theory

The experiments were scaled according to the principles of geometric, dynamic, kinematic and rheological similarity (Hubbert, 1937; Ramberg, 1981; Sokoutis et al.,

2000; Weijermars & Schmeling, 1986), which is a requirement for scale-models to be representative for the natural process they simulate. Therefore, analogue models represent small-scale replicas of natural systems where quantities describing the system behaviour are proportionally scaled down. Geometrical similarity requires angles to be similar in nature and model and the corresponding lengths being proportional, defined by a length-scale ratio L^* , where $*$ refers to model/nature ratio (e.g., $L^* = L_m/L_n$).

To respect dynamic similarities, relevant scaling relationships between the natural prototype and the analogue model requires that the average strength of the ductile layers are correctly scaled with respect to the strength of the brittle layers and the gravity forces (Brun, 1999; Hubbert, 1937; Ramberg, 1981; Weijermars & Schmeling, 1986).

Figure 2.2: (a) Illustration of the analogue modelling setup consisting of a four-layer continental lithosphere and a two-layer oceanic lithosphere resting on a high-density, low-viscosity fluid (mixture of Polytungstate and glycerol) representing the asthenosphere. The red arrow indicates the shortening direction that is applied via the moving wall. (b-c) Comparison of strength profiles calculated for the numerical (Fig. 2.1) and analogue modelling approach, which reflect initial rheological conditions. (b) strength profiles for oceanic and (c) for continental lithospheres. LM stands for lithospheric mantle.

In rock mechanics and structural geology, the scaling of stress is performed by using the Navier–Stokes equation of motion for a continuous medium.

$$\frac{\sigma_{ij}}{\partial x_j} + \rho(g_i - (a_i)) = 0 \quad (i, j = 1, 2, 3) \quad (2.12)$$

where ρ , g , t , h_d , are the density, gravitational acceleration, time and σ_{ij} , x_j , a_i are the components of stress, position and acceleration respectively.

As the processes investigated in this thesis occur at slow, plate-tectonic rates, effects of inertia forces (ρa_i) are very small and can be neglected (Ramberg, 1981). This justifies treating time and length as independent variables (Ramberg, 1981). Therefore, to respect rheological similarity, the stress ratio σ^* is determined by scaling the vertical normal stress throughout the box. It is given by the relationship:

$$\sigma^* = \rho^* g^* L^* \quad (2.13)$$

As all experiments have been performed under normal gravity conditions, g^* equals 1 making the stress-scale ratio dependent on the density and length-scale ratios, respectively.

The kinematic similarity is satisfied when the model and prototype undergo similar changes of shape or position within a certain amount of time set by the time ratio (t^*), which reflects the duration of deformation in the viscous layer. The time-scale ratio can be calculated using the inverse of the strain rate which is the viscosity ratio divided by the stress ratio:

$$t^* = \frac{1}{\dot{\epsilon}^*} = \frac{\eta^*}{\sigma^*} \quad (2.14)$$

where η^* is the model to nature ratio of the analogue material viscosity to the average viscosity of the natural prototype's ductile layer.

Finally, rheological similarity requires that analogue materials respond to forcing in similar way than natural prototypes, which implies that for brittle layers frictional properties must be similar in model and nature and that for ductile layers flow curves of analogue materials are similar to those of rocks they intent to simulate for the same stress level (e.g. Weijermars & Schmeling, 1986). To satisfy these conditions, the analogue models were scaled according to the stress envelopes obtained for the initial conditions of our numerical models using flow law parameters derived from laboratory experiments (i.e., plastic, exponential-dislocation-diffusion creep). Thus, for a given lithospheric thermal thickness, the strength and thickness of the ductile and brittle layers used in the analogue models approximate the strength and thickness of the respective layers used in our numerical models, facilitating the comparison of modelling results stemming from both approaches (Fig. 2.2). In the following sections we detail the fundamental scaling relations for brittle and ductile deformation.

2.3.3 Brittle behaviour

Brittle layers behave according to the Mohr-Coulomb relationship given by:

$$\tau = C + \sigma_n (\tan \phi) \quad (2.15)$$

where τ is the critical shear stress, C the cohesion, ϕ the angle of friction, and σ_n is the normal stress. According to Byerlee (1978), ϕ is $\sim 31^\circ$ for most crustal rocks with a cohesion $C = 20\text{-}50$ MPa. Dry feldspar and quartz sands with grain size ranging from 100-300 μm are used in this study as analogues of materials exhibiting Mohr-Coulomb behaviour (Fig. 2.3a).

Friction coefficients and cohesion of these sands have been determined based on ring shear tests (Fig. 2.3a), and yield friction coefficients (Table 2.1) that are similar to natural rocks underlining their suitability as rock analogues (Sokoutis et al., 2005; Willingshofer et al., 2018). Cohesion has been determined through extrapolation of these ring shear test data and shows a large spread, particularly for low stress levels (Ritter et al., 2016). These uncertainties do not allow for accurate scaling of cohesion, which is often considered to be negligibly small in lithosphere-scale analogue experiments (Davy & Cobbold, 1991). Strength envelopes show that the differential stress in brittle layers increases linearly with depth and is largely independent of strain rate (Fig. 2.2b).

Reformulating equation 2.15 for differential stress in a contractional regime is given by:

$$(\sigma_1 - \sigma_3)_{\text{Brittle}} = 2\rho gz \quad (2.16)$$

where σ_1 and σ_3 are the maximum and minimum principal stresses and z the depth (Brun, 1999).

Table 2.1: Summary of material properties. Granular material properties are from Willingshofer et al. (2018). Flow curves of the silicone putties are measured at room temperature with a conical-cylindrical viscometer at Utrecht University.

Layer	Material	ρ (kg/m ³)	C (Pa)	μ	A (Pa ⁿ s)	n
Brittle continental crust	Felspar sand	1300	~ 10	0.68		
Ductile continental crust	Rhodorsil gomme+ sand	1300			1.78e-5	1.12
Brittle mantle lithosphere	Quartz sand	1500	~ 10	0.63		
Ductile mantle lithosphere		1500				
strong	PDMS + BaSO4				2.33e-6	1.36
medium	Rhodorsil gomme +sand				1.35e-5	1.13
weak	Rhodorsil gomme +sand				5.51e5	1.09
Asthenosphere	Polytungstate +glycerol	1470			1.25e-3	1.00

Here, ρ is the density, A is the material viscosity factor and n the power-law exponent, C is the cohesion and μ is the friction coefficient, equal to $\tan \phi$.

2.3.4 Ductile behaviour

Mixtures of polydimethyl-siloxane polymer (PDMS) or Rhodorsil Gomme-type polymer with fillers (quartz sand or barium sulphate) have been used to simulate flow behaviour in ductile layers. These materials exhibit slightly non-Newtonian behaviour with n values ranging between 1.09 and 1.4 (Table 2.1, Fig. 2.3b). The strength of these materials is scaled by averaging the temperature-dependent differential stress of the natural prototype over the layer thickness (Figs. 2.2b, c). Although this deviates from temperature dependent creep strength in natural systems, it gives a good first order approximation of the total strength.

Following Jaeger et al. (2009) the differential stress for plane strain can be calculated by using the shear stress (equation 2.17). For non-Newtonian materials, shear stress obeys a power law flow law (equations 2.18 and 2.19), linking stresses to strain rate (or engineering strain rate, $\dot{\gamma}$), and material properties (e.g. Kirby, 1985; Weertman, 1978). The following formulations apply:

$$(\sigma_1 - \sigma_3)_{\text{Ductile}} = 2\tau \quad (2.17)$$

$$\tau^n = \frac{\dot{\gamma}}{A} \quad (2.18)$$

$$\dot{\gamma} = \frac{\partial v}{\partial y} \quad (2.19)$$

where n is the flow behaviour index. A is in this case the material viscosity factor for non-Newtonian rheologies obtained from shear tests and equation 2.19 expresses shear strain rate as the ratio of velocity (v) over a length scale (y). The exact value of the strain rate cannot be precisely calculated *a priori* as it depends on the number and location of faults that develop during model deformation (Midtkandal et al., 2013). For the ductile layers in the analogue experiments, the shear strain rate can be approximated as the ratio between the velocity V applied at the model boundary and the ductile layer thickness h_d . Therefore, to represent a ductile layer with a given thickness and viscosity, the differential stress is given by (Nalpas & Brun, 1993; Smit et al., 2003):

$$(\sigma_1 - \sigma_3)_{\text{Ductile}} = 2 \left(\frac{V}{h_d A} \right)^{\frac{1}{n}} \quad (2.20)$$

To scale the strength of the ductile layer in the analogue model, we utilise the rate dependence of the material, which allows for regulating the strength of a ductile layer of certain thickness through the displacement rate V .

To ensure that dynamic similarity is respected, we calculate the Reynolds number Re , which represents the ratio between inertial forces and viscous forces:

$$Re = \rho \cdot V \cdot h_d / \eta \quad (2.21)$$

where ρ , V , η , h_d , are the density, velocity of displacement, effective viscosity, and thickness of the ductile layers. In nature the Reynolds number is generally very small ($Re < 10^{-20}$) (Dombrádi et al., 2010; Ramberg, 1981; Schellart & Strak, 2016; Sokoutis et al., 2000). In the experiments, $Re \ll 1$ (e.g. for the ductile lithospheric mantle, $Re \sim$

$3e^{-10}$) allowing for treating the time and length ratios as independent variables (see also equation 2.13).

Following Faccenna et al. (1999), we calculate the buoyancy number (F) to quantify the negative buoyancy force of the oceanic lithosphere relative to the asthenosphere and its relation to resistive forces arising from the viscous lithosphere.

$$F = \frac{\Delta\rho g h_L}{\dot{\gamma} \eta_{lm}} \quad (2.22)$$

where $\Delta\rho$ is the density contrast between the lithosphere and the asthenosphere, h_L is the total thickness of the lithosphere, η_{lm} is the viscosity of the lithospheric mantle.

Figure 2.3: (a) Shear curves based on ring shear tests for feldspar and quartz sand subject to normal stresses of $\sigma_n = 500, 1000, 2000$, and 4000 Pa. “Peak strength” denotes the stress level at which failure occurs. “Dynamic strength” refers to a stable sliding regime and “reactivation strength” indicates the shear strength upon renewed shearing. For details of the measurement procedure see Lohrmann et al., 2003). (b) Flow curves of the silicone putties used in this study. Plot of shear strain rate vs. shear stress. Viscosities are measured with a conical-cylindrical viscometer. The curves define the power law function of the ductile material as defined in equation 2.18.

Chapter 3

This chapter is based on the following manuscript: Auzemery A., Willingshofer E., Sokoutis D., Brun J.P. Passive margin inversion controlled by stability of the mantle lithosphere. Submitted for publication in Tectonophysics.

Chapter 3

Passive margin inversion controlled by stability of the mantle lithosphere

Highlights

- Analogue models are used to quantify deformational and topographic response during subduction initiation at passive margins.
- Strength of the ductile lithospheric mantle is key for creating instabilities at passive margins.
- Prior to underthrusting, folding at passive margin is the primary response to shortening.

3.1 Introduction

Within the frame of the Wilson cycle (Wilson, 1968; Burke, 1976), passive continental margins will eventually be turned into active margins where new subduction zones emerge. The processes contributing to this change in dynamic conditions and their structural and morphologic expression are, however, not well understood. The lack of modern examples of subduction initiation along the passive margins of the Atlantic or Indian oceans, suggests that passive continental margins are stable features, promoting the idea that subduction zone invasion is required for the formation of subduction zones at continental margins (Crameri et al., 2020; Duarte et al., 2013; Zhou et al., 2020). Although observational evidences from modern examples constraining the subduction nucleation process are inconclusive, inferences from the geological record like the Alps (Manzotti et al., 2014a), Ryukiu (Hall, 2019) and the eastern Pacific (McCarthy et al., 2018) suggest that subduction initiation must have occurred at passive margins in the plate tectonics history.

Deformation at passive continental margins is usually attributed to two contrasting driving mechanisms: (a) body forces arising from density contrasts or (b) far field forces (Stern, 2004). In context of the nucleation of subduction zones, these distinct forcing conditions are referred to as “spontaneous” or “induced” mechanisms of subduction initiation (Stern, 2004; Stern & Gerya, 2018). While spontaneous subduction initiation is predominantly associated with extension of the overriding plate, induced subduction initiation occurs within a contractional tectonic regime and will thus lead to shortening of the passive continental margin (Stern, 2004). For either of these mechanisms, driving forces need to overcome resisting forces among which the shear strength of the lithosphere being of critical importance (e.g. Cloetingh et al., 1989; Mueller & Phillips, 1991).

Previous studies have shown that although passive continental margins are sites of significant density contrasts and thus high buoyancy forces (Vlaar & Wortel, 1976), the nucleation of a new subduction zone is unlikely to occur by spontaneous collapse because of the increasing shear resistance of the lithosphere related to cooling (Cloetingh et al., 1982; Mueller & Phillips 1991). This view is further supported by the lack of

modern examples of subduction initiation along the Atlantic or Indian ocean passive margins where additional forces stemming from ridge push or adjacent continental topography (Marques et al., 2013; Pascal & Cloetingh, 2009) act on the passive margin lithosphere suggesting that passive continental margins are stable features. Therefore, subduction initiation at passive margins requires failure of the load-bearing crustal and mantle layers under a contractional regime (Nikolaeva et al., 2010; Zhong & Li, 2019). In contrast, numerous examples of inverted passive margins (Johnson, 2008; Pedoja et al., 2011; Yamato et al., 2013; Ziegler & Cloetingh, 2004) suggest that subduction initiation at passive margins should predominantly be expressed by deformation of the load-bearing crustal and mantle layers within a contractional regime (Nikolaeva et al., 2010; Zhong & Li, 2019). At shallow levels, stress concentration leads to folding, reverse and thrust faulting and associated seismicity (Stein et al., 1989b; Yamato et al., 2013).

While deformation of the crust is well documented, it is unclear how the mantle lithosphere at passive margins reacts to applied stresses during passive margin inversion. From a mechanical perspective, the strength of the mantle lithosphere often inhibits its failure, calling for the presence of weak mantle lithospheres (Göğüş & Pysklywec, 2008; Li et al., 2016b) or pre-existing weak zones (Stern & Gerya, 2018 and references therein) to localize strain in the lithospheric mantle. From the studies quoted above, it follows that deformation at passive margins, potentially leading to the development of a subduction zone, is strongly related to the strength and the gravitational stability of both the crust and the lithospheric mantle (Boonma et al., 2019; Faccenna et al., 1999). Although natural examples show that deformation localizes at passive continental margins, it is difficult to predict if an inverted passive margin will eventually turn into a subduction zone.

Therefore, we use physical analogue modelling to predict and quantify the deformational and topographic response at passive margins. This allows for shedding light on characteristic length scales of deformation during passive margin inversion and subsequent formation of a new subduction zone. In particular, we seek to obtain new insights in the relationship between the strength of the mantle lithosphere and the locus and style of deformation at passive margins. As such the main variables in this study are the strength of the ductile lithospheric mantle, regulated through changes of mantle viscosity and convergence rate. We demonstrate the relevance of our modelling results by comparison with length-scales of deformation at natural examples.

3.2 Rheology and geometry of passive continental margins

The rheology and geometry of passive margins and neighbouring oceanic domains often exhibit significant lateral changes in thickness, composition, structure and temperature of continental and oceanic lithospheres (Reston & Manatschal, 2011). In general, passive margins are characterised by the wedging out of the continental crust (Fig. 3.1a) over the region where thinning of the continental lithosphere occurred. Depending on rheology and thermal structure, mechanical decoupling may occur between the various layers, such as between the crust and mantle, thereby reducing the overall strength at the margin lithosphere (e.g., Burov & Diament, 1995; Ziegler &

Cloetingh, 2004). As an example, integration of yield-stress-envelopes along a section of a magma-poor passive margin such as the Angolan margin shows a drop in strength at the transition from the oceanic to the continental crust (Fig. 3.1b). This strength minimum is related to the relatively low lithostatic pressure at the passive margin and low strength of the thinned and extended continental crust at the ocean-continent transition OCT (Fig. 3.1a). In the case of the Angolan margin, this low strength region is ca. 40-50 km wide. Continent-ward, the thickness of the lithosphere increases, and the continental lithosphere becomes stronger than the oceanic lithosphere (Fig. 3.1b). However, we emphasise that the layering of continental lithosphere may cause mechanical decoupling between the crust and the mantle, thereby reducing the overall strength at the margin (Burov, 2011; Beniest et al., 2018; Erickson, 1993; Leroy et al., 2004).

3.3 Experimental approach

3.3.1 General concept

Our experiments simulate shortening of a magma-poor passive margin as a consequence of far-field forcing representing the cumulative effect of forces arising from mid-ocean ridges (Stern, 2004), mantle flow (Baes et al., 2018), gravitational potential energy (Pascal & Cloetingh, 2009), or sedimentary loading (Cloetingh et al., 1984). The model geometry and properties of the materials are displayed in Figure 3.2 and in Table 3.1.

Six experiments are shown (Fig. 3.2b) in which different viscosities of the lithospheric mantle and shortening rates (0.5 and 1 cm/hr) applied at model boundaries are used to change the ductile strength of the lithospheric mantle, the main parameter of this study. We thus utilize the stress dependence of viscosity for addressing the effects of different mechanical states of the lithosphere at the onset of deformation. The analogue models were built in a tank as described in chapter 2 and consist of sand (brittle layers) and silicone putty layers (ductile layers) floating on a low-viscosity fluid representing the asthenosphere.

The initial geometric model setup consists of an oceanic, a continental and a wedge shape transitional lithosphere in between (Fig. 3.2a). The latter represents the passive margin, i.e., the necking zone where the continental crust has been thinned most during lithosphere extension. The wedge-shaped domain, separating oceanic from continental lithospheres will be referred to as the “margin”. The length of the model margin scales to a margin length of ~150 km in nature, which falls within the range of passive margin lengths (~200 km on average, Bauer et al., 2000).

Within this framework, the rheology of the oceanic lithosphere can be approximated by a two-layer system (brittle-ductile) where the brittle-ductile transition (BDT) is located at a depth of ca. 42 km (Burov, 2011). The scaled 70 km thick oceanic lithosphere represents the mechanically strong part of a 50-100 Ma old lithosphere (by using the mechanical base of the lithosphere, e.g. Burov, 2011). In contrast, the continental lithosphere is defined by a four-layer system (brittle-ductile-brittle-ductile)

(Burov & Diament, 1995). Key element is the decoupling of the continental crust from the underlying mantle lithosphere along a viscous layer that wedges-out towards the ocean-continent transition (Figs. 3.2a, c). In the models, the ductile lithospheric mantle layers are homogenous in thickness and composition under both the continental and oceanic crust. As such we do not impose pre-defined mantle structures in the lithospheric mantle, such as a weak zone, to localize the deformation but use the geological features inherent to the rheological stratification of a passive margin. For the three types of lithosphere used in the models, the brittle thickness is assumed to be constant. Even if a variation in viscosity with a constant brittle-ductile transition depth seems unrealistic, it allows to test buoyancy conditions as function of the strength of the viscous lithospheric mantle (see buoyancy number, chapter 2, section 2.3.4).

Figure 3.2: (a) Experimental setup: four-layer continental lithosphere and a two-layer oceanic lithosphere resting on a high-density, low-viscosity fluid representing the asthenosphere. The white arrow indicates the shortening direction that is applied via the moving wall. BCC, DCC, BCM, DM and BO denote brittle continental crust, ductile continental crust, brittle continental mantle, ductile continental mantle and the brittle part of the oceanic lithosphere, respectively. (b) Matrix showing the parameter space explored in this study. (c) Graph showing computed integrated strength profile under a compressional stress regime for the total lithosphere for models with intermediate viscosity in the lithospheric mantle. (d) Strength profiles for oceanic lithosphere (left) and the continental lithosphere (right) at the onset of deformation. The strength profiles are calculated based on the material parameters given in Table 3.1.

3.3.2 Materials properties and scaling

Brittle layers exhibiting Mohr-Coulomb behavior are represented by layers of dry feldspar and quartz sand whereas mixtures of silicone putty (PDMS and Rhodorsil Gomme-type) with fillers represent ductile layers (see Tab. 3.1). All layers were scaled for density. As such, different densities of sand and silicone putty are used to account for the density contrast between the oceanic and continental lithosphere and for a density increase with depth. Density of the ductile layers has been regulated through the addition of quartz sand, whereas different viscosities of the mantle lithospheres are obtained through mixing silicone putty with Barium sulphate (BaSO_4).

The experiments were scaled according to the principles of geometric, dynamic, kinematic and rheological similarity as described in chapter 2 (Hubbert, 1937; Ramberg, 1981; Weijermars & Schmeling, 1986).

The experiments are geometrically scaled by a length scale factor of $L^* = 3.33 \times 10^{-7}$ so that 1 cm in the model corresponds to 30 km in nature. Density is scaled based on the brittle oceanic lithosphere. The stress ratio is then calculated using the density ratio between the quartz sand ($\rho_{qs} = 1500 \text{ kg.m}^{-3}$) and the mantle lithosphere in nature ($\rho_{oc} = 3350 \text{ kg.m}^{-3}$), and the length ratio. Then, $\sigma^* = 1.5 \times 10^{-7}$.

The strength profiles of the brittle layers are calculated for compressional stress regime (Chapter 2, equation 2.16). The slope for the brittle stress is 294.3 Pa/cm representing $\sim 58 \text{ MPa/km}$ in nature which is close to Byerlee's law. To simulate ductile behavior, we use silicone putties with different densities and viscosities (Table 3.1), which approximate flow by creep (Broerse et al., 2019). The strength of the ductile layers is calculated for a simple power law flow law linking stress and strain rate (chapter 2, equation 2.20).

With n values ranging between 1.15 and 1.45, the materials exhibit slightly non-Newtonian behaviour. In nature, the effective viscosity of the ductile mantle lithosphere is a function of the strain rate and composition, and typically ranges between 10^{21} - 10^{22} Pa.s (Billen & Hirth, 2005; Bills et al., 1994; Hirth & Kohlstedt, 2003a; Kaufmann & Amelung, 2000; Vergnolle et al., 2003; Watts et al., 2013), where low viscosities ($\sim 10^{20}$ - 10^{21}) characterise mantle lithospheres with fertile composition (Dixon et al., 2004; Chenin et al., 2019; Vergnolle et al., 2003). The viscosity scale ratio was calculated using the time and the stress ratio. The time ratio is constrained through the shear strain rate, assuming a natural shear strain rate in the order of 10^{-14} s^{-1} and a ductile layer thickness of 30 km (in accordance with the analogue model). Solving equation 2.19 (chapter 2) yields a convergence velocity in nature of 1 cm/yr representing 0.5 cm/h in the experiments. As such an experiment lasting for 20 hours represents $\sim 32 \text{ Myr}$ of shortening in nature, setting the time ratio (t^*) to 7.20×10^{-11} and the viscosity ratio (η^*) to $\simeq 1.0 \times 10^{-17}$, so that the range of viscosities used in the model (10^4 - 10^5 Pa s) represents $\sim 10^{21}$ - 10^{22} Pa s in nature.

The buoyancy number F is defined in chapter 2 and is used to quantify the role of buoyancy forces with respect to the viscous strength of the oceanic lithosphere (Table 3.1) where $F > 1$ favours subduction by the negative buoyancy of the lithosphere and $F < 1$ inhibits subduction (Faccenna et al., 1999).

Table 3.1: Summary of model configurations and material properties per experiment (EXP) together with characteristic properties of natural prototypes. The thickness and density of each layer in the experiments is scaled to nature, based on a compilation of gravity modelling derived data of different passive margins (Namibian: Bauer et al., 2000; Canadian: Li et al., 2016a; Norwegian: Maystrenko & Scheck-Wenderoth, 2009) and cooling models for oceanic lithosphere (Chapman & Pollack, 1977).

Layers	EXP	Material	h		ρ		η		n	F
			M	N	M	N	M	N		
			cm	km	kg/ m³		Pa s			
Ocean brittle lithosphere		Quartz sand	1.4	42	1500	3350				
Continent brittle crust		Feldspar sand	0.8	24	1300	2900				
Continent ductile crust		Rhodorsil Gomme + sand	0.4	12	1300	2900	4.6e4	4.6e21	1.12	
Continent brittle L. mantle		Quartz sand	0.5	15	1500	3350				
Ductile L. mantle	Exp1	Rhodorsil Gomme + sand	1.0	30	1500	3350	1.3e4	1.3e21	1.09	3.03
	Exp2	Rhodorsil Gomme + sand	1.0	30	1500	3350	1.3e4	1.3e21	1.09	1.51
	Exp3	Rhodorsil Gomme + sand	1.0	30	1500	3350	5.8e4	5.8e21	1.13	0.85
	Exp4	Rhodorsil Gomme + sand	1.0	30	1500	3350	5.8e4	5.8e21	1.13	0.42
	Exp5	PDMS + BaSO4+ sand	1.0	30	1500	3350	1.4e5	1.4e22	1.36	0.36
	Exp6	PDMS + BaSO4+ sand	1.0	30	1500	3350	1.4e5	1.4e22	1.36	0.18
Asthenosphere		Polytungstate + glycerol	10	300	1470	3280	~5.0e.2	5.0e19	1.00	

Here h is the thickness, η is the effective viscosity, ρ is the density of the layers and n is the stress exponent. F is the buoyancy number. M and N refers to model and nature, respectively.

3.3.3 Analysis and quantification of vertical movements

In order to show the spatial and temporal evolution of deformation and associated surface topography, we present our results (e.g. Fig. 3.3a) in terms of top-view images and cross sections. Experiments are monitored using digital photography and laser scanning of the top surface. Digital elevation models (DEM's) calculated from scan data to evaluate the extent of deformation between consecutive time-steps of the experiments (e.g. Fig. 3.3c). Vertical movements through time are also monitored along profiles to further delineate where deformation takes place. The amplitude of vertical

movements was calculated by subtracting the topographic values between two successive time steps. Mean deviations between two successive topographic profiles which are less than 0.1 mm have been filtered out. This threshold was set to eliminate small undulations of the model surface and to focus on the first order topographic response to deformation. The length of the deformation zone (e.g., Fig. 3.3c) is then defined as the zone where vertical movements from one step to another one are in excess of the threshold. The total sum of vertical movements (TSVM), derived from DEM data, was used to illustrate and quantify regions in the models where most of the deformation occurs during the entire duration of the experiment (e.g. Fig. 3.3d).

3.3.4 Assumptions and simplifications

The experiments aim at simulating the shortening of a thermally equilibrated passive margin lithosphere. Each experiment is carried out with a constant rate of deformation, which is applied through movement of a wall that simulates a horizontal force like ridge push and far field forces. The resolution of the presented analogue experiments accounts for the first-order rheological layering of the lithosphere. The viscous layers of the models are defined by uniform properties over their thickness and absence of temperature dependent variations of the creep strength with depth are included in the model. In contrast, the ductile behavior in natural systems strongly depends on temperature (Brace & Kohlstedt, 1980; Ranalli, 1995). To facilitate setting up of the experiment, materials used in this study have quasi-Newtonian properties ($n=1$). This deviation from nature ($n=2-3$) may impede strain localization within the mantle and thus the development of a shear zone, representing the plate boundary. Previous studies demonstrate that representing the ductile behavior of the lower crust and lithospheric mantle with uniform viscous materials is an acceptable first-order approximation (Chemenda et al., 1996) that does not affect the general results and conclusion (Auzemery et al., 2020; Faccenna et al., 1999; Martinod & Davy, 1992). In nature density contrasts between a mature oceanic lithosphere and the adiabatic mantle range between 20-80 kg.m⁻³, (Afonso et al., 2007; Schubert et al., 2001). Following Faccenna et al. (1999) we implement a density difference between the oceanic lithosphere and the underlying asthenosphere, which is 30 kg.m⁻³ (representing ~60 kg.m⁻³ in nature) in our experiments to allow for gravitational instability at the ocean-continent transition. This density contrast is assumed to be constant and the buoyancy is regulated through the strength of the ductile mantle lithosphere. The importance of vertical and lateral density variations for the initiation of subduction zones at passive margins is discussed in detail by Faccenna et al., (1999), Mart et al., (2003) and Maystrenko et al., (2013). The evolution of topography is analysed without taking into account erosion, transport and deposition of sediments which can be important for the timing of deformation (Malavieille, 2010; Pinto et al., 2010; Willett et al., 1993). The absence of surface processes and the lack of localizing mechanisms in the ductile mantle lithosphere results in exaggerated vertical motions of the model surface. We nevertheless regard the surface scan data as relevant data, which provides critical information on the locus and relative magnitudes of vertical motions.

The kinematic boundary condition (moving wall) creates slight deformation along the sidewalls within a 1-2 cm wide zone. These deformation structures appear at the beginning of the experiments, do not further evolve with time and do not propagate into the interior of the model. As such, they do not bias the modelling results. Despite the simplifications inherent to the analogue modelling technique, the experiments presented here provide first-order deformation patterns and associated topography that characterize the deformational behavior of passive margins under compression.

3.4 Experimental results

For each rheologic setup, two experiments have been performed with different convergence velocities. The results are presented by sets of experiments with weak, medium strength and strong lithospheric mantle. The modelling results are described in terms of three evolutionary stages including: inversion of the passive margin (stage I), underthrusting (stage II) and subduction (stage III). More specifically, stage I refers to the first deformational response to the applied shortening and neighbouring oceanic lithosphere whereas stage II encompasses the switch from distributed to localized deformation. We refer to “subduction” (stage III) when no major changes in deformation width occur and a slab-like feature evolves.

3.4.1 Weak mantle lithosphere (Set 1)

Experiments 1 and 2 are characterized by a low viscosity lithospheric mantle of 1.7×10^4 Pa s resulting in a buoyancy number that is larger than 1 for both experiments (Table 3.1), so that stress induced by the negatively buoyant oceanic lithosphere overcomes the ductile resistance of the mantle lithosphere.

Experiment 1 is characterized by a slow convergence velocity of 0.5 cm/h representing ~ 1 cm/year in nature. Analysis of topographic profiles (Fig. 3.3b) shows that deformation starts with a very narrow flexure and deformation of the oceanic crust prior to 3% bulk shortening (BS). Deformation results in thrusting within the oceanic lithosphere cutting the model surface parallel to the ocean-continent boundary (Fig. 3.3a, thrust n°1). This fault remains active and controls the development of a shallow trench-type basin, which rapidly deepens up to 5% BS (Fig. 3.3b). During this stage, thrusting leads to uplift in the hanging wall, which is limited to the vicinity of the trench (Fig. 3.3d). In between 5 and 10% BS the thrust polarity reverses and a newly developing thrust (no. 2) displaces thrust n° 1 (Fig. 3.3a). Thrust polarity reversal leads to the vertical accumulation of brittle material below the ocean-continent transition inducing a widening of the deformed area and uplift of the entire passive margin (Figs. 3.3b, c). The subsequent distinct decrease in length of the deformed area is related to a stage where the trench position is stationary and where no additional uplift occurs at the passive margin (Fig. 3.3c, between 10-15%BS). Another change of thrusting polarity occurred at ~ 15 % BS (thrust n° 3) without deforming the continental lithosphere. Trench migration toward the ocean is observed at this stage. Vertical motions are mainly localized in the oceanic domain with maximum subsidence at the trench and maximum uplift of the overriding plate in the immediate vicinity of the trench (Fig. 3.3d). The

length of the zone of deformation is relatively narrow and scales to a maximum length of ~350 km in nature (Fig. 3.3c).

Experiment 2, with a convergence velocity of 1 cm/h, evolves similarly to experiment 1. However, the main thrust fault is located much closer to the continental margin. As in experiment 1, the continental margin is not deformed internally, but uplifted in response to thrusting of oceanic crust below the margin (Fig. 3.3f, topographic profile at 5% BS). Prior to underthrusting, a narrow fold with a wavelength of 3 cm marks the flexure of the OCT (Figs. 3.3f, scan and topographic profile at 5% BS). Faster convergence correlates with a more rapid increase in deformation length, but the maximum length of the deformation zone is similar in both cases (Fig. 3.3g). The trench slowly migrates ocean-ward and gradually deepens during the long-lasting activity of thrust n° 1. The flip in thrust polarity (thrust n° 2) commences at 12% of BS and results in a prismatic pop-down bounded by opposite verging thrust faults.

In both experiments, deformation only affects the oceanic lithosphere by the activation of intra-ocean thrust faults. Thrusting is associated with thickening of the crust leading to uplift of the overriding plate and an increase in deformation length. The length of deformation is around 6 to 10 cm in our experiments, representing 200 to 300 km in nature. The maximum width of the experiments (stage I) is reached once the thrust polarity has changed. Thereafter the width of deformation decreased rapidly (experiment 1) and slowly (experiment 2), characterizing stage II in the evolutionary scheme. We note that a stable configuration (stage III) is only achieved in experiment 1. For a buoyancy number larger than one, the viscous resistance of the lithospheric mantle does not prevent the sinking of the crustal blocks. Intra-oceanic deformation is accommodated by major thrusts, which change polarity through time and create large subsidence at trench-type basins. In contrast, the low strength of the lithospheric mantle and its Newtonian behavior results in ductile flow but not in the localization of strain and thus the development of a shear zone at that level.

Figure 3.3: Results of experiments 1 and 2 testing a weak lithospheric mantle. (a, e) (left) Final, 3D lithospheric scale structure and (right) topographic evolution of experiments. The cross-sections are located at two-third of the width from the top of the model. Abbreviations are the same as in Fig. 3.2. (b, f) Topographic profiles are retrieved through DEMs at 0%, 3%, 5%, 10%, 15%, and 20% bulk shortening (BS). Vertical exaggeration is 10 times. (c, g) Length of the deformation zone at different times of the experiment also expressed in bulk shortening. I, II and III are the three evolutionary stages: (I) inversion of the passive margin (II) underthrusting and (III) subduction (d, h) Total sum of vertical movements across the section, calculated during the entire time span of the experiment. The amplitude of vertical movement was calculated by subtracting the topographic values between two profiles between two-time steps given in the graph in part (c), expressed in bulk shortening. The amplitude of vertical movements was used to define the zone of deformation in part (c, g) and the total sum of vertical movements in part (d, h). →

3.4.2 Medium strength mantle lithosphere (Set 2)

The second set of two experiments is characterized by an intermediate viscosity of the lithospheric mantle of 5.8×10^4 Pa s and corresponds to passive margins with moderate age and a medium strength lithospheric mantle. The buoyancy numbers for these experiments are 0.85 and 0.43 for the slow and fast experiment, respectively.

Experiment 3 with a slow convergence velocity of 0.5 cm/h shows deformation of the passive margin and almost no deformation in the oceanic domain (Figs. 3.4a-d). Shortening results in deformation of the OCT commencing after 5% BS (Fig. 3.4a, fault 1 and 2). Coevally, the former passive margin is flexed downward at a wavelength of 11 cm (Fig. 3.4a, scan at 5% BS). This flexural basin is flanked on both sides by bulges. We note that deformation spreads over a wide area covering the entire margin at this early stage (stage I) in model evolution (Fig. 3.4c). Thrusting localizes where the ductile lower crust of the margin wedges-out and develops with a backward sequence in the direction of the continent. A trench-type basin forms between the oceanic and the continental lithospheres, and progressively deepens during shortening, as illustrated by the topographic profiles at 5% BS. After 10% BS, the down-flexed passive margin is cut by the evolving fault 2 (Fig. 3.4a) and underthrusts the continent. This step marks the starting of the underthrusting stage (stage II) where deformation is mainly localized at the passive margin. It is characterized by a decrease in length of the zone of deformation (Fig. 3.4c, between 10% and 17% BS). At this stage, the displacement of the deformed margin toward the continent and the stationary character of the continental edge suggests that the continental crust at the margin gets squeezed by the oceanic lithosphere. The cross-section shows that the whole passive margin is accreted at the trench and the down-going plate drags part of it. In particular the accreted brittle crust is decoupled from the underlying mantle at the level of the ductile lower crust. The cross section at the end of the experiment displays the geometry of a vertical slab-like feature, formed during stage III.

Experiment 4 with a convergence velocity of 1 cm/h resulted in a similar sequence of events and deformation pattern, although the topography is lower than in experiment 3. Deformation at the ocean-continent transition creates more scattered vertical movements and induces rapid uplift at the continental side. Development of a thrust system results in a rapid increase of the length of the deformed area (Fig. 3.4g). Coevally the continental lithospheric mantle is shortened and folded with a wavelength of 13 cm, leading to an uplift of 3 mm. A shallow flexural basin formed behind the thrust-system at the margin and rapidly closed, after 10% BS. During this stage the length of the zone of deformation increases (Fig. 3.4g, 0-10% BS). However, when the trench is formed and underthrusting starts (stage II), deformation localized at the trench, and the length of deformation decreases (Fig. 3.4g, 10-17% BS). The almost stationary character

Figure 3.4: Results of Experiments 3 and 4 testing a moderate viscosity of the lithospheric mantle. for explanation of the figure, see Figure 3.3. →

of the trench and of the continental edge suggest that underthrusting is mainly controlled by the convergence of the oceanic plate. Underthrusting is again facilitated by decoupling above the Moho and also drags down the continental mantle lithosphere, which together with the oceanic mantle lithosphere envelopes a tight syncline cored by ductile lower crust. We interpret the constant length of deformation during the last stage of the experiment (Fig. 3.4g, 17% BS) as marking the final phases of deformation (stage III).

In both experiments, thrusts strike parallel to the evolving trench. The mechanically weak lower continental crust controls the locus of thrust formation and acts as a decollement in which the upper crustal thrusts root. The ductile lower crust allows for crust-mantle decoupling, ultimately leading to partial delamination of the mantle lithosphere. The two experiments differ at this stage such that exp. 4 shows a clear delamination expressed by the strong deformation of the lower crust whereas exp. 3 shows more or less lithospheric dripping. In both experiments, deformation mainly affects the passive margin and is associated with the formation of a flexural basin between oceanic and continental lithosphere over a length of ca. 500-600 km when scaled to nature. Localization of deformation at the passive margin and at the continental edge, induces strong vertical movements (Figs.3.4g, h). Uplift increases progressively during underthrusting. It is caused by the accretion of material at the trench and short wavelength ($\sim\lambda = 300$ km, scaled value) buckling of the continental crust. The experiments also show that the topographic response depends on the convergence rate, creating longer but lower amplitude folding for high convergence rate. The switch from widespread deformation during the early evolution of the experiments (stage I) to localized underthrusting (stage II) is characterized by a reduction in length of the deformation zone amounting to ca. 300-400 km (Figs.3.4c, g). In both cases, stage III is characterized by a constant length of deformation over time.

3.4.3 Strong mantle lithosphere (Set 3)

A lithospheric mantle with a viscosity of 1.4×10^5 Pa s characterizes the third set of two experiments representing passive margins with a strong lithospheric mantle. The buoyancy number for these experiments is 0.36 and 0.18 for the slow and fast experiment respectively.

Experiment 5 corresponds to a slow convergence velocity of 0.5 cm/h. Up to 5% BS, deformation is marked by the development of a frontal thrust (n° 1, Fig. 3.5a), located at the boundary between oceanic lithosphere and passive margin. This thrust induces short wavelength crustal folding ($\lambda_1 \sim 7$ cm wavelength). Coevally, downward flexure of the margin defines a basin at the backlimb of the thrust. At 5% BS, the margin is subject to long wavelength folding λ_2 (18 cm wavelength) and uplift up to 4 mm (see scan and topographic profiles, Fig. 3.5b). The length of the zone of deformation gradually increased (Fig. 3.5c). During thrust propagation, a strong uplift affects the basin edge in response to flexure. At 20% BS, the section reveals that the main thrusts (n° 1 & 2) cut through the brittle mantle layer. However, these faults do not propagate through the ductile lithospheric mantle. Strong ductile resistance of the mantle

Figure 3.5: Results of Experiment 5 and 6 testing a strong lithospheric mantle. For explanation of the figure, see Figure 3.3.

lithosphere opposes the slab descent and folding affects the entire lithosphere (cross-section in Fig. 3.5a).

Experiment 6 corresponds to a fast convergence velocity of 1 cm/h. This experiment shows a geometrical pattern comparable to experiment 5 but with the different stages developing simultaneously. Underthrusting of the oceanic plate does not occur and shortening is accommodated by short wavelength crustal folding ($\lambda_1 \sim 7$ cm) and large wavelength lithosphere buckling of both the continent and the ocean ($\lambda_2 \sim 17$ cm, profile Fig. 3.5f). Similar to experiment 5, the passive margin subsides, and the continental edge is uplifted (Fig. 3.5f). In addition, folding is associated with low-offset, dominant oceanward-verging thrusts and pop-up structures. High topography reveals that deformation amplitude is large in the oceanic domain and decreases at the OCT and in the continental lithosphere close to the trench. Flexure is more pronounced than in experiment 5 and affects the trench, the frontal thrust system and the fore-arc basin.

In both experiments the failure of the crust occurs at the OCT. At high shortening velocity, deformation is distributed over a wider area expressed by regions of uplift and subsidence (Figs. 3.5a, e). The highest topography is observed at the continental edge. Although the cross-section of experiment 5 suggests that underthrusting (stage II) occurs, none of the experiments is characterized by a reduction in deformation width. Instead, the area affected by deformation reaches a steady state (experiment 5) or continuously increases (experiment 6), indicating that shortening is dominantly accommodated by long-wavelength folding ($\sim \lambda = 500$ km, scaled value) controlled by the high strength of the ductile mantle lithosphere. The zone of deformation represents 700-800 km in nature. The combination of near surface brittle deformation and lithosphere folding gives rise to a very pronounced and characteristic topography development (Figs. 3.5d, h). As a consequence, the continental margin edge experiences a strong uplift whereas the foreland is marked by a pronounced subsidence. As in previous experiments, the transition between oceanic and passive margin lithosphere shows local uplift caused by the deformation of the thinned crust and its accretion.

3.5 Discussion

3.5.1 Controlling parameters for passive margin instability

Our experimental results consistently highlight the dependence of strain localization and locus of deformation on the viscosity of the ductile lithospheric mantle and thus its strength, despite the lateral homogenous nature of the ductile mantle lithosphere. For the effect of lateral strength variations in the lithospheric mantle the reader is referred to Calignano et al. (2015) or Santimano and Pysklywec (2020). The rheology of the mantle lithosphere has a strong effect on the distribution of deformation, such that deformation localizes within the oceanic lithosphere with a low strength in the mantle layer, whereas it localizes at the OCT in all other cases (lithospheres with intermediate and high ductile mantle viscosity, Figs. 3.4 and 3.5). Additionally, increasing the viscosity of the mantle lithosphere also leads to a change in deformational behavior within that layer from flow dominated (experiments 1-4) to folding dominated

(experiments 4-5). As in all experiments the oceanic and continental crust remained unchanged, we argue that the above-mentioned differences are a direct consequence of differences in shear resistance in the ductile mantle lithosphere, a parameter that is key for developing shear instabilities which potentially evolve into subduction zones as shown by previous analogue modelling studies (e.g. Faccenna et al., 1999; Goren et al., 2008; Luth et al., 2010; Sokoutis & Willingshofer, 2011; Willingshofer et al., 2013).

Although incipient deformation is easier achieved within the passive margin crust because only low stress levels are required to mobilize the ductile layer, the stability of the lithospheric mantle is critical for the further evolution of the system. Previous modelling studies have shown that the negative buoyancy of the oceanic plate as well as the density contrasts between continental and oceanic lithospheres are important for passive margin inversion and in particular for subsequent subduction initiation (Chemenda et al., 1996; Faccenna et al., 1999; Goren et al., 2008; Leroy et al., 2004; Mart et al., 2005). Following Faccenna et al. (1999), we use the buoyancy number (Fig. 3.6a) to characterize system behavior. In addition, our experiments show a relationship of F to the place where deformation initiates. If $F > 1$, the stability of the oceanic lithosphere is compromised, and underthrusting occurs in the oceanic domain (experiments 1-2). These modelling results apply for cases of highly negative buoyancy of the oceanic lithosphere which seems to be at variance with the low viscosity in the lithospheric mantle (less than 2×10^{21} Pa s in this study). As viscous strength in nature increases rapidly with age (England & Wortel, 1980), intra-oceanic lithosphere instability requires large negative buoyancy and large melt flux through the lithosphere to facilitate its weakening. A natural example would be the inversion of a partially melted oceanic lithosphere by subduction-induced mantle upwelling (Strak & Schellart, 2018) or a mantle plume setting (Burov & Cloetingh, 2010; Gerya et al., 2015; Monnereau et al., 1993; Rolland et al., 2009). The latter might well have been the case for Caribbean subduction initiation (Whattam & Stern, 2015).

In the case of a stable lithosphere ($F < 1$), deformation occurs at passive margins. In general, a low buoyancy number should prohibit underthrusting (Faccenna et al., 1999). However, adding a contractional component of deformation seems to allow for deviating from this concept and facilitates underthrusting even with buoyancy numbers as low as 0.85 or 0.42 (experiments 3 and 4). To find explanations for this apparent discrepancy the pre-underthrusting stage of deformation needs to be considered: Detailed analysis of the deformational sequence revealed that early-stage deformations are characterized by thrusting, which commences at the tip of ductile continental crust, where horizontal shortening is accommodated by layer-parallel shear. This is illustrated by the presence of ductile material along the ramp of the oldest thrust (Fig. 3.4). As such we suggest that the rheological stratification of the passive margin and in particular that of the ductile crust provides a first order control on the locus of deformation during incipient margin inversion (stage I). Similar to Nikolaeva et al. (2011), our experiments suggest that the early-stage deformation in the crust pre-conditions where deformation will be transferred to the mantle lithosphere. More specifically, shear localization and associated deformation at the base of the continental

Figure 3.6: (a) Summary of modelling results for tested parameters of this study. (b) Mode of deformation at passive margins including I passive margin inversion, II underthrusting, and III subduction. Red segments represent the length of the deformed region. The figure shows a schematic illustration of the evolution of deformation at a passive margin with favourable conditions for subsequent subduction initiation. (c) Relationship between wavelengths of lithospheric folding and thermo-tectonic age of lithospheres in nature (modified after Cloetingh & Burov, 2011, references in Table 3.2), including data for Algeria (A), Northern Australia (NA), Southern Australia (SA), Black-Sea (BS), Brazil (B), Canada (Ca), Cantabria (c), Europe (E), Iberia (Ib), Southern India(SI), Norway (N), South-Korea (SK) passive margins. The grey zones correspond to theoretical folding wavelengths, derived from analytical models, for continental lithosphere (Cloetingh et al., 1999) as well as for oceanic lithosphere (McAdoo & Sandwell, 1985). The red dashed line is the folding wavelength as deduced from the analogue modelling.

Table 3.2: Wavelengths and thermo-tectonic ages of folded lithosphere displayed in Fig. 3.6c.

Id	Area	Ages of lithosphere (Myr)	Folding wavelengths λ (km)	References
Al	Algerian margin	16-25	100-150	Hamai et al., 2015; Roure et al., 2012
NA	Northern Australian margin	540-560	200-300	Keep & Harrowfield, 2008
SA	Southern Australia intra-continental	390-450	500-700	Holford et al., 2011; Célérier et al., 2005
SA	South Australian margin	390-450	100-300	Quigley et al., 2010;
BS	Black-Sea margin	75-125	125-150	Cloetingh et al., 2008 Starostenko et al., 2013
B	Brazil Intra-continental	590-630	750-1,000	de Lima, 2000
B	Brazilian margin	590-630	200-300	This study
Br	Brittany margin	210-290	225-275	Bonnet et al., 2000
Ca	Canadian (Arctic) margin	150-250	150-200	Stephenson & Cloetingh, 1991
C	Cantabrian margin	330-370	150-250	This study; Viejo & Gallastegui, 2005a
E	Western Europe	180-230	270-300	Ziegler et al., 1995; Holford et al., 2009
SI	South Indian margin	80	200-250	Beekman et al., 1996
Ib	Iberia intra-continental	330-370	400-600	De Vicente & Vegas, 2009
SK	South-Korea	20-25	70-100	Kim et al., 2018
N	Norway	1,000	100-300	Løseth & Henriksen, 2005

crust regulates the degree of coupling and permits deflection of the oceanic plate under the margin (Fig. 3.6b), thereby creating a mechanical instability allowing for the change of distributed deformation (stage I) to localized deformation (stage II) and ultimately for partial delamination of the mantle lithosphere. When deformation localizes at the base of the crust, the strength of the mantle lithosphere controls the extend of underthrusting and thus the stability of the margin. We note that the convergence velocity plays also an important role in the delamination process. Therefore, favourable conditions for inversion of passive margins that might evolve into a subduction zone lie in a limited spectrum of lithospheric mantle viscosities of around 5.0×10^{21} Pa s (when scaled to nature), which is consistent with intermediate to mature oceanic lithospheres (Auzemery et al., 2020; Oxburgh & Turcotte, 1976; Turcotte & Schubert, 2014), or again a thermally weakened lithosphere at the margin (Jammes & Lavier, 2016). For viscosities higher than that shortening will lead to large-scale folding of the margin and neighbouring lithospheres (Fig. 6a). In the extreme case, lithosphere folding and associated thrusting can affect oceanic plates as observed in the Indian ocean (Beekman et al., 1996; Stein et al., 1989a). Such strong lithosphere produces deformation at crustal levels equivalent to experiment 5 and 6. We acknowledge that thermal inputs could change these conditions and could impact on margin evolution.

3.5.2 Comparison with numerical modelling

The analogue models display striking similarities with numerical models of passive margin inversion and in particular of subduction initiation (e.g. Auzemery et al., 2020; Nikolaeva et al., 2011; Kiss et al., 2020). Both methods highlight the role of the strength of the mantle lithosphere and of crust/mantle decoupling at passive margins. However, the quasi-Newtonian viscosities in the analogue experiments impede strain localization within the mantle and thus the development of a shear zone, representing the plate boundary, which would then accommodate the downward motion of the subducting plate (Fig. 3.6b, III). Instead, downward motion involved both, the subducting as well as the overriding plate (Fig. 3.6c, III) leading to “two-sided” subduction as described by Gerya et al. (2008). In our models, this behavior can be attributed to the lack of externally imposed or intrinsic weakening mechanisms such as shear heating (Auzemery et al., 2020; Kiss et al., 2020; Thielmann & Kaus, 2012) or fluid percolation (Alvarez-Marron et al., 1997; Regenauer-Lieb et al., 2001). Based on our experimental results, we argue that in the absence of inherited weak zones or weakening mechanism, incipient subduction should be two-sided until one of the above-named processes or a combination thereof leads to strain localization. In this case, subduction can also involve part of the continental lithosphere such as in experiment 3.

Additionally, thermo-compositional mechanisms may contribute to the stability of the mantle lithosphere (Boonma et al., 2019) to explain the variation in strength in the mantle lithosphere. Convergence rate, which regulates thermo-compositional mechanisms, may be of particular importance in nature to create instabilities at depth > 140 km (Boonma et al., 2019). Moreover density-enhancing processes, like

eclogitization, should occur in the underthrust lithosphere, such that slab pull may start after 100-150 km of forced underthrusting (Gurnis et al., 2004; Maffione et al., 2015).

3.5.3 Length scales of deformation during passive margin inversion

Similar to our experiments, inverted passive margins are marked by local deflection at the OCT (e.g. Hamai et al., 2015), increased seismicity/deformation at the margin (Stein et al., 1989b), strong uplift of the continental edge (Leroy et al., 2004) and folding of the continental lithosphere (Cloetingh et al., 1999; Ziegler et al., 1995). Moreover, our experimental results consistently show the dependence of length scales of deformation on the strength of the ductile lithospheric mantle (Fig. 3.6a). Low viscosity of the ductile mantle lithosphere results in localized intra-oceanic deformation whereas stronger mantle lithospheres lead to wide deformation of the passive margin. In particular, prior to underthrusting, the passive margins are marked by folding where the wavelength is controlled by the strength of the mantle lithosphere. Converting the periodicity of the folds in models to the natural scale, the associated length-scales of folding are in the order of 50 to 90 km for experiments with a weak mantle lithosphere, 350-400km for experiments with medium lithospheric mantle viscosities and 600-700 km for experiments with a strong mantle lithosphere (Fig. 3.6c). Therefore, our experimental results are consistent with the findings of Cloetingh et al. (1999), suggesting that lithospheric folding is a primary response of the lithosphere to shortening. Moreover, Cloetingh and Burov (2011), emphasize the dependence of fold wavelength on plate age and thus strength of the mantle lithosphere. Along these lines, we compare our experiments with estimates of folding wavelengths from oceanic and continental lithospheres as well as margins with different thermo-mechanical age (Fig. 3.6c and Table 3.2). In our experiments, shortening of lithosphere with a weak mantle layer results in very localized deformation in the weaker oceanic plate and a very gentle deflection at the margin (exp. 2). In nature, contraction of a young lithosphere, such as present at the Algerian or the South-Korean margin, gives rise to localized deformation at the OCT, that includes downward bending of the oceanic plate relative to the continental crust (Hamai et al., 2015) and development of short wavelength folds (70-100km) in the oceanic domain (Kim et al., 2018). Moreover, the Algerian margin also reveals high seismicity at the OCT (Fig. 3.7a). However, there is little evidence for ongoing deformation further onshore. By comparison, our experimental results suggest that a weak oceanic lithosphere accommodates most of the shortening and deformation of the passive margin is expressed by its deflection. As such, we argue that deformation of a very young passive margin (Kim et al., 2018; Hamai et al., 2018) does not necessarily provide evidence for underthrusting at the margin (Fig. 3.7d).

Length-scales of deformation deduced from experiments 3-4 are consistent with wavelength values of mantle lithosphere folding (Fig. 3.6c). In nature, the wavelength of folding of inverted passive margins plots somewhat lower compared to theoretical predictions of folding wavelengths from Cloetingh et al. (1999), suggesting that the thermo-tectonic age of the margin lithosphere is lower because of thermal inputs during

rifting. This is, for example, the case for the passive margins of the Black Sea (Cloetingh et al., 2008; Starostenko et al., 2013), Arctic Canada (Stephenson & Cloetingh, 1991), Brittany (Bonnet et al., 2000), Norway (Løseth & Henriksen, 2005; Pascal & Cloetingh, 2009), or northern and southern Australia (Holford et al., 2011; Keep & Harrowfield, 2008; Quigley et al., 2010). These margins show local deflections at the OCT, uplift at the margin and the development of a continental flexural basins with a wavelength of 100-300 km (Fig. 3.6c). Among them, the Cantabrian margin is probably the best example of margin inversion prior to underthrusting (De Vicente & Vegas, 2009). The Cantabrian domain is the western continuation of the Pyrenees and reflect inversion of the North-Iberian margin during the Eocene-Oligocene rather than a continental collision. The gravity data of the margin show a gentle gradient in gravity anomaly towards the south (Fig. 3.7f, Pedreira et al., 2007) typical of plate flexure (Karner & Watts, 1983). Additionally, a thrust system is convincingly imaged at the OCT associated with an Eocene-Oligocene wedge (Fig. 3.7g). Further south, Eocene syn-compressional sediments are identified on the hinterland marginal basin while the margin shoulder is uplifting. All these observations point toward the development of a thrust system at the OCT during Eocene time, leading to a gentle flexure at the margin ($\lambda \sim 250$ km). In the Cantabrian domain, seismic data suggest that shortening led to underthrusting. Yet, underthrusting occurred toward the ocean, giving rise to an early stage continental subduction or delamination (Teixell et al., 2018).

In contrast to models 3 and 4, lithospheres with a strong mantle lithosphere do not evolve into subduction zones and are characterized by long wavelength lithosphere-scale folding (exp. 5 and 6, Fig. 3.5). These experiments are in agreement with offshore and onshore observations of old and strong passive margin lithospheres such as present at the south Brazilian margin and the north Atlantic margin. The late Cenozoic exhumation episodes identified in the British Isles and adjacent regions were most likely caused by the transmission of compressional stresses into continental NW Europe from the Mid-Atlantic Ridge and the Alpine orogen (Holford et al., 2009; Ziegler et al., 1995).

Figure 3.7: (a, e, i) Relief map (NASA Worldview) with epicentres of crustal earthquakes (red dots). Campos Basin (CB), Chaco basin (C), Santos Basin (SB), Parana basin (Pa), Ocean Continent Boundary (OCB). (b, f, j) Measured free air gravity anomalies (red line) and topography (black line) for the Algerian margin (Leprêtre et al., 2013), Cantabrian (Viejo & Gallastegui, 2005a) and Brazilian (Van der Meijde et al., 2013) margins, respectively. (c, g, k) Crustal models of the passive margins indicating length-scales of deformation: (λ/λ_1) short and ($\lambda/2$) long wavelength folding as a function of mantle lithosphere strength. The compilations are based on: the Algerian margin: (Benhenni et al., 2019; Hamai et al., 2015; Leprêtre et al., 2013); the Cantabrian margin: (Teixell et al., 2018; Viejo & Gallastegui, 2005a), the Brazilian margin: (Chulick et al., 2013; Milani & Zalán, 1998; Stanton et al., 2019). (d, h, l) Cartoons showing wavelength of deformation as function of mantle lithosphere strength. The red layer shows the decoupling of deformation at the passive margin. →

Deformation is identified across long distances exceeding 1000 km and includes different styles such as reactivation and inversion of pre-existing faults (Lundin & Doré, 2002), uplift at the margin (Holford et al., 2009) and long-wavelength buckling of the European lithosphere (Bourgeois et al., 2007; Holford et al., 2009, Ziegler et al., 1995). Similarly, the South American Plate is experiencing horizontal shortening controlled by Andean tectonics and compressional interaction of the continental lithosphere with the strong oceanic lithosphere is found eastward (de Lima, 2000). Along the Brazilian margin, the offshore earthquakes distributed along the continental shelf and slope are predominantly due to reverse faulting (Assumpção & Sacek, 2013). There, the edge of the continental margin records Neogene-Pleistocene fault inversion (Cogné et al., 2012; Riccomini et al., 2004) and strong uplift in the Campos do Jordão plateau (Modenesi-Gauttieri et al., 2011) and the Taubaté Basin (Cogné et al., 2012). In the same area, several studies show that the direction of the maximum horizontal stress trends E-W (Assumpcao, 1992; Barros et al., 2009; Chimpliganond et al., 2010; Coblenz & Richardson, 1996; Ferreira et al., 1998; Neto et al., 2013). Further onshore, a similar compressional E-W trend is observed in the Parana, Pantana and Chaco basins (Milani & Zalán, 1998; Rocha et al., 2016; Ussami et al., 1999).

The presented analysis of the south American plate accounts for ongoing deformation including short wavelength lithosphere folding at the margin λ/λ_1 that is ca. 300 km long, typical of a mantle lithosphere folding (Fig. 3.6c), associated with long wavelength antiforms λ_2 of the order of 750-1000 km (de Lima, 2000), representing whole lithosphere folding. Assuming that the majority of the data are indeed related to inversion, it is unlikely that these margins are evolving into subduction zones.

In summary, comparing the above quoted observations with our modelling results suggest that shortening of an oceanic lithosphere and passive margin with intermediate strength provides the best conditions for localizing deformation at the passive margin and the subsequent nucleation of a subduction zone.

3.6 Conclusions

Our physical analogue models focus on the mechanisms and deformational response during the process of passive margin inversion. Our study supports views that both the décollement at the base of the ductile crust and the strength of the viscous mantle lithosphere are key elements for creating an instability at the passive margin. Experimental results show that intermediate strength of ductile mantle lithospheres (i. e. with a mean viscosity η of 5.8×10^{21} Pa s) favours deformation of the passive margin whereas a low strength mantle lithosphere (η of 1.7×10^{21} Pa s) leads to unstable intra-oceanic subduction and a strong mantle lithosphere (η of 1.4×10^{22} Pa s or larger) to lithosphere folding.

Passive margin inversion for favourable mechanical conditions includes three stages: i) widespread deformation of continental crust at the passive margin by folding ii) deformation that is confined to the passive margin where shortening is accommodated by a continent-ward propagating thrust system that evolves from and roots in a

decollement at the base of the ductile passive margin crust facilitating underthrusting of the oceanic plate, and iii) subduction of the oceanic lithosphere.

On the basis of our experimental results, we constrain early stage length scales of deformation at passive margin to predict the likeliness of subduction initiation. Our models predict that underthrusting would occur in case of folding at passive margin over a distance of $\lambda = 100\text{-}300$ km, whereas short ($\lambda = 50\text{-}100$ km) and long wavelength foldings ($\lambda = 500\text{-}1000$ km) are characteristic of intra-oceanic underthrusting and continental lithosphere folding, respectively. Our results are consistent with length-scales of deformation of natural inverted passive margins and allow for narrowing down favourable mechanical conditions for subduction initiation.

Appendix A

This supporting information provides the detail of the numerical modelling used for the calculation of strength and integrated strength in figure 3.1.

The two-dimensional (2D) finite-difference, marker-in-cell code (MDoodz; Duretz et al., 2016) solves the equations of momentum and mass conservation, as well as the heat equation. For details regarding the mathematical model and algorithms, see chapter 1. Deformation is governed by frictional, dislocation, diffusion and Peierls creep equations.

The strength profiles across the Angolan margin was calculated with the geometry from figure 3.1a. The model consists of a two layered continental crust and an oceanic crust overlying the mantle lithosphere (fig 1a), with sediments on top. Thermal boundary conditions include fixed temperatures at the surface and at the base of the lithosphere: $T_0 = 20^\circ\text{C}$ and $T_m = 1330^\circ\text{C}$ (e.g., Turcotte & Schubert, 2014). The initial temperature field is obtained by solving the steady state heat equation using reference thermal parameters (Tab. S1) excepted below the lithospheric mantle where the conductivity was set artificially high in order to produce a quasi-adiabatic asthenosphere. The heat equation includes the effect of radiogenic heat production (Q_r) and a constant asthenosphere temperature of 1330°C bellow the thermal boundary. Because this boundary is not very well defined in nature, the thermal base of the continental and oceanic lithosphere is kept constant and a linear interpolation is applied between the base of oceanic and continental plate (Fig. 3.1a). The thermal thickness of the lithosphere is defined using the thermal data from Maystrenko et al. (2013) and Artemieva (2009), giving 100km for the oceanic lithosphere and 140km for the continental lithosphere. The differential Stress field and resulting stress envelopes (fig. 3.1) were calculated at a strain rate of 10^{-14} s⁻¹, using flow law's parameters for dry grano-diorite for the continental upper crust (Ranalli, 1995) dry dolerite for the mafic crust (Mackwell et al., 1998), Calcite for the sedimentary layer (Kronenberg et al., 1990), dry diabase for the oceanic crust (Mackwell et al., 1998) and dry olivine for the lithospheric and asthenospheric mantle (Hirth & Kohlstedt, 2003a).

Table 3.S1: Rheological and thermal values used for the numerical model. References for rheology are R_95: Ranalli (1995) ; H&K_04: Hirth and Kohlstedt (2003b) ; K_90: Kronenberg et al., (1990); M_83: Mackwell et al., 1998. Peierls mechanisms are described in detailed in the appendices.

Layers	ρ (kg. m ⁻³)	k (W.m ⁻¹ . K ⁻¹)	Qr (W.m ⁻³)	Φ (°)	A (Pa ⁻ⁿ .s ⁻¹)	n	Q (J.mol ⁻¹)	Ref.
Continent crust (dry granodiorite)	2800	2.7	1.0e-6	30	3.16e-26	2.3	154.0e3	R_95
Sediments (calcite)	2600	2.5	1.0e-6	20	1.59e-25	4.7	297.0e3	K_90
Continent Lower crust (Diabase)	2900	2.7	1.0e-7	30	5.04e-28	4.7	485.0e3	M_83
Hydrated mantle (wet olivine)	2900	2.9	1.0e-10	30	5.68e-27	3.5	480.0e3	H&K_03
Ocean crust (Diabase)	2900	3	1.0e-10	30	5.04e-28	4.7	485.0e3	M_83
Mantle lithosphere (dry olivine)								
Dislocation creep	3360	3.2	1.0e-10	30	1.10e-16	3.5	530.0e3	H&K_03
Diffusion creep					1.50e-15	1.0	375.0e3	H&K_03
Asthenosphere	3300	3.2	1.0e-10	30	1.10e-16	3.5	530.0e3	H&K_03

Here p is the density, k is the thermal conductivity, Qr is the radiogenic heat production, ϕ is the friction angle, A is a pre-factor, n is the stress exponent. Q is the activation energy. The shear modulus G is fixed to $3e10$ Pa.

Chapter 4

This chapter is based on the following manuscript: Auzemery A., Willingshofer E., Yamato P., Duretz T., Sokoutis D. (2020). Strain localization mechanisms for subduction initiation at passive margins. Published in *Global and Planetary Change*.

Chapter 4

Strain localization mechanisms for subduction initiation at passive margins

Highlights

- Analogue and numerical models are used to infer strain localization mechanisms for subduction initiation at passive margins.
- At passive margins subduction initiation is controlled by the strength contrast between the subducting oceanic plate and the mantle lithosphere.
- Evolution towards sustainable subduction is controlled by both Peierls creep mantle rheology and shear heating.

4.1 Introduction

Although subduction initiation at passive continental margins is lacking direct observations, several subduction zones are believed to have been formed at these settings (Hall, 2019; Karig, 1982; Manzotti et al., 2014; Marroni et al., 2017). Inverted margins possibly reflect early stages in the process of subduction initiation and several examples (Pedoja et al., 2011; Yamato et al., 2013) promote the idea that the difference in strength between the oceanic and continental lithosphere may be critical for initiating subduction. Hence, rheological contrasts between the oceanic and continental lithospheres have been considered as key parameters for subduction initiation (SI) (Baes et al., 2018; Cloetingh et al., 1989; Erickson, 1993; Faccenna et al., 1999; Goren et al., 2008; Mart et al., 2005; Nikolaeva et al., 2010; Hamai et al., 2018; Yamato et al., 2013). Although these studies used different approaches, they consistently emphasize that the aging of the oceanic plate is key for providing favorable mechanical conditions for SI at passive margins. In the case of a young passive margin, the strength contrast required at the ocean-continent transition (OCT) is not enough and intra-oceanic subduction and obduction is then likely to happen (Düretz et al., 2016a; Gülcher et al., 2019b; Maffione et al., 2015; Zhong & Li, 2019). For older oceanic lithosphere, SI at passive margins requires additional deformation mechanisms such as deformation of the ductile crust at the margin and deflection of the oceanic lithosphere (Nikolaeva et al., 2010).

Aging of the oceanic plate does not only regulate the strength of the lithospheric mantle at passive margins but also leads to increased mechanical coupling between the oceanic and continental plate (Cloetingh et al., 1989; England & Wortel, 1980; McKenzie, 1977; Mueller & Phillips, 1991). Consequently, most attempts of analogue and numerical modelling of subduction initiation required a weak lithospheric mantle or a pre-existing weak-zone in the lithospheric mantle (Faccenna et al., 1999; Mart et al., 2005; Nikolaeva et al., 2010). More recent studies mention necessary convergent forces to initiate induced subduction initiation in the absence of weak zone in the mantle lithosphere, without detailing rheological mechanisms (Zhong & Li, 2019). Additionally, it has been shown that mechanisms that reduce the strength of the

lithospheric mantle such as Peierls creep in olivine (Cramer & Kaus, 2010; Drury, 2005; Korenaga & Korenaga, 2016), grain size reduction (Ricard & Bercovici, 2009) or shear induced thermal softening (Duretz et al., 2016a; Jaquet & Schmalholz, 2018; Kiss et al., 2020; Thielmann & Kaus, 2012) are important for the development of large-scale shear zones in the mantle lithosphere, which are precursor structures of future subduction plate interfaces. Peierls creep mechanism in olivine strongly reduces the brittle strength in the uppermost strong mantle lithosphere and may be responsible for strain softening (Kameyama et al., 1999). Thermal softening enables shear localization in rocks deforming under temperature-sensitive creep (Duretz et al., 2016a; Jaquet & Schmalholz, 2018; Kiss et al., 2020; Thielmann & Kaus, 2012).

In contrast to previous studies, we aim to investigate rheologically controlled mechanisms that can be responsible for strain localization in the case of induced subduction initiation at a passive margin, in the absence of pre-defined weak zones. Our combined physical analogue and numerical modelling approach inquires the role of the mechanical stratification of ocean-continent transition and associated lateral strength contrasts at various depth levels within the lithosphere for incipient stages of SI. In particular, we explore the role of Peierls creep deformation and shear heating mechanism for the development of shear zones that might become subduction interfaces.

4.2 General concept and modelling setup

In this work, analogue modelling provides simple high-resolution 3D outputs of crust and lithosphere geometries, which facilitate interpretation of complex geological scenarios, whereas numerical modelling allows exploring temperature driven feedback mechanisms. To ensure a meaningful complementary use of both methods we scaled the experiments to obtain, within limits, similar geometrical and mechanical conditions at the onset of shortening (Fig. 4.1).

Through both approaches we simulate contraction of an ocean-continent transition as a consequence of far-field forces or forces arising from an existing subduction system (Marques et al., 2014), a mid-ocean ridges (Pascal & Cloetingh, 2009) and/or from mantle flow (Baes et al., 2018; Yamato et al., 2013). The initial geometric setup of the analogue and numerical experiments consists of an oceanic, a continental and a wedge shape transitional lithosphere in between (see chapter 2 and Figs. 4.1a, b). The latter represents the thinned continental lithosphere at the passive margin and will subsequently be referred to as “margin”.

The mechanical stratigraphy of the models is characterized by an alternation of brittle and viscous layers. Four in the case of continental (brittle-ductile-brittle-ductile) and two in the case of oceanic lithosphere (brittle-ductile), respectively (Burov & Watts, 2006). Continental lithosphere is decoupled at the base of the continental crust along a viscous layer that wedges out towards the ocean-continent transition (Figs. 4.1a, b). The margin’s crust is tapering off towards the oceanic lithosphere at an angle of $\sim 5^\circ$, which falls within the range of natural examples (e.g. $4\text{--}10^\circ$ the case of the Angolan margin). The wedge shape geometry of the crust at the passive margin implies that this domain coincides with the largest lateral strength gradients (Fig. 4.1b). In all our models, the

Figure 4.1: 2D sectional view of model setups, (a) analogue and (b) numerical, illustrating geometric, rheologic and kinematic conditions at the onset of shortening. Black arrow shows the directions of outflow (- V_{out}) and the velocity applied at the left boundary of the models (+ V_{in}). Strength envelopes of the continental lithosphere are shown on the right side of the section. The zoom in (b) shows details of the differential-stress and highlights the decoupling between the continental crust and the mantle at the margin. Stress calculation was computed using parameters displayed in Table 4.1 at a strain rate of 10^{-15} s^{-1} . Layering in the continental crust corresponds to passive strain markers.

strength of the lithosphere at passive margins is lower than in the ocean. Similar strength variations are predicted for natural examples (Maystrenko & Scheck-Wenderoth, 2009). The length scale of the margin (150 km) is chosen to be representative for natural cases, which typically range between 60 and 400 km (Bauer et al., 2000; Osmundsen & Redfield, 2011; Unternehr et al., 2010). Density contrasts between a mature oceanic lithosphere and the adiabatic mantle range between 40-80 kg/m³ (Afonso et al., 2007; Schubert et al., 2001). For this reason, a density contrast of 60 kg/m³ is implemented between the mechanical lithosphere and the asthenosphere creating a negative buoyancy of the oceanic lithosphere. In our study this density contrast is assumed to be constant

and we refer to Faccenna et al., (1999) and Mart et al., (2003) who discuss the importance of vertical and lateral (Maystrenko et al., 2013) density variations for the initiation of subduction zones at passive margins in detail.

In our models, the strength of the lithosphere is controlled by its thermal age, simulated using different plate thicknesses (Fig. 4.2c). The temperature and composition of the continental lithosphere is kept constant and a linear interpolation is applied between the base of oceanic and continental plate. Hence, by testing variable ages of oceanic lithosphere in the experiments, it is possible to investigate the impact of passive margins rheology on SI.

4.2.1 Analogue modelling specificities

The experiments were scaled according to the principles of geometric, kinematic, rheological and dynamic similarity (Hubbert, 1937; Ramberg, 1981; Weijermars & Schmeling, 1986; Sokoutis et al., 2000; Sokoutis et al., 2005) following the approach described in chapter 2, section 2.3. The model geometry and properties of the materials are displayed in figure 4.1 and listed in Table 4.1.

Each experiment is carried out with a constant rate of deformation, which is applied through movement of a wall that simulates a necessary horizontal force to maintain this deformation rate. In the series of models presented, the convergence velocities have been set to 0.5 cm/h.

The experiments are geometrically scaled by a length scale factor of 3.33×10^{-7} so that 1 cm in the model corresponds to 30 km in nature. Density is scaled based on the brittle oceanic lithosphere. The stress ratio is then calculated as product of the density ratio and the length-scale ratio (chapter 2, equation 2.13).

Brittle crustal and mantle layers exhibiting Mohr-Coulomb type behaviour are represented by sieved layers of dry feldspar and quartz sand, respectively (see Willingshofer et al., 2018 for a detailed description of material properties). Mixtures of silicone putty with fillers like sand and barium sulphate (BaSO_4) represent ductile layers. The addition of sand is required to reach suitable densities whereas the BaSO_4 is used to increase the viscosity. The power law-exponent of the ductile layers range between 1.09 and 1.40) exhibiting quasi-Newtonian behaviour.

We do not impose lateral changes in rheology in the lithospheric mantle in order not to pre-scribe the locus of deformation. As such the ductile lithospheric mantle layers are homogenous in thickness and composition under both the continental and oceanic crust. This assumption allows for addressing the mechanical effects of the thermal age of the lithosphere at passive margins (Fig. 4.2c) by varying the strength of the ductile and brittle layers in the lithospheric mantle.

4.2.2 Numerical modelling specificities

Thermo-mechanical models were used to simulate the shortening of a visco-elastic-plastic lithosphere. Details regarding the constitutive equations that are solved are provided in chapter 2. Deformation is governed by frictional, dislocation, diffusion and Peierls creep equations. The heat equation includes the effect of shear heating (Q_{sh})

Table 4.1: Summary of model configurations and material properties per analogue experiment (EXP) together with characteristic properties of natural prototypes. Here h is the thickness, η is the viscosity, ρ is the density of the layers and n is the stress exponent. F is the buoyancy number as detailed in chapter 1. AST. stands for the asthenosphere used in all the experiments.

ID	Oceanic brittle lithosphere			Continental brittle crust			Continental ductile crust			Continental brittle lith. mantle			Ductile lith. mantle			F
	model cm kg/m ³	nature km kg/m ³ Pa s		model cm kg/m ³	nature km kg/m ³ Pa s		model cm kg/m ³	nature km kg/m ³ Pa s		model cm kg/m ³	nature km kg/m ³ Pa s		model cm kg/m ³	nature km kg/m ³ Pa s		
1	$h=1.4$ $\rho=1500$	$h=42$ $\rho=3350$		$h=0.8$ $\rho=1300$	$h=24$ $\rho=2900$		$h=0.4$ $\rho=1300$ $\eta=4.6e4$ $n=1.12$	$h=12$ $\rho=2900$ $\eta=4.6e21$ $n=1.12$		$h=0.5$ $\rho=1500$	$h=15$ $\rho=3350$		$h=1.0$ $\rho=1500$ $\eta=1.7e4$ $n=1.09$	$h=30$ $\rho=3350$ $\eta=1.7e21$ $n=1.09$		3.00
	$h=1.4$ $\rho=1500$	$h=42$ $\rho=3350$		$h=0.8$ $\rho=1300$	$h=24$ $\rho=2900$		$h=0.4$ $\rho=1300$ $\eta=4.6e4$ $n=1.12$	$h=12$ $\rho=2900$ $\eta=4.6e21$ $n=1.12$		$h=0.5$ $\rho=1500$	$h=15$ $\rho=3350$		$h=1.0$ $\rho=1500$ $\eta=5.8e4$ $n=1.13$	$h=30$ $\rho=3350$ $\eta=5.8e21$ $n=1.13$		0.85
3	$h=1.4$ $\rho=1500$	$h=42$ $\rho=3350$		$h=0.8$ $\rho=1300$	$h=24$ $\rho=2900$		$h=0.4$ $\rho=1300$ $\eta=4.6e4$ $n=1.12$	$h=12$ $\rho=2900$ $\eta=4.6e21$ $n=1.12$		$h=0.5$ $\rho=1500$	$h=15$ $\rho=3350$		$h=1.1$ $\rho=1500$ $\eta=8.4e4$ $n=1.2$	$h=33$ $\rho=3350$ $\eta=8.4e21$ $n=1.2$		0.65
	$h=1.4$ $\rho=1500$	$h=42$ $\rho=3350$		$h=0.8$ $\rho=1300$	$h=24$ $\rho=2900$		$h=0.4$ $\rho=1300$ $\eta=4.6e4$ $n=1.12$	$h=12$ $\rho=2900$ $\eta=4.6e21$ $n=1.12$		$h=0.5$ $\rho=1500$	$h=15$ $\rho=3350$		$h=1.0$ $\rho=1500$ $\eta=1.4e5$ $n=1.36$	$h=30$ $\rho=3350$ $\eta=1.4e22$ $n=1.36$		0.35

Table 4.1 (page 2)

5	$h=1.1$ $\rho=1500$	$h=33$ $\rho=3350$	$h=0.8$ $\rho=1300$	$h=24$ $\rho=2900$	$h=0.4$ $\rho=1300$ $\eta=4.6e4$ $n=1.12$	$h=12$ $\rho=2900$ $\eta=4.6e21$ $n=1.12$	$h=0.5$ $\rho=1500$	$h=15$ $\rho=3350$	$h=1.0$ $\rho=1500$ $\eta=5.8e4$ $n=1.13$	$h=30$ $\rho=3350$ $\eta=5.8e21$ $n=1.13$	0.75
6	$h=0.8$ $\rho=1500$	$h=24$ $\rho=3350$	$h=0.8$ $\rho=1300$	$h=24$ $\rho=2900$	$h=0.4$ $\rho=1300$ $\eta=4.6e4$ $n=1.12$	$h=12$ $\rho=2900$ $\eta=4.6e21$ $n=1.12$	$h=0.5$ $\rho=1500$	$h=15$ $\rho=3350$	$h=1.0$ $\rho=1500$ $\eta=5.8e4$ $n=1.13$	$h=30$ $\rho=3350$ $\eta=5.8e21$ $n=1.13$	0.60
7	$h=1.7$ $\rho=1500$	$h=51$ $\rho=3350$	$h=0.8$ $\rho=1300$	$h=24$ $\rho=2900$	$h=0.4$ $\rho=1300$ $\eta=4.6e4$ $n=1.12$	$h=12$ $\rho=2900$ $\eta=4.6e21$ $n=1.12$	$h=0.5$ $\rho=1500$	$h=15$ $\rho=3350$	$h=1.4$ $\rho=1500$ $\eta=1.4e5$ $n=1.36$	$h=42$ $\rho=3350$ $\eta=1.4e22$ $n=1.36$	0.40
AST.	$h=10$ $\rho=1470$ $\eta=8e2$ $n=1$	$h=3000$ $\rho=3290$ $\eta=8e19$ $n=1$									

and radiogenic heat production (Q_r). The presented models do not account for any predefined strain softening parameterization. Similarly to the analogue approach, the numerical thermo-mechanical models were designed based on a 2D initial setting inspired from a magma-poor margin (Unternehr et al., 2010). The behaviour of the layers was simulated by a visco-plastic rheology with parameters displayed in Table 4.2. The model consists of a two-layer continental lithosphere made of a 35 km crust and a lithospheric mantle with a thermal thickness (h_l) of 150 km. The oceanic lithosphere consists of an 8 km crust and a lithospheric mantle and has a thermal thickness that depends on its age. Dry granite, dry olivine and dry diabase rheology are used for the continental crust, the mantle lithosphere and the oceanic crust, respectively (Table 4.2).

To obtain a rheology of the oceanic lithosphere depending on the thermal age (a_L), the thermal base can be approximated by using the half-space cooling function, which writes:

$$H_L = 2 \operatorname{erfc}^{-1}(\Delta T) \sqrt{\chi t} = -c_1 \sqrt{a_L} \quad (4.1)$$

$$\text{and,} \quad q_0 = 2 k T_m / \sqrt{\pi \chi t} = c_2 / \sqrt{a_L} \quad (4.2)$$

$$\text{so,} \quad H_L = -c_1 \frac{c_2}{q_0} \quad (4.3)$$

By assuming that the thermal base experiences 20% of temperature anomaly with respect to the temperature in the mantle ($\Delta T = 260\text{K}$) and a thermal diffusivity of $\chi = 10^{-6} \text{ m}^2.\text{s}^{-1}$, $c_1 = 14\,500 \text{ m}.\text{Myr}^{-\frac{1}{2}}$ and $c_2 = 473 \text{ mW}.\text{m}^{-2}.\text{Myr}^{\frac{1}{2}}$ (Artemieva, 2011). The half-space cooling model successfully explains the observed heat flow and bathymetry for ocean floor younger than 80 Myr (Turcotte & Schubert, 2014). However, this model fails to predict the thickness of the lithosphere at old ages ($> 80 \text{ Myr}$) where lithosphere thickness approaches an equilibrium, time-independent thermal structure. To resolve this issue, the $\sqrt{a_L}$ is better constrained by using another variable depending on the surface heat flow. We used the model defined by Cheng (2016) which fits the age of the sea floor with q_0 (in $\text{mW}.\text{m}^{-2}$) at the surface:

$$q_{01} = 433.63 a_L^{-0.445} \quad \text{for young oceanic lithospheres}$$

and

$$q_{02} = 224.03 a_L^{-0.307} \quad \text{for old oceanic lithospheres}$$

and we chose

$$q_0(a_L) = q_{01} \text{ when } q_{01} > q_{02}$$

and

$$q_0(a_L) = q_{02} \text{ when } q_{01} < q_{02} \quad (4.4)$$

By using this method, the thermal base of the lithosphere as a function of age follows the same trend as the half-space cooling model for ages $< 80 \text{ Myr}$, but subsequently plateaus off for ages $> 80 \text{ Myr}$ (Fig. 4.2c).

The model domain is a section of $2000 \times 500 \text{ km}$ and the numerical resolution is 500 m in both dimensions. The model top boundary is a true free surface (Duretz et al., 2016b). A constant normal velocity is applied at the left boundary of the model

($|V_{in}| = 1$ cm/yr) from the surface to a depth of 200 km. In order to satisfy mass conservation, an outflow velocity (V_{out}) is distributed at the base of the model and on the sides from 200 km deep to the base of the box (Fig. 4.1b, $H_{out1,2,3}$). The outflow is proportionally distributed over the length of each boundary:

$$V_{in}H_{in} = V_{out} \sum H_{out} \quad (4.5)$$

The initial geotherm is computed assuming steady state and accounts for different radiogenic heat productions in each layer (Table 2) and a constant asthenosphere temperature of 1330°C. The resulting temperature at the continental Moho is $\sim 425^\circ\text{C}$. We consider the mechanical base of the lithosphere as the base of the lithosphere that can support stresses (Cloetingh & Burov, 1996; Burov & Diamant, 1995). The mechanical of the lithosphere is set at t_0 , at the 1000°C isotherm which fits with the base of the strength envelope.

Table 4.2: Rheological and thermal values used for the numerical models. References for rheology are R_95: Ranalli (1995) ; H&C_83: Hansen & Carter, 1983; H&K_04: Hirth and Kohlstedt (2003b) ; M_98: Mackwell et al., 1998.

	ρ (kg.m ⁻³)	k (W.m ⁻¹ .K ⁻¹)	Qr (W.m ⁻³)	ϕ (°)	A (Pa ⁻ⁿ .s ⁻¹)	n	Q (J.mol ⁻¹)	Ref.
Continent crust (dry granite)	2900	2.7	0.8e-6	30	3.16e-26	2.3	154e3	H&C_83
Strong lower crust (diabase)	2900	2.7	0.8e-6	30	8.04e-25	3.4	260e3	R_95
Oceanic crust (Diabase)	2900	3	1.0e-10	30	5.04e-28	4.7	485.0e3	M_98
Mantle lithosphere (dry olivine)								
Disl. creep	3360	3.2	1.0e-10	30	1.10e-16	3.5	530.0e3	H&K_03
Dif. creep					1.50e-15	1.0	375.0e3	H&K_03
Asthenosphere	3300	3.2	1.0e-10	30	1.10e-16	3.5	530.0e3	H&K_03

Here p is the density, k is the thermal conductivity, Qr is the radiogenic heat production, ϕ is the friction angle, A is a pre-factor, n is the stress exponent, Q is the activation energy. The shear modulus G is fixed to $3e10$ Pa.

4.2.3 Combined approach

To respect the stress-scale ratio, we adapt the material in our analogue models by scaling the stress envelopes obtained in our numerical models using flow law parameters derived from laboratory experiments (i.e., plastic, exponential-dislocation-diffusion creep). Thus, for a given age of the oceanic lithosphere, the strength and thickness of the ductile and brittle layers used in analogue models reflects the strength and thickness of the respective layers used in our numerical models (Fig 4.2c). As the

brittle strength is function of normal stress, the differential stress increases linearly with depth and the brittle layers are scaled to brittle layer thicknesses in nature.

For scaling of the ductile layer, we consider the average strength of the ductile layer over its thickness (Fig. 4.2a). From the strength envelopes calculated for our numerical models, the mechanical base (hm) of the lithosphere represents a suitable lower boundary for the ductile layer. However, the location of the brittle-ductile transition remains unclear. By setting the stress boundary at the base of the brittle layer for a shortening stress regime, the overall mean strength of the ductile layer is unreasonably strong. For this reason, we consider the semi-brittle domain where the exponential creep (Peierls) strain is active as the domain where the brittle-ductile transition occurs (Aharonov & Scholz, 2019). The brittle-ductile transition (BDT) for our analogue model is then chosen as the middle of this domain (Fig. 4.2a, grey area). We reckon that this approximation is more realistic as it reduces the strength of the mantle lithosphere and accounts for an intermediate mixed mode of deformation in nature.

The averaged ductile strength $\overline{\Delta\sigma}$ is calculated in between the BDT and the hm as defined above and is then scaled to laboratory scale. A target value for the effective viscosity is calculated using the strain rate of the experiment (referred as scaled value in table 4.3). The material for the ductile layers in the analogue experiments is selected to reflect the scaled target value of viscosity, for a given strain rate (Table 4.3). As an example (see Fig. 4.2c), in the numerical model, the average viscosity in the ductile layer of a 70 Myr lithosphere is about $\eta \simeq 1.0 \times 10^{22}$ Pa s. Thus, the viscosity ratio is $\eta^* \sim 1.0 \times 10^{-17}$. This yield to a time ratio $t^* = \frac{\eta^*}{\sigma^*} = 7.20 \times 10^{-11}$ and so, the 20 hours of the experiment represents ~ 32 Myr in nature, the 0.5 cm/h convergence rate represents ~ 1 cm/yr in nature.

Following Faccenna et al. (1999) we calculate the buoyancy number F (see chapter 2) to infer the of role of buoyancy forces with respect to the viscous forces acting on the ductile lithospheric mantle (Fig. 4.2b).

Table 4.3: Summary of average lithospheric ductile mantle rheological values, laboratory scaled values, and actual values used in the experiments. Three experiments were selected to represent oceanic lithosphere with various thermal ages (first column).

Id		$(\overline{\Delta\sigma})_{\text{ductile}}$ (Pa)			$\bar{\eta}$ (Pa s)		
nature	analogue	nature	scaled values	experiments	nature	scaled values	experiments
30 Myr	exp6	1,07e8	16	14	6.4e21	6.4e4	5.8e4
70 Myr	exp3	1,67e8	25	23	9.0e21	9.0e4	8.4e4
140 Myr	exp7	2,54e8	38	40	1.3e22	1.3e5	1.4e5

Here, the values represent the average strength $\overline{\Delta\sigma}$ and the average viscosity $\bar{\eta}$ of the ductile lithospheric mantle, computed with the scaling approach described in section 4.2.4. Scaled values were calculated using the naturel value, scaled with the stress and viscosity ratio.

Figure 4.2: (a) Strength profiles illustrating rheological layering of a 70 Myr oceanic lithosphere and the deformation mechanism implemented in the numerical study. Extrapolated strength envelopes computed for the analogue model 6 by using scaling approach described in section 2.4. BDT refers to the brittle-ductile transition. (b) Graph showing the parameter space explored by the analogue and numerical models as a function of mean ductile and mean brittle differential stress of the oceanic lithosphere. Analogue models address the influence of the ductile and brittle strength of the oceanic lithosphere, respectively. $\Delta\sigma$ and $\overline{\Delta\sigma}$ are the differential stress and the mean differential stress. Subscripts a and n refer to analogue and numerical. (c) Strength profiles for different oceanic lithospheres tested as function of age and thermal structure. The mechanical strong part of the lithosphere (in blue) is used as proxy for the thickness of the brittle layer and the yellow one for the thickness of the ductile layer in the analogue experiments. The red envelopes represent the strength for the analogue model 6, 3 and 7.

4.3 Analogue modelling results

In order to investigate favourable parameter combinations for SI, we present experimental results in figure 4.3. Two sets of experiments document the influence of the strength of the ductile lithospheric mantle (set 1), and the strength of the brittle oceanic lithosphere (set 2), regulated through the thickness of its brittle part. Then, we compare the results with experiments representing an increase of the thermal age of the oceanic lithosphere (set 3), characterized by an increase of the two above-mentioned parameters. This modelling strategy facilitates exploring the relationship between lithosphere thermal age and locus of deformation. Deformation of the experiments is monitored using digital photography and laser scanning of the model surface. The observed deformation pattern as well as the final tectonic internal structure are analysed and interpreted using cross-sections retrieved at the end of the experiments (Fig. 4.3).

4.3.1 Effect of lithospheric mantle ductile strength

To simulate the increase in ductile strength at the base of the lithosphere with age (Fig. 4.3h, set 1), the viscosity of silicone putty representing the lithospheric mantle was increased. The thickness of the brittle layer remained unchanged to isolate the effect of ductile strength. As explained in previous sections, the ductile strength strongly affects the buoyancy of the lithosphere as the viscous strength resists the gravitational normal stress. The evolution of experiments 1, 2 and 4 has been described in detail in chapter 3. Here we only highlight key features for easier comparison with experiments where the brittle strength of the oceanic lithosphere is the main variable.

In case of a lithosphere characterized by a low viscosity lithospheric mantle (experiment 1, Fig. 4.3a), intra-oceanic underthrusting takes place with the development of a thrust parallel to the continent-ocean boundary. Deformation remains within the oceanic lithosphere throughout the duration of the experiment and does not affect the margin. Intra-oceanic thrusting is characterized by an in-sequence change of thrust vergence resulting in vertical stacking and sinking of thrust sheets (see cross section, Fig. 4.3a). However, the Newtonian behaviour of the low strength ductile mantle does not lead to localised deformation to form a subduction plate boundary-type structure, but allows for vertical sinking of the brittle blocks and the development of pop-down structures. The frequent changes of thrust polarities also impede unidirectional subduction.

In case of an intermediate viscosity of the lithospheric mantle (experiments 2 and 3) deformation predominantly affects the passive margin. Early stage deformation (5-10% BS) involves thrust initiation at the ocean-continent transition and the continent-ward propagation of the thrust system leading to the formation of a flexural basin at the margin (Figs. 4.3b, c). Localization of deformation at the down-flexed passive margin into a major thrust (e.g. fault 3, Fig. 4.3b) marks the underthrusting stage. Subsequent shortening is accommodated by further underthrusting of the oceanic plate dragging down the continental mantle lithosphere to form a double-sided subduction (Fig. 4.3b, c).

Experiment 4 with a high viscosity mantle lithosphere also evolves by deformation of the passive margin through thrusting up to 10% BS (Fig. 4.3d). Thereafter, shortening is predominately accommodated by folding of the passive margins and adjacent lithospheres as a consequence of the high shear resistance within the lithospheric mantle that prevents strain localization (Fig. 4.3d).

These results highlight the importance of the ductile strength of the lithospheric mantle, which controls (1) where deformation localizes (at the passive margin or within the oceanic domain) and (2) the style of deformation and thus the extent of underthrusting of the oceanic lithosphere at the passive margin.

4.3.2 Effect of brittle strength of the oceanic lithosphere

As suggested in a previous study (e.g. Nikolaeva et al., 2010) and in experiment 2, underthrusting at the passive margin occurs by deformation of the continental crust but can be controlled by the shear resistance in the ductile mantle lithosphere. In this section, we assess the importance of strength contrasts at passive margins, which are implemented at the level of the brittle crust through reducing its thickness (Fig. 3h, set 2). The viscosity of the lithospheric mantle remains unchanged. Three experiments with different thickness of the brittle layer are compared: experiment 2 (1.4 cm), experiment 5 (1.1 cm), and experiment 6 (0.8 cm). The evolution of experiment 2 has already been described in context of the previous section and will not be repeated here.

As described above, experiment 2 leads to inversion of the passive margin and underthrusting of the oceanic lithosphere underneath the passive margin. In case of thinner oceanic lithospheres, such as in experiments 5 and 6, deformation also starts at the passive margin by thrusting, at a position where the ductile lower crust of the margin wedges-out (e.g. thrust 1 at 5%BS, Fig. 3f). In contrast to experiment 2, deformation jumps thereafter to the oceanic domain, as documented by the formation of fault 2 (Fig. 4.3f). In experiment 5, this structure cuts through the brittle oceanic lithosphere in opposite direction to thrust 1 thereby leading to a phase of obduction of the oceanic plate onto the continental margin (20% BS and cross section, Fig. 4.3f).

Results of this set of experiments show that for a low strength oceanic plate, (1) deformation can still initiate at the margin but (2) then jumps to the oceanic domain where it resides. We also note that that intra-oceanic underthrusting takes place close to the margin where the deformation localizes during the early stage of deformation.

4.3.3 Effect of the thermal age of oceanic lithospheres

The experimental results, described in the previous sections, highlight the importance of the strength of both brittle and ductile oceanic lithosphere for the locus of subduction initiation and the underthrusting process. However, by increasing the thermal age, both brittle and ductile strength increase (Fig. 4.2c). Here we present the result of experiments, referred to as set 3, (experiments 6, 3 and 7) where both, brittle and ductile strength for the oceanic lithosphere have been consecutively increased. The results are summarized in figure 4.4.

Figure 4.3: Experimental results of the analogue study. Experiments 1 to 7 are represented in corresponding panels (a, b, c, d, e, f, g). Topographic evolution is retrieved through surface scans at 0%, 5%, 10%, 15%, and 20% bulk shortening (BS). Vertical exaggeration is 10 times. The final scan is supported by a lithospheric scale cross-section showing internal tectonic structures. The numbers at the surface of the scans represent the apparition order of the different tectonic structures. PM: passive margin. (h) Matrix showing the parameter space explored in the analogue modelling study as well as the strength profiles for the third set of experiments. BCC, DCC, BM, DM denote brittle continental crust, ductile continental crust, brittle mantle, ductile mantle, respectively.

Figure 4.3 (page 2)

Shortening of an oceanic lithosphere with an age of ~30 Myr is represented by experiment 6 (Figs. 4.3f and 4.4a). The model evolution follows the following pattern: (1) strain localization at the passive margin, (2) abortion of underthrusting and migration of deformation to the oceanic domain and (3) incipient obduction of oceanic crust onto the continental margin.

Shortening of an oceanic lithosphere with an age of ~60-70 Myr is represented by experiment 3 (Figs. 4.3c and 4.4b). The model development follows the following pattern: (1) strain localization at the passive margin and development of a thrust system, (2) rupture of the flexural basin and underthrusting of the oceanic lithosphere by using the continental ductile crust as a decollement and (3) deformation of the ductile lithospheric mantle that accommodates the descent of the subducting plate.

Contraction of an oceanic lithosphere with an age of ~140 Myr is represented by experiment 7 (Figs. 4.3g and 4.4c). This experiment represents an end-member case of very high strength of the oceanic lithosphere, characterised by a thick brittle lithosphere and a high viscosity in the ductile lithosphere. Overall, it is characterized by lithospheric folding at the margin and the adjacent lithospheres. The model development is similar to experiment 3. It follows the following pattern of: (1) strain localization and underthrusting at the margin with the development of a frontal thrust system, (2) abortion of underthrusting and (3) large amplitude buckling of both the continental and the oceanic lithosphere in response to strong resistance of the ductile lithospheric mantle. In addition, lithosphere buckling in the oceanic domain is associated with low-offset thrusting (fault 3', Fig. 4.3g).

Figure 4.4: Summary of deformation styles and evolution found in the analogue parametric study: a) intra-oceanic subduction for a young oceanic lithosphere, b) subduction initiation at the passive margin for an intermediate strength, and c) lithospheric folding for a strong oceanic lithosphere.

4.4 Numerical modelling results

To investigate the impact of the rheology of the oceanic lithosphere on subduction initiation, numerical experiments have been carried out by changing thermal ages of the oceanic lithosphere, ranging from 20 Myr to 140 Myr. In figure 4.5 we show the results of experiments representing young (30 Ma), intermediate (70 Ma) and old (140 Ma) oceanic lithosphere. These numerical experiments show distinct deformation evolutions and three end-member results similar to the results obtained in analogue experiments: (1) intra-oceanic underthrusting for young oceanic lithosphere (2) subduction at the passive margin for intermediate age of oceanic lithosphere, and (3) buckling/deformation of continental lithosphere at the passive margin for old oceanic lithosphere. In this section, we describe the evolution of these end-member models. Additionally, we show in figure 4.6 the role of key parameters relevant for strain localization, including the coupling between the crust and the mantle lithosphere at the passive margin, the importance of shear heating and Peierls creep mechanism.

4.4.1 Effect of the thermal age of oceanic lithospheres:

In experiments that are representative for young oceanic lithospheres with thermal ages ranging between 20 and 40 Myr, early stages of deformation affect the oceanic lithosphere as well as the passive margin, as shown by shear zone type structures with elevated strain rates (Figs. 4.5a, b, c). Reference model 1 shows that the margin is affected by both plastic and ductile deformation in the upper and lower crustal layer, respectively. After 7 Myr of shortening, deformation predominantly affects the oceanic lithosphere, where buckling instabilities develop with major shear deformation being concentrated at inflexion points of the buckle folds (Fig. 4.5a, b). We note that deformation is gradually distributed along the oceanic domain, attaining a peak value near the margin (Fig. 4.5b). During the advanced stages of shortening, deformation localizes at the passive margin, where the highest strain-rates are observed at the base of the ductile crust (Fig. 4.5c). However, this strain concentration does not lead to the development of a subduction-like fault. Instead, under-thrusting occurs through the development of a pop-down structure in the oceanic lithosphere near the OCT.

The locus of strain and style of deformation changed in the case of oceanic lithosphere older than 40 Myr. In the model representing an age of 70 Myr (Figs. 4.5d, e, f), we observe in early stages of the model evolution that deformation localizes at the base of the ductile crust of the continental margin, where a shear zone initiates (Fig. 4.5d). Simultaneously, shortening is accommodated by thrusting within the brittle crust (Fig. 4.5e). After ca. 10 Myr of contraction, deformation is transmitted from the lower crust to the lithospheric mantle, first in the semi-brittle domain (Peierls creep), and then to the ductile one (dislocation creep). The effect of this mechanism is presented in figure 4.6, that shows the locus of deformation in relation to deformation mechanisms. At the top-part of the mantle lithosphere, Peierls mechanisms becomes activated (Fig. 4.6c) whereas ductile deformation in the lowermost part of the lithosphere is governed by dislocation creep mechanisms. At the same location, shear heating reaches its maximum value (Fig. 4.6d). At this stage, mapping of the velocity field in figure 4.5e highlights

Figure 4.5: Model evolution for the three reference numerical models involving three different ages of oceanic lithosphere. From left to right: 30 Myr, 70 Myr and 140 Myr. Evolution of lithological and temperature fields are depicted for the following shortening time steps: 7 Myr, 15 Myr, 25 Myr. For each time step we also show the distribution of strain-rate to emphasize strain localization. White arrows represent the velocity field.

that the transfer of deformation from the crust to the mantle induces deflection of the oceanic lithosphere under the passive margin. After 15 Myr of contraction, the formation of a shear zone at the plate interface leads to underthrusting of both the oceanic lithosphere and the margin, resulting in the formation of a slab after ~25 Myr of shortening (Figs. 4.5e, f). In surface, a growing fold and thrust belt is formed in the forearc region in response to underthrusting (Fig. 4.5e).

For an oceanic lithosphere older than 100 Myr, underthrusting is deterred by shear resistance in the lithospheric mantle. The early evolution of this model shows first order similarities with experiments of intermediate age of oceanic lithosphere (Fig. 4.5g) and includes (1) strain localized at the passive margin, (2) formation of a shear zone at the base of the continental ductile crust, which acts as a décollement and (3) transfer of deformation to the mantle underlying the wedge shape margin. However, after 20 Myr of shortening, the strong resistance of the lithospheric mantle restricts the strain localization in the lithospheric mantle. The strain is mainly distributed in the continental domain and underthrusting is prevented. After 15 Myr of shortening, contraction leads to the formation of conjugate shear zones in the lithospheric mantle that are referred to as S1 and S2 in figure 4.5h. In the last stage, after 25 Myr of shortening, failure of the lithospheric mantle occurs at shear zone S1 leading to the subduction of the continental

Figure 4.6: Respective contributions of (a) frictional- (b) dislocation- (c) Peierls creep deformation mechanisms on strain rates for the reference model with a 70 Myr old oceanic lithosphere. (d) shows the magnitude of shear heating. The time step is at 12 Myr to show deformation mechanisms during underthrusting at a margin.

lithospheric mantle under the oceanic lithosphere. However, with ongoing shortening, the oceanic lithosphere penetrates through the decollement (Fig. 4.5i), leading to the decoupling of the mantle from the crust, referred as “partial delamination” by Nikolaeva et al. (2011). Crustal deformation in the continental lithosphere is then accommodated by long wavelength folding and associated shear band formation.

4.4.2 Effect of crust-mantle coupling

Several studies (e.g., Nikolaeva et al., 2011) highlight that the presence of a continental ductile lower crust is key for underthrusting at passive margins. To test the role of crust-mantle coupling at passive margins for strain localization, we run a model that includes a strong crust, leading to coupling between the crust and the lithospheric mantle (Fig. 4.7a). This model confirms that the decoupling horizon at the base of the margin crust is essential for strain localisation and underthrusting at the margin as under coupled conditions the amount of underthrusting is very limited. Strain localizes at passive margin, but the overall strength of the lithosphere prevents plate rupture. Subsequently deformation shifts to the oceanic domain where the lithosphere buckles and pop-down structures may develop. Thus, crust-mantle coupling at passive margins strongly favours intra-oceanic strain localization for oceanic lithospheres younger than 60 Myr, whereas it leads to lithospheric folding for oceanic lithosphere older than 80 Myr. These additional results are available in the appendix B to this chapter and are summarized in figure 4.8b.

4.4.3 Effect of Peierls creep

When Peierls creep of olivine is not considered, larger stresses are allowed to build up in the mantle. The overall lithosphere strength thus increases by about 30%. In these conditions, neither underthrusting nor subduction initiation occurred regardless of the age of the oceanic lithosphere. As an example, for a plate age of 70 Myr, deformation localises at the base of the continental crust but never propagates into the mantle (Fig. 4.7b). Instead, intra-oceanic underthrusting occurs, characterized by the development of a pop-down structure (Fig. 4.7b).

4.4.4 Effect of shear heating

The modelling results underline the importance of thermal softening for shear localization in the lithospheric mantle during underthrusting (Fig. 4.6). In the absence of shear-heating, underthrusting occurs at the passive margin, but it never reaches the subduction stage. The modelling results suggests that shear heating does not change the place where deformation starts but it is essential for keeping deformation localized on individual structures like those along which underthrusting takes place. Neglecting shear heating typically leads to double-sided subduction involving both the oceanic and continental lithospheres (Fig. 4.7c) similar to the analogue modelling experiments (see experiment 2, fig. 4.3b).

Figure 4.7: Effects of variations in rheological condition for a 70 Myr old oceanic lithosphere at the onset of shortening. For the color coding see Figure 4.1. White lines correspond to isotherms (200°C spacing). Model time is 30 Myr after onset of shortening. (a) model with shear heating, Peierls creep and a strong continental crust leading to crust-mantle coupling at the margin. (b) model with shear heating but without olivine Peierls creep. (c) model with olivine Peierls creep but deactivated shear heating. For b and c continental crust is the same than for the reference model.

4.5 Discussion

Both analogue and numerical modelling results show first order similarities and emphasize that subduction initiation at or near passive margins follow a 3-stages tectonic evolution including (1) early-stage deformation of the continental passive margin crust (2) underthrusting of the oceanic plate and (3) subduction. We note that underthrusting characterizes the major change from distributed to localized deformation in the mantle lithosphere. We refer to “subduction” when the failure of ductile lithospheric mantle is complete and the slab sinks into the asthenosphere. These deformation stages are

comparable with evolutionary schemes proposed by Nikolaeva et al. (2010) or Faccenna et al. (1999). In the following sections, we discuss the conditions leading to subduction initiation and summarize the key outcomes of the study in figure 4.8. To correlate analogue and numerical results, we focus on the mean ductile and brittle strength calculated over their respective thicknesses as explained in section 4.2.4. We remark that the analogue experiments are scaled such that the brittle-ductile transition lies slightly deeper, and that therefore the overall brittle strength is higher compared to the numerical models (Fig. 4.8a). To highlight the role of thermo-mechanical weakening processes, we display domain diagrams showing the effect of each studied parameter on the model development as a function of oceanic lithosphere age at the onset of shortening (Fig. 4.8c).

4.5.1 Mechanisms of strain localization for subduction initiation

Several studies (e.g., Faccenna et al., 1999; Goren et al., 2008; Nikolaeva et al., 2011), as well as this one (Fig. 4.5e), show that the locus of subduction initiation is controlled by the strength contrast at the margin which depends on compositional and rheologic differences between the oceanic and continental lithospheres. In nature, this lateral strength contrast is evidenced by recurrent reverse-faulting (Stein et al., 1989b; Pereira et al., 2011; Ziegler & Cloetingh, 2004) and increased seismicity at the ocean-continent boundary (Assumpção & Sacek, 2013). In addition, our analogue and numerical experiments show that the ductile crust at the margin acts as a decollement and facilitates the development of thrusts at the passive margin, similar to what has been described for collisional systems upon continental subduction (Willingshofer et al., 2013). When deformation has been localized, the presence of a decollement at the base of the margin is key for the development of a long-lasting shear zone that eventually propagates into the mantle lithosphere (Fig 4.6). Analogue modelling results highlight that the extent of underthrusting is controlled by the strength contrast between the brittle part of the subducting plate and the ductile mantle lithosphere (Fig 4.8a). An increase of lithospheric mantle ductile strength decreases the likelihood of underthrusting to evolve to subduction (Exp. 2 & 4). A similar result is obtained when decreasing the thickness of the brittle layer but keeping the viscosity unchanged. Suitable conditions for underthrusting include an intermediate age of oceanic lithosphere. However, when subduction does not initiate at the margin, deformation shifts either to the oceanic domain (young lithosphere) or to the continental domain (old lithosphere). Our numerical models including age-dependant rheologies are consistent with these results. We also note that intra-oceanic strain localization is characterized by the development of pop-down structures, which are important during the shortening of hot and weak lithospheres (Poh et al., 2020). Although we obtained intra-oceanic subduction-like behaviour through the formation of pop-down structures, sustaining subduction over long periods of time (>10 Myr) would require the inclusion of stabilizing mechanism at the plate interface such as sediments or a serpentinite mantle wedge (see chapter 5 or Sobolev & Brown, 2019; Wada et al., 2008). The slab pull would also be enhanced by densification of the mantle during eclogitisation (Jolivet et al., 2013).

Figure 4.8: Summary of model results from both analogue and numerical experiments. Three deformation styles have been found and are presented at the top of the figure: Intra-oceanic deformation for a young oceanic lithosphere, subduction initiation at the passive margin for an intermediate age, and continental deformation for an old oceanic lithosphere. (a) Effects of mean brittle and ductile strength of the oceanic lithosphere on strain localization. Yellow dots correspond to analogue experiments and each number corresponds to an experiment number. White dots correspond to numerical experiment with the age of the oceanic lithosphere increasing from the bottom to the top (b) Domain diagrams showing the effect of each studied parameter on the model development as a function of oceanic lithosphere age at the onset of shortening (c) Graph showing the distribution of the buoyancy number (F) of the oceanic lithosphere for each experiment as a function of the mean ductile strength.

In addition to lateral changes of rheology, it has been shown that density contrasts across the OCT might significantly contribute to SI (Faccenna et al., 1999; Mart et al., 2005; Nikolaeva et al., 2011). Expressing density contrasts in terms of the buoyancy number reveals that deformation preferentially affects the oceanic lithosphere for buoyancy numbers > 1 (analogue experiment 1). This is in agreement with the findings of Faccenna et al. (1999). However, strength calculations coupled to cooling oceanic lithosphere suggests that the oceanic lithosphere never reaches a suitable buoyancy conditions to overcome the ductile strength of the lithosphere, even for density contrasts in the order of 60 kg.m^{-3} (Fig. 4.8c). Therefore, in case of induced convergence at passive margins we consider density contrasts as a contributing, but not a dominant parameter for plate rupture and underthrusting. It can however be an important parameter during the sinking of the subducting plate during the subduction stage.

4.5.2 Role of thermal softening and Peierls creep

Although our analogue and numerical experiments provide first order similarities, the differences lie mainly in the weakening mechanisms. In this sense, with the lack of intrinsic mechanisms leading to shear localization for stress levels as obtained in our analogue experiments, the sustainability of the subduction is uncertain. As reported, this limitation has been overcome by numerical modelling of which we summarize the importance of weakening mechanisms in figure 4.8b and the following paragraphs.

Models without shear heating and Peierls creep rheology do not lead to the initiation of a subduction zone. In our models Peierls creep mechanism reduces the lithosphere mantle strength by $\sim 30\%$, which is necessary to trigger subduction at a passive margin (Fig. 4.8). In absence of Peierls creep mechanisms, underthrusting occurs in the oceanic domain. Such results confirm the idea that Peierls mechanism, acting as a stress limiter, enhances strain localization in the lithospheric mantle (Kameyama et al., 1999). Additionally, there is a strong correlation between analogue experiments, scaled from rheologies including Peierls creep, and numerical experiments, including Peierls creep rheology and no shear heating. Although the Newtonian mantle in our models impedes the development of shear zone, figure 4.7c shows that a power law rheology does not make a major difference. In models with no shear heating, underthrusting occurs but deformation never localizes in the ductile mantle lithosphere (Fig. 4.7c). Therefore, an overall weaker mantle lithosphere is not sufficient for shear localization in the mantle lithosphere, which requires thermal softening.

In the simulations, temperature increase in the ductile mantle lithosphere due to shear heating reaching $\sim 120^\circ\text{C}$, which is similar to previous modelling results (Jaquet & Schmalholz, 2018; Schmalholz & Duretz, 2015). Moreover, Peierls creep and shear heating reach maximum values at the same time and same location during underthrusting at the passive margin (Figs. 4.6c, d). It has been suggested that shear heating is the main mechanism controlling the thermal activation of ductile shear zone in the mantle lithosphere (Kiss et al., 2020). In addition, our models of SI provide a good example of the interdependence between Peierls creep and shear-heating mechanism. In the frame

of this study, we then conclude that, in the absence of pre-defined mantle structures, subduction initiation is physically feasible by the evolution of major shear zones in the crust that propagates into the mantle where thermally induced shear heating processes within Peierl's creep domain facilitate the formation of a subduction.

Although not tested within this study, we acknowledge the potential role of grain size-sensitive creep (Ricard & Bercovici, 2009), percolating fluids for weakening of the system fluids (Alvarez-Marron et al., 1997; Regenauer-Lieb et al., 2001) as well as the role of serpentinitisation at passive margins where the mantle lithosphere has been exhumed (Bayrakci et al., 2016; Hirauchi et al., 2016).

4.5.3 Timing and locus of subduction initiation

This study proposes that shortening of a young margin lithosphere, ($a_L < 50$ Myr) gives rise to intra-oceanic strain localization (e.g. analogue experiments 6, Fig. 4.3d). The low strength of the oceanic lithosphere does not permit deformation of the mantle lithosphere at the margin and deformation shifts toward the oceanic domain, and in our models, close to the margin (e.g. Fig. 4.3d). It leads to a phase of obduction on the continental margin. The locus of subduction initiation close to the margin is in agreement with several examples of obduction (e.g. Agard et al., 2016; van Hinsbergen et al., 2016). These studies point out that the short time span between supra-subduction spreading centres and ophiolites emplacement is an evidence of subduction initiation close to the margin. A good example could be the subduction initiation in the Neo-Tethys Ocean between Gondwana and Eurasia. Field-based tectonic reconstructions of Maffione et al. (2017) and van Hinsbergen et al. (2016) suggest that the Late Cretaceous subduction system formed parallel and near (~100 km) the main continental margins of the Anatolide-Tauride block. Strain localization would then be governed by pre-existing weakness zones such as detachment faults at oceanic core-complexes (Maffione et al., 2015). However, in this reconstruction, the age of the oceanic lithosphere at the margin is believed to be of at least 100 Myr, which is significantly higher than our predictions. Nevertheless, this deviation needs to be put into perspective as the rheology of the margin is not constrained and several parameters such as the coupling between the crust and the mantle can increase the limit of 50 Myr.

Additionally, our results suggest that suitable conditions for SI initiation at passive margins occur for oceanic lithosphere with an age range between 50 and 110 Myr. Subduction initiation at passive margins resembles a major problem in plate tectonics as no observations can be tied unequivocally to this process. However, several subduction zones are believed to have been formed at continental margins such as subduction systems along the west Pacific margins and in eastern Indonesia (Hall, 2019, Karig, 1982), or in the European Alps (Manzotti et al., 2014a, Marroni et al., 2017). SI in these examples is characterized by nappe stacking of continental crust followed by formation of a sedimentary wedge at the subduction interface. In the case of the Alps, the age of the oceanic lithosphere during the inception of subduction was estimated between 60 to 80 Myr (Manzotti et al., 2014a, Marroni et al., 2017). It suggests that the

age of the oceanic lithosphere is within in the range of what is inferred as “favourable” for SI in this study.

If the margin is too old (age > 110 Myr) the strength of the mantle lithosphere is too high to allow for underthrusting and shortening is accommodated by widespread deformations in the continental lithosphere, as observed for example along the North Atlantic margin (Holford et al., 2009), the Brazilian margin (Cogné et al., 2012; Marques et al., 2014; Riccomini et al., 2004) or the Iberian margin (Duarte et al., 2013). It is however unlikely that these passive margins will ever become an active plate boundary. For such old passive margins, the migration of an existing plate boundary is probably the easiest way to initiate subduction (Duarte et al., 2013).

4.6 Conclusions

Analogue and numerical modelling were carried out in order to isolate the key mechanisms responsible for subduction initiation at passive margins. A selection of experiments, involving both oceanic and continental lithospheres subject to shortening, is used to investigate the role of oceanic and continental lithosphere rheology for scenarios where convergence is orthogonal to the passive margin. The modelling results show that subduction initiation takes place in three stages including: I) early-stage deformation of the continental passive margin crust II) underthrusting of the oceanic plate and III) sinking of the oceanic lithosphere into the asthenospheric mantle. Each stage is then controlled by specific parameter combinations and deformation mechanisms that can be summarized as follows:

The location of subduction initiation depends on the strength contrast between the oceanic lithosphere and the lithosphere at the margin. It is in strong correlation with the thermal age of the oceanic lithosphere. For an age < 50 Myr, subduction initiates in the oceanic domain, whereas an age > 50 Myr produces at the ocean-continent boundary. However, for an age > 110 Myr, ductile shear resistance of the mantle lithosphere prevents underthrusting and shortening is accommodated by the deformation of the continental lithosphere. Additionally, deformation localizes at passive margins where the ductile continental crust wedges out. There, thrust faults develop above this decoupling horizon, which acts as a decollement and promotes the localization of deformation into a major thrust that accommodates most of the shortening. In contrast, strong crust-mantle coupling at passive margins leads to a transfer of deformation toward the oceanic domain.

Propagation of deformation into the mantle lithosphere allowing for its underthrusting is facilitated by processes that reduce the strength of the mantle such as Peierls creep. Simultaneously and subsequently further localization of deformation and the development of shear zones, which eventually evolve into a subduction zone, requires shear-heating to be effective. The combination of the two mechanisms is necessary to localize strain in the ductile mantle lithosphere and thus, accommodate the sinking of the subducting slab.

Appendix B

Additional numerical modelling results for oceanic lithospheres with an age of 30 Myr (a, b, c) and 140 Myr (e, f, g) at the onset of shortening, respectively. Colour coding is the same as in Figure 4.1. White lines correspond to isotherms (200°C spacing). Model time corresponds to 30 Myr of shortening. (a, d) Models including shear heating, olivine Peierls creep and a strong continental crust leading to crust-mantle coupling at the margin. (b, e) Model with shear heating but no Peierls creep in olivine. (c, g) Model including olivine Peierls creep but no shear heating. For the models shown in figure b and c, continental crust is the same as for the reference model.

Chapter 5

This chapter is based on the following manuscript: "Auzemery A., Yamato P., Duretz T., Willingshofer E., Matenco L., Porkolab K. Influence of magma-poor versus magma-rich passive margins on subduction initiation. Submitted for publication in Gondwana research"

Chapter 5

Influence of magma-poor versus magma-rich passive margins on subduction initiation

Highlights

- Numerical models are used to evaluate the sensitivity of magma-poor and magma rich passive margins to subduction initiation.
- Crust-mantle decoupling at magma-poor margins facilitates subduction initiation at passive margins
- Magma-rich margins are too strong after cooling and subduction initiates within the oceanic lithosphere.

5.1 Introduction

Recent studies have shown that the mechanics of subduction initiation remain less understood (e.g. Hall, 2019; Stern & Gerya, 2018) and no “one fits all” model can be postulated. These studies emphasize that pre-existing tectonic plates configurations and density contrasts across mechanically weak zones are key ingredients for subduction initiation (e.g. Arcay et al., 2020; Nikolaeva et al., 2010). These conditions can be met at oceanic transform and major fracture zones or in mantle plume settings, where strong lateral density and strength contrasts exist (e.g. Cloetingh et al., 2021; Whattam & Stern, 2015). Ocean-continent transitions also meet such conditions, but direct evidence for subduction initiation at passive margin is scarce (Crameri et al., 2020; Gurnis et al., 2004), promoting the idea that externally driven processes such as subduction zone invasion are needed for the genesis of subduction zones at continental margins (Crameri et al., 2020; Duarte et al., 2013; Zhou et al., 2020). Although this mechanism may explain some natural examples (e.g. western edge of the Pacific plate, Hall, 2019; South Sandwich, van de Lagemaat et al., 2021), its spontaneous retreating mode of subduction (Zhou et al., 2020) contrasts with the geological record of subduction initiation at magma-poor continental margins such as in the Pyrenees or the European Alps orogenic systems (McCarthy et al., 2018), suggesting that subduction initiation must have occurred at passive margins in the plate tectonics history.

Subduction initiation at passive margin requires driving forces to exceed the strength of the ocean-continent transition (Baes & Sobolev, 2017; Cloetingh et al., 1989; Faccenna et al., 1999; Nikolaeva et al., 2010), which is subject to transient changes during thermal relaxation (e.g. Zhong & Li, 2019). Together with the rheological stratification of the passive margin lithosphere (Faccenna et al., 1999; Goren et al., 2008), these parameters are inferred to be critical for the locus of subduction initiation (Auzemery et al., 2020; Nikolaeva et al., 2011). These modelling studies have shown that, under compressive stress regime, deformation is focussed in the ductile lower continental crust when the oceanic lithosphere at the margin is older than ~50 Myr and the continental lithosphere is relatively thin (~100-150 km). When these conditions are

met, subduction initiation is physically feasible by the evolution of major shear zones in the ductile crust that propagates into the mantle where thermal softening facilitates the formation of a subduction zone (Auzemery et al., 2020; Kiss et al., 2020). In contrast, deformation localises within the rheologically weaker oceanic lithosphere, when they are younger than 50 Myr and/or in case of high convergence rate ($> 2 \text{ cm yr}^{-1}$), creating intra-oceanic subduction (Auzemery et al., 2020; Faccenna et al., 1999; Zhong & Li, 2019).

All previous modelling studies have not yet quantified the impact of significant differences in geometry and rheology between magma-rich and magma-poor continental passive margins (MRPM and MPPR, respectively) in subduction initiation (Fig. 5.1), constrained by the available observational record (Geoffroy, 2005; Tugend et al., 2020; Sengör & Burke, 1978; White et al., 1987). MRPM are typically driven by mantle upwelling and are associated with significant magmatic activity during rifting (Fig. 5.1a, Corti, 2009; White et al., 1987). In particular, important is the presence of magmatic intrusions and extrusions as well as underplating of mafic bodies at the base of the crust, which significantly increases its effective strength after cooling (Kelemen & Holbrook, 1995). In contrast, MPPM are typically driven by lithospheric stretching without significant decompressional melting magmatism, resulting in exhumation and hydration of the mantle in the footwall of major detachments (Fig. 5.1b, Brun & Beslier, 1996; Clerc et al., 2018; Manatschal et al., 2001; Mohn et al., 2012). The hydration-induced serpentinization penetrates in the mantle lithosphere and creates rheologically weak decollement layers that are prone to tectonic reactivation (Guillot et al., 2015; Hilairet et al., 2007).

Figure 5.1: Cartoons showing first order characteristics of a) magma-rich (inspired from Geoffroy, 2005); and b) magma-poor continental rifted margins (inspired from Mohn et al., 2012) for two evolutionary stages.

In this study, we aim to understand and quantify subduction initiation in the vicinity of contrasting end-member types of passive continental margins (MRPM and MPPR) during far-field induced shortening by means of thermo-mechanical numerical modelling, constrained by the available observational record. In particular, we test the influence of differences in crustal and lithospheric structure, composition and evolution of deformation. A two-steps strategy is adopted to understand both first order controlling mechanisms and the more detailed response. In a first step, we assess the deformational response to the degree of rheological crust-mantle coupling at continental passive margins. In a second step, we use more realistic geometries and compositions of magma-poor and magma-rich continental passive margins to test their specific impact in subduction initiation. The modelling results are validated against observations and quantitative reconstructions of the Adriatic micro-continent, an ideal well-studied area that contains the record of both end-member types of passive continental margins (MRPM and MPPM). This record is exposed in two European orogenic systems, the Dinarides-Hellenides and the Alps, for the MRPM and the MPPM, respectively.

5.2 Modelling approach

5.2.1 Methods

We performed two-dimensional (2D) thermo-mechanical modelling to simulate the shortening of a visco-elastic-plastic lithosphere (Fig. 5.2, Tab. 5.1 and 5.2). The finite-difference, marker-in-cell code (MDoodz; Duretz et al., 2016b) solves the equations of momentum and mass conservation as well as heat transfer, as detailed in chapter 2. The model domain is a section of 3000×500 km and the grid resolution is 1×1 km in both dimensions (Fig. 5.2). A constant normal velocity is applied at the left and right boundary of the model ($|V_{in}| = 0.625$ cm/yr) from the surface to a depth of 200 km. The total convergence rate of 1.25 cm/yr reflects an average rate of slowly converging systems, in accordance with estimates specific to the closure of the Vardar and Piemonte Liguria Ocean (e.g., Capitanio & Goes, 2006; Handy et al., 2010; Le Breton et al., 2020). In order to satisfy mass conservation, an outflow velocity (V_{out}) is distributed at the base of the model and on the sides from 200 km depth to the base of the box (Fig. 5.2). The outflow is proportionally distributed over the length of each boundary.

Deformation is governed by Drucker-Prager plasticity, dislocation creep, diffusion creep and Peierls creep and elasticity (see parameters in tables 5.1 and 5.2 and chapter 2 for relevant mathematical formulations). The presented models do not account for any predefined strain softening parameterization or weak zone for subduction initiation. The model top boundary is a true free surface (Duretz et al., 2016b). Because the effects of sediments have been inferred to be critical for the development of a self-sustaining subduction zone (Cloetingh et al., 1984; Nikolaeva et al., 2010; Sobolev & Brown, 2019), erosion and sedimentation are parameterized in the equation, following a kinematic approach (e.g. Candiotti et al., 2020) where erosion and sedimentation are implemented above or below a base level fixed at 0 km. To limit the loading effect,

which is not the focus of our study, we adopted a low sedimentation/erosion rate, setting diffusivity D to $10^{-6} \text{ m}^2.\text{s}^{-1}$ (for $h > 0$), and sedimentation to 0.5 mm.yr^{-1} (for $h < 0$). The accommodation space is filled in all models with a sedimentary material composed of alternating calcite and mica layers, which is a rheologically weak composition.

5.2.2 Modelling strategy

We have chosen a two-steps modelling strategy: (1) we initially run a series of reference models consisting of a simplified rheological stratification where the entire crust is represented by the same material, and (2), we perform another set of models that assumed a layered and laterally variable crustal composition that matches as close as possible the two end-member types of passive continental margins (MPPM & MRPM). The models (Fig. 5.2) consist of a continental lithosphere made of a 36 km thick crust and a lithospheric mantle with the thermal base (H_L) is fixed at a depth of 170 km reflecting a lithosphere of Palaeozoic thermo-tectonic age (Artemieva, 2009; Burov, 2011). The oceanic lithosphere consists of a 2 km sedimentary layer, a 8 km thick crust and a lithospheric mantle with a thermal base defined by the 1330°C isotherm, representing a ca. 70 Myr old lithosphere in accordance with plate cooling models (Parsons & McKenzie, 1978; Houseman et al., 1981; Turcotte & Schubert, 2014). Such configuration, together with a relatively slow convergence rate is supposed to favour subduction initiation at a passive margin (Auzemery et al., 2020; Kiss et al., 2020; Zhong & Li, 2019). The passive margin geometry includes a crust that wedges-out towards the ocean-continent transition. The base of the lithosphere (controlled by the 1330°C isotherm) at the passive margin is defined by a linear interpolation between the ocean and the continent. The initial geotherm is computed assuming steady state and accounts

Figure 5.2: Initial setup and results of the reference models: (a) Model setup. White and black arrows show the directions of outflow ($-V_{out}$) and the velocity applied at the left boundary of the models ($+V_{in}$), respectively. PM is the passive margin.

Table 5.1: Rheological and thermal parameters used for the reference numerical models. Here ρ is the density, k is the thermal conductivity, Q_r is the radiogenic heat production, ϕ is the friction angle, A is a pre-factor, V is the activation volume, f is a correction factor, n is the stress exponent, Q is the activation energy. References for rheology are R_95: Ranalli (1995); H&C_83: Hansen & Carter, 1983; H&K_03: Hirth and Kohlstedt (2003b); M_98: Mackwell et al., 1998; R&D: Rybacki & Dresen, 2004; K_90: Kronenberg et al. (1990); H_07: Hilairet et al. (2007).

	ρ (kg. m ⁻³)	k (W.m ⁻¹ .K -1)	Q_r (W.m ⁻³)	ϕ (°)	A (Pa ⁻ⁿ .s ⁻¹)	V (m ³ . mol ⁻¹)	f	n	Q (J.mol ⁻¹)	Ref.
Sediments										
sediment1 (calcite)	2600	2.5	2.0e-6	20	1.59e-25	0	0	4.7	297.0e3	K_90
sediment2 (mica)	2400	2.5	2.0e-6	20	1.00e-138	0	0	18.0	51.0e3	K_90
Continental crust										
Granite (westerly granite)	2800	2.7	9.0e-7	30	3.16e-26	0	0	2.3	154.0e3	H&C_83
Anorthite (dry anorthite)	2800	2.7	9.0e-7	30	3.98e-16	0	0	3.0	346.0e3	R&D_04
Granulite (mafic granulite)	2800	2.7	9.0e-7	30	8.83e-22	0	0	4.2	445.0e3	R_95
Diabase (strong diabase)	2800	2.7	9.0e-7	30	5.04e-28	0	0	4.7	485.0e3	M_98
Oceanic crust (wet olivine)	2900	3	1.0e-7	30	5.68e-27	11.0 e-6	1	3.5	480.0e3	H&K_03
Mantle lithosphere (dry olivine)										
Dislocation creep	3350	3.2	1.0e-10	30	1.10e-16	11.0e-6	0	3.5	530.0e3	H&K_03
Diffusion creep	3350	3.2	1.0e-10	30	1.50e-15	1.5e-16	0	1.0	375.0e3	H&K_03
Asthenosphere (dry olivine)										
Dislocation creep	3300	3.2	1.0e-10	30	1.10e-16	11.0e-6	0	3.5	530.0e3	H&K_03
Diffusion creep	3300	3.2	1.0e-10	30	1.50e-15	1.5e-16	0	1.0	375.0e3	H&K_03

Table 5.2: Rheological and thermal parameters used for the magma-poor and magma-rich passive margins models.

	ρ (kg.m ⁻³)	k (W.m ⁻¹ .K ⁻¹)	Qr (W. m ⁻³)	Φ (°)	A (Pa.s ⁻¹)	v (m ³ .mol ⁻¹)	f	n	Q (J.mol ⁻¹)	Ref.
Sediments										
sediment1 (calcite)	2600	2.5	2.0e-6	20	1.59e-25	0	0	4.7	297.0e3	K_90
sediment2 (mica)	2400	2.5	2.0e-6	20	1.00e-138	0	0	18.0	51.0e3	K_90
Continental crust MPPM										
Upper (dry quartzite)	2800	2.7	9.0e-7	30	3.98e-19	0	0	2.4	156.0e3	R_95
Lower (felsic granulite)	2800	2.7	9.0e-7	30	2.01e-21	0	0	3.1	243.0e3	R_95
Hydrated mantle (serpentinite)	2800	2.7	9.0e-7	30	8.83e-22	0	0	4.2	445.0e3	H_07
Continental crust MRPM										
Upper (dry quartz)	2800	2.7	9.0e-7	30	3.98e-19	0	0	2.4	156.0e3	R_95
Lower (felsic granulite)	2800	2.7	9.0e-7	30	2.01e-21	0	0	3.1	243.0e3	R_95
Extrusive (wet olivine)	2900	3.2	1.0e-10	30	1.10e-16	11.0e-6	0	3.5	530.0e3	H&K_03
Underplating (dry diabase)	3000	2.7	9.0e-7	30	5.04e-28	0	0	4.7	485.0e3	M_98
Oceanic crust (w. olivine)	2900	3	1.0e-7	30	5.68e-27	11.0 e-6	1	3.5	480.0e3	H&K_03
Mantle lithosphere (dry olivine)										
Dislocation creep	3350	3.2	1.0e-10	30	1.10e-16	11.0e-6	0	3.5	530.0e3	H&K_03
Diffusion creep	3350	3.2	1.0e-10	30	1.50e-15	1.5e-16	0	1.0	375.0e3	H&K_03
Asthenosphere (dry olivine)										
Dislocation creep	3300	3.2	1.0e-10	30	1.10e-16	11.0e-6	0	3.5	530.0e3	H&K_03
Diffusion creep	3300	3.2	1.0e-10	30	1.50e-15	1.5e-16	0	1.0	375.0e3	H&K_03

for radiogenic heat production in each layer and a constant asthenosphere temperature of 1330°C. The resulting temperature at the continental Moho is ~460°C. When displaying modelling results, we focus on the mechanically significant part of the lithosphere defined by the 1000°C geotherm (~base of the strength envelope Burov, 2011).

In this first step, we carried out 32 numerical models (Fig. 5.3d) and systematically varied the length and the crustal composition of the passive margin lithosphere, thereby reducing the mechanical coupling between the crust and the mantle lithosphere at the margin. The modelling results are used for constructing domain diagrams (Fig. 5.3d), which allow for mapping the mode of deformation against the length and the crust-mantle coupling of the margin. The reference models test 8 different lengths of passive margins changing from 50 km to 400 km, covering most of natural passive margin lengths (Reston & Manatschal, 2011), as well as 4 different rheologies for the crust (Table 5.1 and Fig. 5.3d). They thus provide an overview of suitable parameter combinations for subduction initiation and provide a geometric and rheological frame for models with specificities tailored to represent magma poor and magma rich passive margins.

The second set of models include the specific crustal characteristics of MPPM and MRPM as illustrated in figure 5.1, with the same thermal base in the oceanic and continental domain for both models. A different flow law was attributed to each material, in agreement with the simulated type of oceanic or continental layered lithosphere, magmatic material and overlying sediments, as well as their lateral distribution, to account for a representative simplification of a more realistic natural scenario (Figs. 5.5a, 5.6a). To be able to highlight the relevance of crust-mantle coupling and to correlate models with our parametric study, the crust and mantle thicknesses were kept similar. However, a 2-layer model is now used for the continental crust consisting of an upper crust of 24 km thickness with dry quartzite rheology and a lower crust between 24 and 36 km depth where deformation is described by a felsic granulite flow law (see parameters in table 5.2).

Figure 5.3: Influence of crust-mantle coupling and length of the lithosphere at passive margin on subduction initiation. Right: modes of subduction initiation (a) subduction at the passive margin, (b) intra-continental subduction and (c) intra-oceanic subduction. Enlargements display the strain rate and the viscosity at the early stage of underthrusting as well as the direction of the maximum principal stress (σ_1). wgr., an., gr., dol., stand for westerly granite, anorthite, granulite, dolerite, respectively. (d, e) Parametric study: (d) mode of subduction initiation, (e) time period of subduction initiation. →

5.2.3 Model limitation

The aim of this study is to provide insights into favourable rheological conditions for subduction initiation at magma-rich and magma-poor passive continental margin where convergence is orthogonal to the ocean-continent transition. Inherent to the approach are model limitations related to the choice of initial and boundary conditions and the dynamics of the system. Our models do not account for initial 3D variations of crust and lithosphere architecture, structure, and composition, inherited from rifting and passive margin formation. These have recently been implemented and discussed in the modelling studies of Beaussier et al. (2019), Gülcher et al. (2019a), or Zhou et al. (2020) and are particularly important when convergence is oblique to the inherited structures (Beaussier et al., 2019) and when subduction initiation takes place by lateral propagation of an existing subduction zone (e.g. Zhou et al. 2020). The forces required for subduction initiation are typically lower in 3D (Ghosh et al., 2006), our 2D modelling results represent upper boundary conditions to the forcing of the system. In our study subduction initiation is driven by horizontal forcing imposed by convergent kinematic boundary conditions. Therefore, we do not account for mantle upwelling (Gerya et al., 2015; Koptev et al., 2019; 2021) or mantle suction flow (Baes & Sobolev, 2017; Yamato et al., 2013). Surface processes on the other hand are implemented in simplified form in this study but we do not account for climate driven variability of sediment supply to the trench controlling the lubrication of the evolving subduction zone (Sobolev & Brown, 2019). Although many of these parameters are likely important for subduction initiation at magma-poor and magma rich margins, our 2d models provide a significantly advanced approach by incorporating inherited features such as rheologically weak layers (serpentinised mantle lithosphere). Moreover, our results have the required resolution to accurately describe the system behaviour and obtain novel insights in the key parameters controlling the mechanism associated with SI.

5.3 Reference model results

Using numerical experiments, we have systematically investigated the role of the passive margin length and the degree of crust-mantle coupling on subduction initiation (Fig. 5.3d). For each rheological configuration presented above, the length of the margin (the wedge in the models) has been varied yielding three end-modes of subduction initiation: subduction at the passive margin, intra-continental subduction and intra-oceanic subduction (Figs. 5.3a, b, c, respectively). More specifically, intra-continental subduction refers to the subduction of the continental lithospheric mantle under the oceanic lithosphere.

5.3.1 Role of crust-mantle coupling

For a constant length of the passive margin (e.g. 150 km, Fig. 5.4), variation of continental crust rheology impacts on the degree of crust-mantle coupling and consequently controls the locus of subduction initiation. For instance, the presence of a granitic continental crust allows the decoupling of the crust from the mantle and favours

subduction initiation at the margin. In this case, deformation starts at the base of the ductile crust at the ocean-continent transition, where a shear zone develops, which eventually propagates into the mantle lithosphere (Fig. 5.4a). In contrast, shortening of a strong doleritic or granulitic continental crust, which results in strong crust-mantle coupling, consistently leads to folding of the lithosphere followed by intra-oceanic subduction (Fig. 5.4c-d). In this case, early-stage deformation affects the passive margin, which bends downward (Fig. 5.4c).

Models where continental crust is simulated using an anorthite flow law, (intermediate case of coupling) follow a similar evolution as those with a granitic crust. However, subduction at the margin is intricate and folding is observed in the oceanic domain (Figs. 5.4a, b). We note that folding is better expressed close to the passive margin than in the middle of the oceanic plate (Fig. 5.4c), likely reflecting stress concentration at the ocean-continent transition. Evidence of this mechanism is highlighted by a decrease of the amplitude and the wavelength of folding from the margin to the ocean (Figs. 5.4b, c).

In case of stronger coupling (dolerite crust), deformation at the margin is limited and folding mainly affects the oceanic domain (Fig. 5.4d). However, strain does not localize over a long period of time and deformation shifts from the passive margin towards the oceanic domain where subduction occurs.

Figure 5.4: Left: Role of crust-mantle coupling on the locus of subduction initiation. Crust-mantle coupling is controlled by the composition of the continental crust. Right: Strength profiles where σ_{II} is the second invariant of the deviatoric stress. Length of passive margin is 150 km for all experiments. The different colours in the crust represent the four different compositions of the crust used in this study. The rest of the layering is similar to Fig. 5.2.

5.3.2 Role of the passive margin length

Our results show that subduction initiation at passive margins requires the development of a shear zone that subsequently propagates into the mantle lithosphere (Fig. 5.3d). Deformation localizes at the base of the ductile crust of the continental margin, where a shear zone initiates. Varying the passive margin length implies changing the wedge or taper angle (α) of the margin, and thus, the angle at which the ductile lower crust dips towards the continent (Figs. 5.3a-c). Our results highlight that the likelihood of subduction initiation at passive margins strongly decreases with increasing margin length. For margin lengths in excess of 300 km, equivalent with wedge angles of $< 7^\circ$, intra-continental subduction is observed (Fig. 5.3b). In contrast, short margin lengths result in initially steeper dip, which favours flow of the ductile crust at the ocean-continent transition and fosters subduction initiation.

A small wedge angle represents a case of a long passive margin where the ductile crust gets gradually thinner towards the ocean-continent transition. When the thickness of the ductile crust gets lower than ca. 3,5 km, stresses within this layer get too high and the ductile crust does not act as a decollement anymore. Instead, deformation is transmitted further into the ductile crustal layer toward the continent (Fig. 5.3b). In this situation, shortening is accommodated by folding of the mantle lithosphere at the margin, resulting in deflection of the continental Moho. In the later stage of the model, strain propagates into the mantle leading to intra-continental subduction (Fig. 5.3b). The ocean-ward polarity of the subduction zone is probably controlled by the buttressing effect of the long and overall strong passive margin crust, which is coupled to the mantle lithosphere in the area of the ocean-continent transition. Therefore, our models suggest that the formation of a subduction zone at passive margins is more easily achieved with a steep and decoupled passive margin (Fig. 5.3a). In other cases, deformation is transmitted toward the continental crust (Fig. 5.3b) or the oceanic lithosphere (Fig. 5.3c).

5.3.3 Timing of subduction initiation

Our results highlight an important correlation between geometrical (length of passive margin and margin angle) and rheological (thickness of ductile continental crust) initial conditions with the period of time required for subduction initiation (Fig. 5.3e). We refer to duration of subduction initiation in our models as the period between the start of shortening (at $t=0$) and the moment when the continuous shear zone analogous to an incipient subduction plate boundary is established. Furthermore, we use the thickness of the ductile crust as a proxy for the degree of crust-mantle coupling, where a thicker ductile crust represents a decoupled case and vice versa. For a similar convergence rate, initiation of subduction requires less time for a decoupled lithosphere (<10 Myr) than for a coupled lithosphere (>15 Myr, see Figs. 5.4b-c), because a weak lower crust favours strain localization early in the evolution of a subduction zone. For the same reason, the length of the margin increases the duration of subduction initiation. An exception are cases with strong crust-mantle coupling at the passive continental margin (crust simulated using dolerite). In such cases subduction initiation does not occur at the passive margin, which deforms over a long period of time, but within the

oceanic plate (Fig. 5.3e). We emphasise that the duration of subduction initiation also includes early phases of distributed deformation of the passive margin prior to strain localization. This phase is longer by ca. 10 Ma for weak margins (e.g. anorthite crust) as opposed to strong margins (e.g. dolerite) as indicated in figure 5.3e.

5.4 Application to magma-poor and magma-rich passive margins

The parametric study allowed us to infer the sensitivity of parameters controlling localization of deformation and subsequent subduction initiation. The next step is to understand the impact of a more realistic magma-poor (Fig. 5.5) and magma-rich (Fig. 5.5) passive margin geometries and composition for subduction initiation.

5.4.1 Subduction initiation at magma-poor passive margins

The magma-poor passive margin model (Fig. 5.5a) accounts for the necking of continental crust, resulting in a large omission of the lower crust in distal margin domains and the exhumation and hydration of sub-crustal mantle peridotites at the ocean-continent transition (Brun & Beslier, 1996; Manatschal et al., 2001; Mohn et al., 2012). As in geophysical models, the exhumed mantle shows variable degrees of serpentinization (Bayrakci et al., 2016), which is accounted for in the model by placing distributed patches of serpentinite material within the uppermost mantle lithosphere, where the mantle is located at the sea floor and below the extended margin (Fig. 5.5a). The latter represents serpentinization due to normal faulting cutting across the extended margin crust. By implementing these complexities, we account for a relatively weak lithosphere at the ocean-continent transition that represents a thinned continental lithosphere overlying a serpentinized mantle, and therefore, a relatively weak crust-mantle coupling (Fig. 5.5a).

The evolution of our simulation follows the general trend of a model with crust/mantle decoupling at the base of the margin (Fig. 5.5b) and includes: the onset of deformation at the ocean-continent transition, the formation of shortening structures across the margin, subsequent propagation of deformation into the mantle lithosphere and finally underthrusting of oceanic lithosphere at the passive margin. The results show that the thrust-type deformation sequence is younging towards the continent (fault 1 is older than fault 2, Fig. 5.5b) during the early stage of deformation. Different to the reference models, strain rates are highest ($\sim 10^{-13} \text{ s}^{-1}$) within the serpentinized mantle, (Fig. 5.5b1) and not within the ductile crust. The model shows that at ca. 10 Ma of shortening, shear along the weak serpentinites links up with shear within the ductile crust below the necking zone to form a throughgoing shear structure at the base of crust (Fig. 5.5b2). Furthermore, this shear structure is linked to a major thrust-type structure (n° 2 in Fig. 5.5b2) that evolves near the necking zone where the margin is getting thicker, and the wedge angle increases. Subsequently, the distal margin is thrust below the continental lithosphere and accreted to the overriding plate at depth of 25-35 km (Fig. 5.5b3). At the position of the former distal margin a thick sedimentary wedge developed ~ 15 Myr after the onset of convergence.

Figure 5.5: Subduction initiation for a magma-poor passive margin. Length of passive margin is 150 km. (a) Model setup with an enlargement of the passive margin to show the geometry and the rheological stratification of a magma-poor margin. Strength profile indicates crust-mantle decoupling. (b) Model evolution, illustrating key stages in the subduction initiation process where the left panel shows the deformation of the layers and associated temperature structure. Red numbers indicate the sequence of interpreted shear structures. The right panel shows the distribution of strain rate and viscosity for each time-step. White arrows represent the velocity field. PM is the passive margin.

5.4.2 Subduction initiation at magma-rich passive margin

The model involving a magma-rich passive margin accounts for magmatic underplating, as well as mafic intrusions and extrusions, in accordance with the dynamics of an active and melting mantle as a result of partial melting of mantle lithosphere (Geoffroy, 2005). Magmatic underplates are often characterized by high-velocity seismic zones in the lower crust (Kelemen & Holbrook, 1995). Mafic intrusions are gabbroic complexes (sills, dikes) distributed across the MRPM. Their equivalents at the surface correspond to large basaltic lava flows identified as seaward dipping reflectors on seismic reflection data (White et al., 1987). We ascribe a dolerite rheology to the magmatic underplatings while magmatic extrusions are simulated by wet olivine. In the model we implement these features by accounting for a margin that exhibits a continental crust intruded by mafic rocks. The evolution of our model during subduction initiation of a MRPM follows the general trend of a model with a strong crust/mantle coupling at the margin (Fig. 5.6b). It features first distributed deformation at the ocean-continent transition, migration of deformation and folding of the oceanic lithosphere, followed by intra-oceanic subduction.

The modelling shows that the addition of magmatic lithologies strengthens the passive margin close to the ocean-continent transition and increases the crust-mantle coupling, preventing strain localization. Therefore, most of the strain is localized in the oceanic lithosphere near the ocean-continent transition, while less deformation is observed in the ductile layer of the proximal continental margin (Fig. 5.6b1). The oceanward underthrusting takes place 350 km away from the margin, at one of the conjugate shear structures (Fig. 5.6b2). Although the oceanic domain is 1500 km long, strain localizes relatively close to the ocean-continent transition, similarly to the reference model (Fig. 5.4c). The subsequent development of a subduction zone is associated with the development of an accretionary wedge, which is recognisable ~12 Myr after the onset of convergence (Fig. 5.6b3).

5.5 Discussion

5.5.1 Comparison to previous models

Although subduction zones often initiate through the lateral propagation from pre-existing subduction or polarity reversals (Crameri et al., 2020), other mechanisms that allow for the formation of new subduction plate boundaries must exist (Cloetingh et al., 2021; Crameri et al., 2020; Hall, 2019; McCarthy et al., 2018), most likely driven by far-field tectonic forcing (Gurnis et al., 2004). The probability of subduction initiation increases in the vicinity of continental margins (e.g. Ulvrova et al., 2019). Previous modelling studies have shown that the ability of a passive margin to convert into a subduction zone is largely governed by the age and compositional differences in the mantle lithosphere across the ocean-continent transition (e.g., Auzemery et al., 2020; Baes & Sobolev, 2017; Faccenna et al., 1999; Gurnis et al., 2004; Nikolaeva et al., 2011; Zhong & Li, 2019). Our models are in agreement with these general inferences, but further demonstrate that the degree of crust-mantle coupling has also a critical impact on

Figure 5.6: Subduction initiation for a magma-rich passive margin with a margin length of passive margin is 150 km. (a) Model setup with an enlargement on the passive margin to highlight the geometry and rheological stratification of a magma-rich margin. (b) Model evolution illustrating key stages in the subduction initiation process where the left panel shows the deformation of the layers and associated temperature structure. The right panel shows the distribution of strain rate and viscosity for each time-step. White arrows represent the velocity field. PM is passive margin.

the localisation of subduction zones (Fig. 5.4), which is further controlled by the geometry and the rheological layering of the continental crust at passive margins (Figs. 5.5 and 5.6). The latter controls where ductile flow dominates over plastic failure facilitating strain localization during the early phases of shortening. In our models this feature is inherent to the rheological stratification of the experiments, which is different to strain localization triggered by a thermal perturbation (Crameri & Kaus, 2010; Jaquet & Schmalholz, 2018; Thielmann & Kaus, 2012), although the deformational characteristics including the development of a major shear zone are similar in both cases. The degree of coupling at the passive margin can be governed by serpentinization of the lithospheric mantle (Fig. 5.5) or within the ductile continental crust. The linkage of a serpentinized shear zone with the ductile crust below proximal parts of the passive margin is key for the development of a large-scale shear zone that propagates into the mantle lithosphere to form the subduction plate boundary. Furthermore, we argue that serpentinized mantle lithosphere at MPPM is critical for initiating deformation at low stress levels. Kiss et al (2020) obtained low stress levels for Moho temperatures between 525 and 600 °C at the onset of contraction. Our results further show that by explicit implementation of flow laws simulating weak serpentinite rheologies, subduction initiation is mechanically feasible for Moho temperatures as low as ~ 460 °C. As such our results are in line with the study of Candiotti et al., (2020) who emphasise the importance of mechanical and geometric heterogeneities at the onset of shortening for reducing the force needed for subduction initiation.

It has been shown that the favourable age of oceanic lithosphere to develop a subduction zone at the margin is in the range of 50-110 Myr, (e.g. Auzemery et al., 2020). Within this age window, density and strength contrasts across the margin coupled to strain localization mechanisms provide suitable mechanical conditions for subduction initiation. However, a strong crust-mantle coupling of the continental lithosphere would prevent strain localization at the ocean-continent transition and generate a shift of deformation toward the oceanic domain (Fig. 5.6). Therefore, we agree that intra-oceanic subduction is more favourable in young oceanic basins (Auzemery et al., 2020; Faccenna et al., 1999; Zhong & Li, 2019), but it can also happen in old oceanic lithospheres for cases where the strength of the passive margin lithosphere is higher (e.g. through magmatism) compared to the oceanic lithosphere.

Together with the thickness of the ductile lower crust, the margin length is an important factor that influences strain localization through the crust and the mantle lithosphere. Therefore, for long (more than 300 km), decoupled and low angle passive margin, deformation is transmitted toward the continental crust (Fig. 5.3b) or the oceanic lithosphere (Fig. 5.3c). In the latter scenario, the subduction is expected to occur at a strength minima which coincides with the presence of serpentinite (Maffione & van Hinsbergen, 2018). Even if our concept of continental underthrusting is rarely mentioned in studies discussing plate tectonics, similar models have been obtained through reconstructions of the north African margin (Roure et al., 2012) Cantabrian margin (Teixell et al., 2018; Viejó & Gallastegui, 2005b) or in numerical studies (Kiss et al., 2020; Hamai et al., 2018).

5.5.2. Application to observations and reconstructions of subduction initiation near the former Adriatic passive continental margins

The evolution of continental passive margins along the former Adriatic micro-continent provides a unique opportunity to study subduction initiation in the vicinity of two continental rifted margins end-members. A MPPM and a MRPM developed over a similar pre-rifting continental structure and composition, with a similar oceanic lithospheric age at the time of subduction initiation and with comparable convergence rates. The Mediterranean region is tectonically complex and various views on the tectonic evolution are available (Le Breton et al., 2020; Capitanio & Goes, 2006; Csontos & Vörös, 2004; Handy et al., 2010; van Hinsbergen et al., 2020; Schmid et al., 2020; Stampfli & Hochard, 2009). In our reconstruction (Fig. 5.7), we consider the models of van Hinsbergen et al. (2020) and Schmid et al. (2020) which propose up-to-date quantitative definitions of the tectonic units.

5.5.2.1 Subduction initiation along the former Adriatic MPPM presently exposed in the Alps

The onset of convergence in the Neotethys Ocean was roughly coeval with the onset of Middle Jurassic opening of the Piedmont-Liguria branch of the Alpine Tethys Ocean at ~170-160 Ma (Fig. 5.7a-c, Liati et al., 2005; Schaltegger et al., 2015). Rifting magmatism is minor or non-existent in the Alpine chain, with the exception of the eastern Southern Alps transitional area, where a Triassic phase of extension is associated with magmatism (e.g. Lustrino et al., 2019). A MPPM has been defined across several Alpine transects (Manatschal & Bernoulli, 1999; Manatschal & Müntener, 2009; Mohn et al., 2012; Müntener et al., 2004). After 60-70 Myr (McCarthy et al., 2020), the closure of the Piemont-Liguria branch of the Alpine Tethys ocean was initiated by subduction at the Adriatic passive margin (Fig. 5.8a, Manzotti et al., 2014a; Marroni et al., 2017) and was driven by the northward motion of the Adriatic plate at a convergence rate of 0.9 to 1.5 cm/yr (Handy et al., 2010; van Hinsbergen et al., 2020). The Alpine Tethys Ocean was closed by the Eocene onset of continental collision, followed by the subsequent indentation of the Adriatic micro-continent in the Alps (e.g., Schmid et al., 2004).

Our numerical modelling of MPPM (Fig. 5.5) displays striking similarities with reconstruction proposed for the Sesia-Dent Blanche transect (Fig. 5.8a). The Adriatic passive continental margin preserved a necking domain that recorded the formation of detachments in the upper-middle crust (Mohn et al., 2012; Manzotti et al., 2014a). These detachments accommodated the extreme thinning of the Adria margin and created the large tectonic omission observed between the lower continental crust and exhumation of subcontinental mantle (Manatschal & Müntener, 2009; Müntener et al., 2004). The margin had a reconstructed width of 150-250 km at the time of subduction initiation (e.g. Beltrando et al., 2010), which was controlled by the presence of shear zones in the serpentinized exhumed mantle (Beltrando et al., 2010; Malatesta et al., 2011). The subsequent nappe emplacement was associated with high pressure-low temperature (HP-

Figure 5.7: Paleogeographic reconstructions of the western Neotethys domain, (compiled and adapted from Capitanio & Goes, 2006, van Hinsbergen et al., 2020, see also Table 5.3 of the appendix C to this chapter) (a) Middle Jurassic (170-160 Ma) (b) Upper Cretaceous (90-65 Ma). Black arrows indicate the relative motion of Adria. (c) Main tectono-magmatic and metamorphic events for oceanic basins in the Western Alps and the Dinarides relevant for subduction-initiation. B-U, SI, O and C denote break-up, subduction initiation, obduction and collision, respectively. A-B, C-D are the location of the profiles shown in Fig. 5.8.

LT) metamorphism (Manzotti et al., 2014a and references therein) and the development of an accretionary wedge consisting of oceanic, continental, and subcontinental mantle material (Marroni et al., 2017). Moreover, the record of high pressure rocks in the Sesia unit, reaching eclogite facies conditions (e.g., Regis et al., 2014) suggest subduction of part of the passive continental margin. More specifically, higher pressure conditions recorded in the former proximal part of the passive margin (Manzotti et al., 2014a) indicates that underthrusting started in the middle of the margin, probably close to the necking zone (Fig. 5.8a) at ~90-95 Ma. The basal units of the Sesia–Dent Blanche nappes were also affected by intense ductile shearing associated with pre-Alpine deformation (Beltrando et al., 2010). Observations from the Alps are in agreement with our modelling scenario where the initiation of subduction is driven by a basal thrust at the base of the continental crust along the passive margin, subsequently crosscut and duplicated in the

overall nappe system (Fig. 5.5). The original reactivation of the serpentized mantle during contraction enabled the propagation of deformation toward the necking zone. Therefore, our study agrees that the necking domain localises subduction initiation where the ductile lower crust is sufficiently thick to accommodate shortening (Fig. 5.5b), as previously inferred by Mohn et al., (2012). Our modelling also infers that a reactivation of the detachment at the base of the Sesia-Dent Blanche, Adriatic-derived extensional allochthon is not required to explain subduction initiation at the former Alpine passive margin (Manzotti et al., 2014a).

5.5.2.2 Subduction initiation along the former Adriatic MRPM presently exposed in the Dinarides-Hellenides

Separating from the European continent, the opening of the northern branch of the Neotethys Ocean (or Vardar) started at ~250-245 Ma (Fig. 5.7b-c, e.g., Schmid et al., 2020), when large volumes of Middle - Late Triassic intrusive and extrusive magmatism was documented along the Adriatic continental margin, from the Hellenides (e.g., Pe-Piper, 1998; Tsikouras et al., 2008) to the Dinarides (e.g., Pamic, 1984; Trubelja et al., 2004) and the eastern Southern Alps transitional area to the Alpine Tethys margin (e.g., De Min et al., 2020; Schmid et al., 2004a). The abundance of this basic to intermediate magmatism sourced from deep crustal or mantle levels (e.g., De Min et al., 2020; Pamic, 1984), associated with normal faulting and syn-kinematic deposition (e.g., Crisci et al., 1984; Robertson et al., 2009) infers partial melting during mantle upwelling and active rifting (e.g., Beltrán-Triviño et al., 2016). These observations combined with the Middle-Late Triassic age of calc-alkaline intermediate and mafic volcanism found in thrust ophiolitic mélanges (e.g. Pamic, 1984) testify the high magmatic budget along the Adriatic continental passive margin of the Dinarides-Hellenides (Fig. 5.8b). The subsequent closure of the northern Neotethys ocean by intra-oceanic subduction started during the late Early to Middle Jurassic (Fig. 5.7a, ~160-180 Ma), inferred from dating metamorphic soles and mid-oceanic ridge (MOR) ophiolites, from occurrences of Late Triassic to Early Jurassic slices of scraped cherts and MOR basalts in thrust ophiolitic melanges and from the Middle-Late Jurassic age of the obducted ophiolitic sheet (Bortolotti et al., 2013; Mikes et al., 2008; Maffione et al., 2015; Ustaszewski et al., 2009). Similar to the Adriatic continental rifted margin facing the Alpine Tethys, the age

Figure 5.8: Reconstruction of subduction initiation for the Alps and the Dinarides – Hellenides (adapted and simplified from Manzotti et al., 2014a; Schmid et al., 2004b, 2008,). (a) Shortening initiate subduction at the Alpine Tethys Adriatic MPPM. Development of shear zones in the serpentized exhumed mantle leads to nappe emplacement at the Adria continental margin. This evolution can be compared with our model of MPPM (Fig. 5.5b). (b) Development of intra-oceanic subduction in the vicinity of the Neotethys Adriatic MPPM. Locations AB, CD, of the cross-section are indicated in figures 5.7a and 5.7b. Black arrows indicate direction of extension and shortening respectively. CAB, MOR, SZZ stand for calc-alkaline basalt, mid-oceanic ridge and supra-subduction zone (magmatism). →

of the Neotethys ocean was ca. 50–70 Myr when subduction initiated (e.g. Schmid et al., 2004b, van Hinsbergen et al., 2020, Fig. 5.8). Plate reconstruction and paleo-magnetic studies suggest a convergence rate in the Neotethys Ocean of 1.5-2 cm/yr during 20 Myr (Capitanio & Goes, 2006; Handy et al., 2010; van Hinsbergen et al., 2020; Le Breton et al., 2020). This ocean was ultimately closed by the onset of the latest Cretaceous continental collision and the formation of the Sava suture zone between Adriatic- and European-derived continental units (e.g., Pamic, 1984; Schmid et al., 2020).

The comparison with our modelling study infers that the high magmatic budget of the Neotethys Adriatic continental rifted margin favoured the intra-oceanic subduction (Figs. 5.6, 5.7b). This mode of subduction initiation is driven by the strong rheological coupling of the passive margin, inhibiting the continent-ward propagation of deformation (Fig. 5.6). The subduction initiation took place during 10-15 Myr (Bortolotti et al., 2013), which is in agreement with our modelling results (Fig. 5.6). The age of oceanic relics found in ophiolitic melanges, combined with the age of obducted ophiolites, indicate that subduction initiated within the Early Jurassic part of the oceanic plate, although the occurrence of mid-ocean ridge ophiolites in contact with supra subduction zone ophiolites of the same age may also point to an subduction initiation close to the mid-oceanic ridge (Maffione et al., 2015). Although our models do not include a mid-oceanic ridge, they show that most shortening is initially localized in the place where the strength contrast is highest, i.e. at the ocean-continent transition, but migrates subsequently towards the oceanic domain, where the subduction zone forms at distances of less than 400 km from the margin (Fig. 5.4c).

Furthermore, our modelling predictions are comparable with previous observations in many other orogens containing ophiolitic belts, such as Anatolia (van Hinsbergen et al., 2016), Oman (Rioux et al., 2016; Saddiqi et al., 2006) or New-Caledonia (Lagabrielle et al., 2013). These observations have shown that the short time span (<10 Myr) between supra-subduction spreading centres and ophiolites emplacement points towards intra-oceanic subduction initiation close to the margin. These observations are also in agreement with our inferences showing that a coupled margin deforms significantly less at the ocean-continent transition and a subduction zone develops further into the ocean. We also acknowledge the potential role of plate re-organization and oblique convergence (Cramer et al., 2020; van Hinsbergen et al., 2020), or mantle rejuvenation by plume activity (Rolland et al., 2009), processes which are not accounted in our study.

5.6 Conclusions

Our thermo-mechanical models show that magma-poor passive margins with hyper-extended crust and exhumed serpentized mantle lithosphere are favourable sites for induced subduction initiation. The serpentized mantle facilitates strain localization and progressive development of a major shear zone that ultimately links up with zone of high strain that developed within the ductile lower continental crust in the necking zone. Shearing then propagates into the mantle lithosphere to initiate a subduction plate boundary. In contrast, passive margins intruded and underplated by magmatism favour

intra-oceanic subduction initiation. These margins are rheologically coupled and stronger, which induces distributed deformation during the early shortening, while the subsequent deformation migrates within the oceanic lithosphere. These results indicate that the rheological and geometric configuration of the passive margin, inherited from the magmatic budget during continental rifting phase, is critical for strain localisation mechanisms during the early stages of deformation and shows that subduction initiation is mechanically feasible for Moho temperatures below 500°C.

Our modelling results can be applied to the unique case of subduction initiation along two contrasting types of margins associated with the same Adriatic continental microplate. These results provide a mechanical explanation for subduction initiation at the ocean-continent transition in the western Alps and within the Neotethys ocean in the Dinarides-Hellenides orogenic systems, representing end-members of subduction initiation at magma-poor and magma-rich margins, respectively.

Appendix C

Table 5.3: references used for the construction of the timelines in Fig.5.7.

References	Area	Timing	Type	Details
Babić et al., 2002	Dinarides	Hettangien-late Bajocian	flysch(SOM)	Continental and marine clasts
Bortolotti et al., 2013	Dinarides	Anisian	sediments (SOM)	Pelagic syn rift carbonate
Dimitrijevic & Dimitrijevic, 1991	Dinarides	Triassic	sediments	Syn-rift volcano-clastic
Maffione et al., 2015	Dinarides	165-155 ma	magmatism	MORB ophiolites
Mikes et al., 2008	Dinarides	150-145 ma	sediments	Flysch margin
		Titonian-Barremian	sediments	Chert syn-obduction
Gawlick et al., 2009	Dinarides	150-130 ma	sediments	Micro-facies in melange
Robertson et al., 2009	Dinarides	Late Carnian	magmatism	Transitional-alkaline and MORB
Trubelja et al., 2004	Dinarides	Ladinian	magmatism	Calc-alkaline and MORB
		172-165 ma	magmatism	SSZ ophiolites
Bortolotti et al., 2005	Pelagonia	150-130	sediments	Flysch prism and MORB
Most, 2003	Pelagonia	150 ma	metamorphism	Exhumation HP rocks
Pe-Piper, 1998	Pelagonia	240-210 ma	magmatism	Alkaline and sub-alkaline

Photiades et al., 2003	Pelagonia	Mid-Triassic	magmatism (SOM)	Alkaline and MORB
Scherreiks et al., 2014	Pelagonia	160-140 Ma	sediments	Erosion continent
Saccani et al., 2011	Pelagonia	Mid-Triassic	magmatism (SOM)	Transitional-Alkaline and MORB
Tsikouras et al., 2008	Pelagonia	245 ma	magmatism	Subalkaline, alkaline
Šoštarić et al., 2014	Pelagonia	170-165 ma	metamorphism	4–6 kbar metamorphic soles
Beltrán-Triviño et al., 2016	southern Alps	240-235 ma	magmatism	MORB-CA
Cassinis et al., 2008	southern Alps	210-220 ma	magmatism	Sub-alkaline-tholeiitic
Crisci et al., 1984	southern Alps	225 ma	magmatism	SA with CA affinity
De Min et al., 2020	southern Alps	245-235 ma	magmatism	CAB contaminated
Thöni & Miller, 2000	eastern Alps	250-240 ma	magmatism	Alkaline
Visonà & Zanferrari, 2000	eastern Alps	245 ma	magmatism	Alkaline/Calc-alkaline
Babist et al., 2006	western alps	80	metamorphism	HP, blueschist, eclogite facies
Kapferer et al., 2012	western alps	32 ma	magmatism	cal-alkaline lavas
Hürlimann et al., 2016	western alps	35-30	magmatism	calc-alkaline magmatism
Manzotti et al., 2014a	western alps	90-65	metamorphism	SI blue to eclogite facies
Marroni et al., 2017	western alps	86-80	sediments	flysch sedimentary prism
Masini et al., 2013	western alps	180	sediments	sediments in supradetachment rift basins
	western alps	165	sediments	mantle exhumation
Meffan-Main et al., 2004	western alps	43-35 ma	metamorphism	HP metamorphism
Mohn et al., 2011	western alps	180-165	sediments	syn-rift clastic sediment
Monjoie et al., 2007	western alps	280-260	magmatism	transitional alkaline
Müntener et al., 2000	western alps	180-140 ma	metamorphism	exhumation granulite

	western alps	140-90 ma	metamorphism	stable LP-LT magmatism
Radulescu et al., 2009	western alps	35-37 ma	metamorphism	HP metamorphism
Regis et al., 2014	western alps	84-70	metamorphism	HP eclogite facies
Schaltegger et al., 2015	western alps	165-160	magmatism	tholeiitic MORB
Vavra et al., 1999	western alps	280-226	magmatism	thermal pulses in lower crust

Chapter 6

This chapter is based on the following manuscript: Auzemery A., Willingshofer E., Yamato P., Düretz T., Beckman F. (2021). Stress dependence of passive margin inversion: a key factor controlling subduction initiation. In preparation for the special issue on subduction initiation in *Frontiers in Earth Science*.

Chapter 6

Stress dependence of passive margin inversion: a key factor controlling subduction initiation

Highlights

- Numerical models are used to investigate the kinematic and stress conditions for subduction initiation to occur at passive margins or amid the oceanic plate.
- Slow convergence rate favours subduction initiation at passive margins over subduction at mid-oceanic ridges.
- Distribution of deformation is key to regulate levels of stress in the lithosphere.

6.1 Introduction

Studies of subduction systems show that oceanic subduction either nucleates at passive margins or within oceanic plates (Cramer et al., 2020; Gurnis et al., 2004; Stern, 2004; Stern & Gerya, 2018). Over the past decades several analogue, numerical and analytical modelling studies have been conducted to infer the preferred locus and suitable geometric, kinematic and mechanical conditions for the initiation of subduction zones (e.g. Auzemery et al., 2020; Candiotti et al., 2020; Cloetingh et al., 1989; Faccenna et al., 1999; Goren et al., 2008; Gurnis et al., 2004; Kiss et al., 2020; Maffione et al., 2015; Mart et al., 2005; Nikolaeva et al., 2010; Zhong & Li, 2019). These studies have shown that subduction initiation at passive continental margins critically depends on the buoyancy of the oceanic lithosphere as well as density and strength contrasts across the ocean-continent transition. Together with the rheological stratification of the passive margin lithosphere (Auzemery et al., 2020; Goren et al., 2008; Nikolaeva et al., 2010; chapters 3 and 5 of this thesis), these parameters control the locus of subduction initiation. Intra-oceanic subduction initiation is favoured for oceanic lithosphere younger than 50 Myr, whereas subduction nucleation occurs at the ocean-continent transition for cases of intermediate age (50-110 Myr) oceanic lithosphere and when the margin crust is decoupled from the underlying mantle lithosphere. The latter condition is particularly important because it facilitates strain localization and subsequent strain propagation within weak layers of the passive margin crust (Auzemery et al., 2020; Nikolaeva et al., 2010; Kiss et al., 2020 and chapters 4 and 5 of this thesis). Although the above quoted modelling studies successfully simulate the initiation of subduction zones at passive margins upon vertical or horizontal loading, the stress levels required for the nucleation of subduction are usually significantly higher than plate tectonic forces (Cloetingh et al., 1989; England & Wortel, 1980; Gerbault, 2000; Gurnis et al., 2004; Mueller & Phillips, 1991; Zhong & Li, 2019). In fact, the stress needed for subduction initiation at passive margins is in general one order of magnitude larger than the horizontal component of stress generated by ridge push (Mahatsente & Coblentz, 2015). From a mechanical

perspective, it is important to note that the critical stress for subduction initiation is also higher by ~ 5 TN than the lithospheric yield strength at (slow spreading) mid-ocean ridges (Luttrell & Sandwell, 2012). This suggests that upon contraction, deformation should predominantly affect the mid-ocean ridge, the weakest part of the system, where subduction would then initiate. Although active compressional tectonics is well-documented in several oceanic basins (e.g. Forsyth, 1973; Stein & Stein, 1993; Wyssession et al., 1991), recent and past examples of subduction initiation at or close to mid-ocean ridges as suggested for the Tethys or Pacific realms, are less well documented and subject to debate (Agard et al., 2016; Crameri et al., 2020), suggesting that mechanisms such as the dissipation of energy through shear heating act on the oceanic lithosphere to regulate and limit the stress level (Brun & Cobbold, 1980; Schmalholz et al., 2009). How such mechanisms then contribute to favour subduction initiation at passive margins over subduction at mid-oceanic ridges remains unclear.

As dissipation mechanism are linked to the strength of the lithosphere and thermo-mechanical feedback mechanism leading to strain localization, we argue that (1) convergence rate and the (2) thermo-tectonic age of oceanic and continental lithospheres are key parameters controlling the stress levels in the lithosphere. We test this hypothesis through thermo-mechanical modelling to infer the stress – strain-rate response to different kinematic and stress conditions eventually leading to subduction initiation at passive margins over subduction at mid-oceanic ridges. We conclude, by discussing how the interplay between far-field tectonic forcing and the strength of oceanic lithosphere impacts on the time-scales of subduction initiation and highlight similarities of modelling results to subduction initiation in the European Alps.

6.2 Stress magnitudes in oceanic basins

On earth, the magnitudes of depth-integrated compressive deviatoric stress is in the order of 3–10 TN/m (Cloetingh & Wortel, 1986; Coblenz & Richardson, 1996; Ghosh et al., 2013; Naliboff et al., 2009; Richardson et al., 1979). At continental passive margins sources of compressive stresses include: ridge push (e.g. England & Wortel, 1980), gravitational potential energy (GPE, e.g. Pascal & Cloetingh, 2009), tectonic forcing (Bird, 2017), or mantle convection (Ghosh et al., 2013; Kendall & Lithgow-Bertelloni, 2016) (Fig. 6.1). Ridge push arises from lithostatic pressure related to the elevation of the hot mid-ocean ridge above the cooler ocean basins surrounding it (e.g. Forsyth, 1973; Turcotte & Schubert, 2014). Although the contribution of each mechanism is unclear (Swedan, 2015), ridge push represents integrated deviatoric stress values between 1 and 5 TN/m (Mahatsente & Coblenz, 2015; Mueller & Phillips, 1991; Swedan, 2015), with an average value in the order of 3.5 TN/m for a 75 Myr oceanic lithosphere (Mahatsente, 2017). However, in some places, stress arising from the oceanic plate are of equal magnitude as the GPE from thick continental lithosphere (e.g. Tibetan plateau, 5–9 TN/m; Schmalholz et al., 2019), suggesting that in addition to ridge push, other forces must contribute to oceanic plate motion (Flesch et al., 2001; Ghosh et al., 2006; Naliboff et al., 2009). The horizontal shear tractions induced by mantle flow (Ghosh et al., 2013), could generate an additional integrated deviatoric stress in the order

of 1-3 TN/m (Kendall & Lithgow-Bertelloni, 2016). Consequently, a combination of ridge push and shear traction could generate stress levels in the order of 2 to 8 TN/m, with an average of ~ 5 TN/m for a 75 Myr oceanic lithosphere.

Higher loading levels can only be produced by transfer of stress through additional tectonic processes. As loads of 8-10 TN/m are needed to initiate subduction at passive margins (Zhong & Li, 2020) additional forces are required next to ridge push for the formation of a new subduction zone. It is, therefore, expected that stress levels in excess of ridge push would be relaxed through deformation of the mid-oceanic ridge, where the lithospheric strength is low. The strength of spreading ridges is subject to considerable uncertainty because the magma supply mechanisms are not sufficiently well understood and quantified (Luttrell & Sandwell, 2012). However, inferences from geochemical (Beaussier et al., 2019), geophysical (Luttrell & Sandwell, 2012) and numerical modelling (Husson, 2012) studies suggest that there is a strong correlation between spreading rate and thermal thickness of the lithosphere at mid-ocean ridges. When spreading rates are lower than ca. 1.5 cm yr^{-1} , the melt concentration is low, and the thermal thickness of oceanic lithosphere (isotherm 1300°C) at the mid-ocean ridges can be as large as 55 km (Husson, 2012). Therefore, an ultra-slow spreading ridge is of similar strength than a 5-8 Myr old oceanic lithosphere. This is in agreement with stress predictions associated with oceanic lithospheric folding suggesting that an ultra-slow mid-oceanic ridge could support horizontal loads of 8 TN/m (e.g. Indian ocean, Gerbault, 2000).

Figure 6.1: Simplified sketch illustrating magnitudes of integrated deviatoric stress observed on earth arising from: (a) ridge push (Mahatsente, 2017), (b) gravitational potential energy (GPE, Schmalholz et al., 2019) and (c) shear tractions generated by mantle flow (Kendall & Lithgow-Bertelloni, 2016). (d) In comparison, minimum deviatoric stress required for subduction initiation at a passive margin is 8-10 TN/m (Zhong & Li, 2020). The orange and dark blue layers represent the crust and the mantle lithosphere, respectively.

6.3 Numerical model

To investigate stress-controlled mechanisms for subduction initiation at passive margins, 2 D numerical thermo-mechanical models with a visco-elasto-plastic rheology were used. The finite-difference, marker-in-cell code (MDoodz; Duretz et al., 2016b)

was used to solve the equations of momentum and mass conservation, as well as the heat equation. For details regarding the mathematical model and algorithms, see chapter 2.

6.3.1 Modelling approach

In order to test our hypothesis, we perform numerical modelling simulating shortening of an oceanic basin and the adjacent continental margins. These models do not include melt production processes at mid-ocean ridges but are designed to capture the essential deformation features that characterize the shortening of an oceanic basin of a given age, as a consequence of the interplay between plate cooling and stresses arising from far-field forcing. Ultimately, they serve as a basic model to determine the stress and strain-rate boundary conditions required for subduction initiation at passive margins or within the oceanic domain (Fig. 6.2c).

The modelling approach of this study is two-fold (Fig. 6.2a, b). We first run a simplified model setup that adopts a laterally uniform physical properties of the oceanic lithosphere, where the age of the oceanic lithosphere is everywhere the same (Fig. 6.2a).

In step two, we incorporate the age dependant thickening of the lithosphere from an extinct spreading center to the passive margin (Fig. 6.2b). Hereafter, the extinct spreading center will be referred to as the “ridge”. The results of the parametric study (step 1) facilitate obtaining favourable boundary conditions for subduction initiation at passive margins in step 2, where the role of lithosphere thickness variation which varies through time with cooling was investigated.

6.3.2 Model geometry and rheology

The model domain is a section of 3000×500 km and the numerical resolution is 1×1 km in both dimensions. Models of step 1 comprise a 3000 km wide oceanic plate flanked by two continental plates on either side (Fig. 6.2a and Table 6.1). The continental lithosphere consists of an 18 km thick granitic upper crust, a 12 km thick feldspathic middle crust, a 5 km thick granulitic lower crust and a lithospheric mantle with a fixed thermal thickness (h_l). The oceanic lithosphere entails an 8 km thick crust and a lithospheric mantle with a thermal thickness that depends on its age following the plate model described in chapter 4. The passive margin is characterized by a crust that thins progressively towards the ocean over a distance of 150 km. The thermal base of the lithosphere at the passive margin is defined by a linear interpolation between the ocean and the continent.

Deformation is governed by frictional, dislocation, diffusion and Peierls creep equations, with parameters displayed in Table 6.1. The presented models do not account for any predefined strain softening parameterization. The initial geotherm is computed assuming steady-state conditions and accounts for different radiogenic heat productions in each layer and a constant asthenosphere temperature of 1330°C . The heat equation includes the effect of shear heating (Q_{sh}) and radiogenic heat production (Q_r). Shear heating describes the conversion of mechanical energy into heat by visco-plastic dissipation.

Figure 6.2: Model setup for (a) a uniform thickness of oceanic lithosphere; and (b) a model incorporating a mid-oceanic ridge. $+V_{in}$ is the velocity applied at the left model boundary and $-V_{out}$ the outflow. I_b , I_{pm} and I_o are the integrated stress at the model boundary, passive margin, and oceanic basin, respectively. Bottom sketches in (c) are the two predicted scenarios of subduction initiation: (right) subduction initiation at a passive margin for old oceanic lithospheres, (left) intra-oceanic subduction initiation for young oceanic lithospheres

6.3.3 Investigated parameters

For each model, we varied the thermal thickness of the oceanic lithosphere and the convergence rate at the boundary, which both influence the stress level within the oceanic lithosphere.

The first model setup (uniform thickness) comprises a 2000 km wide oceanic plate flanked by two continental plates on either side (Fig. 6.2a). 14 ages of oceanic lithosphere were tested, at 10 Myr intervals, counting from 0 to 140 Myr. For each age of oceanic lithosphere, a wide range of convergence rates were tested to arrive at a threshold value for which we observe a change from intra-oceanic subduction to subduction at the passive margin. We carried-out two sets of numerical experiments with two different thermal thicknesses (160 and 180 km) of continental lithosphere fixed in time, representing continental lithosphere of Palaeozoic and Proterozoic age, respectively (Artemieva, 2009).

Based on the results of step one, the thermal thickness of the continental lithosphere has been fixed at 160 km in stage two, as this thickness facilitates subduction initiation at passive margins. The thermal thickness and the length of the oceanic domain depends on the age of the lithosphere along the x-axis, as functions of distance from the former ridge. Therefore, by assuming that the ocean was formed at an average half-spreading rate of 1 cm yr^{-1} (in accordance with a slow spreading rate, Dick et al., 2003), the oceanic lithosphere gets 10 Myr older every 100 km (counting from the ridge) and the width of a 30 Myr oceanic basin is 600 km. We carried out three sets of numerical experiments with three different ages of oceanic basin namely 30 and 60 and 90 Myr. For each, the tested parameters include the duration of the cooling period at the start of the experiment and the convergence rate at the boundary during subsequent shortening. The implementation of a cooling period prior to shortening is necessary in order to obtain scenarios where subduction nucleates at the passive margin.

6.3.4 Boundary and loading conditions

The model top boundary is a true free surface (Duretz et al., 2016b). A constant inward normal velocity ($+V_{\text{in}}$) is applied to the left boundary of the model (Fig. 6.2c) to simulate horizontal tectonic force loading arising from the continent (Duretz et al., 2016a). The right boundary remains fixed. In order to satisfy mass conservation, an outflow velocity ($-V_{\text{out}}$) is distributed at the base of the model and on the sides from 200 km deep to the base of the box (Fig. 6.2a). The outflow is proportionally distributed over the length of each boundary such as:

$$\Sigma V_{\text{in}} H_{\text{in}} = V_{\text{out}} \Sigma H_{\text{out}} \quad (6.1)$$

These models with different thermal age of the oceanic lithosphere have been subject to horizontal velocity boundary conditions. For each age of oceanic lithosphere, a wide range of convergence rates were tested to arrive at a threshold value for which we observe a change from intra-oceanic subduction to subduction at the passive margin.

Table 6.1 Rheological and thermal parameters used for the reference numerical models.

Sediments	ρ (kg.m ⁻³)	k (W.m ⁻¹ .K ⁻¹)	Qr (W.m ⁻³)	Φ (°)	A (Pa.s ⁻ⁿ)	v (cm ³)	f	n	Q (J.mol ⁻¹)	Ref.
sediment1 (calcite)	2600	2.5	2.0e-6	20	1.59e-25	0	0	4.7	297.0e3	K_90
sediment2 (mica)	2400	2.5	2.0e-6	20	1.00e-138	0	0	18.0	51.0e3	K_90
Continental crust										
Upper (dry quartz)	2800	2.7	1.0e-6	30	3.98e-19	0	0	2.4	156.0e3	R_95
Middle (felsic granulite)	2800	2.7	0.6e-6	30	2.01e-21	0	0	3.1	243.0e3	R_95
Lower (mafic granulite)	3000	2.7	0.1e-6	30	5.04e-28	0	0	4.7	485.0e3	R_95
Oceanic crust (wet olivine)	2900	3	1.0e-10	30	5.68e-27	11e-6	1	3.5	480.0e3	H&K_03
Mantle lithosphere (dry olivine)										
Dislocation creep	3350	3.2	1.0e-10	30	1.10e-16	11e-6	0	3.5	530.0e3	H&K_03
Diffusion creep	3350	3.2	1.0e-10	30	1.50e-15	1.5e-16	0	1.0	375.0e3	H&K_03
Asthenosphere (dry olivine)										
Dislocation creep	3300	3.2	1.0e-10	30	1.10e-16	11e-6	0	3.5	530.0e3	H&K_03
Diffusion creep	3300	3.2	1.0e-10	30	1.50e-15	1.5e-16	0	1.0	375.0e3	H&K_03

Here ρ is the density, k is the thermal conductivity, Qr is the radiogenic heat production, Φ is the friction angle, A is a pre-factor, f is a correction factor, n is the stress exponent, G is the shear modulus, Q is the activation energy. References for rheology are H&K_03: Hirth and Kohlstedt (2003b); K_90: Kronenberg et al. (1990); R_95: Ranalli (1995).

6.3.5 Model output and stress analysis

With the aim of monitoring stress magnitude through time, at every location along the models, we compute for each time step the total integrated stress in the lithosphere I over the thermal thickness of the lithosphere hl :

$$I = \int_0^{hl} \tau II \, dz \quad (6.2)$$

where τII is the square root of the second invariant of the deviatoric stress tensor (Schmalholz et al., 2019). If the shear component is close to zero, like at the model boundary, τII is equal to the horizontal deviatoric stress σ_{xx}^d , (Schmalholz et al., 2019). Though hl corresponds to the 1330°C isotherm, we use the mechanical base of the lithosphere hm , which is fixed at the 1000°C geotherm ($\sim$ base of the strength envelope, e.g. Fig. 6.2a), for a better visualization of the modelling results.

A stress analysis is performed by computing the integrated deviatoric stress as a function of time (Fig. 6.3) at three positions: the model boundary, the passive margin, and the oceanic lithosphere (Fig. 6.2a). The integrated stress for the oceanic lithosphere is an averaged value over the oceanic domain. At each position, subduction initiation is assumed to start at the time where the integrated stress has reached the highest value and is followed by a sudden stress drop at ts , the time of subduction (Fig. 6.3). Figure 6.3 is an example of how the integrated stress I varies along the model (for a given set of model parameters). The variation of I describes the state of stress in the lithosphere in terms of

Figure 6.3: Example of integrated stress evolution through time and amount of convergence expressed as bulk shortening for an experiment where subduction initiates at the passive margin. The steep increase in the early stage represents elastic loading. The stress values presented in figure 6.4 are stresses integrated over the whole lithosphere for the three locations indicated by the coloured arrows in figure 6.2a (model boundary Ib , passive margin Ipm , oceanic basin Io). t_0 and t_s refers to the time at the start of the model and the time of subduction initiation, respectively.

increase or reduction of stress relative to the integrated boundary stress at the left side of the model. As such we interpret that when I at the passive margin or in the oceanic domain is lower than at the boundary, the stress is dissipated somewhere, for instance by the presence of ductile shear zones. If I is higher, we interpret that lithosphere undergoes a stress loading until stresses release due to the elastic strain and/or viscous thickening of the plate.

Models were run for a range of convergence velocities, ages of oceanic lithosphere, and thicknesses of continental lithosphere. In each model, the integrated stress at the model boundary was calculated and plotted in various domain diagrams (Fig. 6.4). We also use the evolution of integrated stress through time to explain how it controls the lithosphere dynamics (Fig. 6.5c). A total of 150 models were run in order to establish a robust stress transfer model for the oceanic basin.

6.4 Modelling results

6.4.1 Oceanic lithosphere with laterally uniform thickness

In this section, we present the results of the first set of experiments that shows the relationship of convergence rate and the locus of subduction initiation as a function of the age of the oceanic basin (Figs. 6.3a, c). Throughout this study we define t_0 as the age of the oceanic lithosphere at the start of the model and t_s as the age of the oceanic basin when subduction initiates. For each experiment, we calculate the value of integrated stress I_b at the onset of subduction initiation t_s (Figs. 6.3b, d). If subduction initiates at the passive margin the value is plotted as a red dot whereas it is shown in blue, when subduction initiates within the oceanic domain.

Figures 6.4a and 6.4c indicate that for young oceanic lithospheres ($\text{age} < 50$ Myr) subduction initiation at a passive margin is only feasible at low convergence rates. The upper bound value of convergence rate for subduction at passive margin ($\text{max } v$) is relatively constant ($\sim v = 0.8 \text{ cm yr}^{-1}$) and thus independent of the age of the oceanic lithosphere. Moreover, this value is relatively similar for both a 160 and 180 km thick continental lithosphere.

In contrast, for oceanic lithosphere older than 50-60 Myr, the upper bound values of convergence rate for subduction initiation at passive margin varies with the age of oceanic lithosphere at the start of the model as well as the thickness of the continental lithosphere. Intra-oceanic subduction is consistently observed for higher convergence rates across all tested ages of oceanic lithosphere (Figs. 6.4a, c). Moreover, the range of

Figure 6.4: Results of model series 1 with laterally uniform thickness of oceanic lithosphere: role of stress and convergence rate on the locus of subduction initiation for two thermal thickness of continental lithosphere, 160 km and 180 km. (a, c) age of oceanic lithosphere vs convergence rate at model boundary at the onset of convergence (t_0). (b, d) age of subduction vs integrated stress at model boundary at the onset of subduction initiation. The red and blue dot lines represent the minimum and maximum integrated stress for subduction at passive margin. →

velocities leading to subduction initiation at a passive margin is much larger in case of a 160 km thick continental lithosphere (Fig. 6.4a), than for a 180 km thick continental lithosphere (Fig. 6.4c), suggesting that subduction at passive margin is less likely in case of thick continental lithospheres. It shows again that for similar kinematic conditions, strain localization at passive margins is controlled by the strength of the continental lithosphere, regulated through its thermal thickness.

From the distribution of modelling results, we delineate upper and lower bounds of stress levels for subduction initiation (max. and min. stress, Figs. 6.4b, c). The results show that overall subduction at passive margins requires less stress than intra-oceanic subduction suggesting that subduction at passive margins is possible for stress levels lower than that within the oceanic plate (Fig. 6.4b, blue dash-dot line), which is predicted for oceanic lithospheres older than ~60 Myr (Fig. 6.4b). Consequently, with cooling of the oceanic lithosphere, intra-oceanic subduction becomes particularly intricate and requires large amount of stress.

The minimum stress necessary for subduction nucleation at a passive margin (red dashed line at 60 Myr, Fig. 6.4b) is in the order of 20 TN/m, which corresponds to an age of 60-70 Myr for the oceanic lithosphere. The stress magnitude required for subduction at passive margins is sensitive to the thermal thickness hl of the continental lithosphere. The minimum stress for subduction initiation at passive margins ranges from 18 to 25 TN/m for hl = 160 km and from 26 to 32 TN/m for hl = 180 km (Figs. 6.4b, d), which corresponds to age limits of oceanic lithosphere of 60 Myr for hl =160 km and 85 Myr for hl = 180 km. This result shows that among all the parameters analysed in this study, the thermal thickness of the continental lithosphere is probably most important. For similar ages of subduction initiation at passive margins, the disparity in stress is due to a variation in convergence rate with low convergence rate models requiring lower stress levels for subduction initiation at passive margins (red dashed line, Fig. 6.4b).

6.4.2 Mechanisms for subduction initiation

Subduction initiation at passive margins requires high loading level of shear stress acting on the margin but also low levels of shear stress in the oceanic lithosphere. Our modelling results show that convergence velocity is an important parameter that regulates stress in the lithospheric layers and therefore controls the locus of deformation.

Figure 6.5: Effect of convergence rate on mode of subduction initiation in model series 2. Inversion of a 30 Myr old oceanic basin at convergence rate of 0.8 cm yr^{-1} (left) and 1.4 cm yr^{-1} (right). (a, b) Model evolution, displayed in terms of composition of the lithosphere overlain by the second invariant of the strain rate tensor to show strain localization mechanism (top panel), and integrated stress I (black curve) and integrated heat production by visco-plastic dissipation IHS (red curve) over the thickness of the lithosphere (bottom panels). (c, d) Evolution integrated stress for three locations indicated by the coloured arrows (model boundary Ib , passive margin Ipm , oceanic basin Io). Note that Io is an averaged value of integrated stress over the oceanic domain. BS is bulk shortening. →

Models with low convergence rate ($\sim 0.8 \text{ cm yr}^{-1}$) predict strain localization and subsequent subduction initiation at the passive margin (Fig. 6.5a, 7%BS), whereas models with high convergence rate predict intra-oceanic subduction (Fig. 6.5b, 7%BS). These results show that stress loading preferentially occurs at passive margins for cases of low convergence rate (Fig. 6.5c), because deformation is distributed over many small structures within the brittle layer that accommodate low amounts of strain and is even more distributed within the ductile layers of the models.

Consequently, distributed deformation (black curve) leads to low shear rates and low heat production (Fig. 6.5a). The access heat (H_s , Fig. 6.5a) is then efficiently removed by thermal diffusion within the viscous layer (red curve, Fig. 6.5a, 7%BS). Therefore, homogeneous distribution of deformation and shear heating at low strain-rate limits the magnitude of stress in the oceanic domain and thus prevents stress loading and failure during cooling of the lithosphere (Fig. 6.5c). When the oceanic lithosphere reaches a certain thickness (after $t = 50\text{-}60 \text{ Myr}$), it barely deforms, but acts as a buttress and shortening leads to deformation of the passive margin until subduction initiates (Fig. 6.5a, 11%BS).

The passive margin lithosphere consists of brittle layers in the crust and the underlying lithospheric mantle where deformation localises in shear bands, which eventually link up to form a through-going shear structure (Fig. 6.5a). This moment of formation of the incipient subduction plate boundary allowing for underthrusting of the oceanic plate correlates with a significant stress drop as shown in figure 6.5c.

In contrast, for models with relatively high convergence rate deformation is localized in the oceanic domain (Fig. 6.5b, between $-500\text{-}0 \text{ km}$). The increase of integrated stress is largest within oceanic lithosphere for both young and old oceanic basins (Fig. 6.5d, $t = 20 \text{ Myr}$). Higher strain rates within shear band type-structures lead to increased shear heating (H_s) triggering thermally induced shear-localization in the ductile layer and the formation of a subduction plate boundary (Fig. 6.5b).

6.4.3 Oceanic lithosphere with age dependant thickness

In this section we present the results for three sets of experiments defined by the age of the oceanic basin (30 Myr, 60 Myr and 90 Myr; Figs. 6.6a, b, c) that include age dependent thickness variations within the oceanic lithosphere (Fig. 6.2b). Figure 6.6 delineates the mode of subduction as function of the cooling period implemented at the start of the experiments and the convergence rate during the subsequent shortening period. Overall, these modelling results show that a cooling period of at least 5 Myr before the start of shortening (Fig. 6.6a) is necessary to prevent underthrusting at the position where the thermal thickness of the oceanic lithosphere is smallest, representing in simplified way a former ridge. Consistent with results derived from step 1, we infer that subduction initiation at a passive margin is only feasible for low convergence rates ($v < 0.6 \text{ cm yr}^{-1}$, Fig. 6a). However, different to the models with a uniform oceanic lithosphere, figure 6.6 shows that subduction initiation is more favourable in case of a

Figure 6.6: Results of model series 2 with an age dependant thickness of oceanic lithosphere: Domain diagrams mapping the location of subduction initiation against the cooling time (t_c) and the convergence velocity (v) during the subsequent shortening period. Each diagram represents a different age of the oceanic basin (age of the oceanic lithosphere at the margin) before the start of the experiment.

young oceanic basin rather than an old one (Fig. 6a vs. Figs.6b, c). To understand this counter-intuitive result, we present in figure 6.7 two modelling results with similar boundary conditions ($t_c = 5$ Myr, $v = 0.3$ cm.yr⁻¹) but with different ages of the oceanic basin (age = 30 & 60 Myr; Fig. 6.6a, b) at the onset of shortening.

For both models, the oceanic lithosphere shows an increase in thickness and integrated stress from the basin centre to the continental lithosphere (Figs. 6.7a, b), which is largest for the older (60 Myr) oceanic basin (Figs. 6.7a, b). In case of a 30 Myr oceanic basin, the level of integrated stress at the ridge is close to the value at the margin (~ 8 -9 TN/m, black curve in Fig. 6.7g). Shortening of such young lithosphere at a slow rate of 0.3 cm.yr⁻¹ leads to strain localization at the former ridge early in the deformation history (Fig. 6.7a, 7 Myr). However, in the later stage (Fig. 6.7a, 30 Myr), strain is rather distributed, because lateral thermal thickness variations get reduced as the oceanic lithosphere cools faster at the ridge compared to the passive margin. Under-such conditions, the subsequent development of a shear-zone at the base of the continental crust leads to underthrusting of the oceanic plate under the margin (Fig. 6.7a, 46 Myr). In comparison, in case of a 60 Myr old oceanic basin the level of integrated stress at the ridge is significantly lower than at the passive margin ($\Delta I = 5$ TN, Fig. 6.7b, 7 Myr). Consequently, strain localises at the mid-oceanic boundary which is much thinner (Fig. 6.7b, 45 Myr), resulting in intra-oceanic under-thrusting (Fig. 6.7b, 45 Myr). These results suggest that the strength contrast (expressed in this section by the thermal thickness) between the lithospheres at the centre of the oceanic basin and the passive margin controls the locus of subduction initiation. Therefore, a young oceanic basin with minor variation in strength between the extinct ridge and the margin requires only 5 Myr of cooling prior to convergence to permit subduction at a passive margin. In contrast, an

Figure 6.7: Effect of oceanic basin age on subduction initiation in model series 2. Shortening of a 30 Myr old (left) and a 60 Myr old (right) oceanic basin at convergence rate of 0.3 cm yr^{-1} (left) after a cooling period of 5 Myr. (a-c) Model evolution, illustrated through plots of model composition overlain by plots of strain-rate (top panels) and through plots of integrated stress I (black curve) and integrated heat production by visco-plastic dissipation IHs (red curve) over the thickness of the lithosphere (lower panels). (c-d) Evolution of integrated stress for three locations: the model boundary, the passive margin, and in the oceanic basin.

old oceanic basin would require a significant period of cooling ($t_c > 20 \text{ Myr}$) to reduce the strength differences between the lithosphere at the ridge and at the passive margin to allow for subduction initiation at the passive margin (Fig. 6.6b).

6.5 Discussion

6.5.1 Favourable conditions for subduction initiation at passive margins

We have set to investigate what conditions and mechanisms lead to intra-oceanic subduction vs. passive margin subduction (Fig. 6.8). Because mid-oceanic ridges have a thinner and weaker lithosphere compared to passive margins, they represent *a priori* preferential locations for subduction initiation (e.g. Agard et al, 2016; Maffione et al., 2015). Moreover, the large difference between ridge push force ($\sim 1\text{-}5 \text{ TN}$) and stresses arising from tectonics, or topography related gravitational potential energy ($5\text{-}15 \text{ TN}$) suggests that subduction should predominantly initiate at mid-oceanic ridges, which is consistent with our modelling predictions for fast convergence rates. Additionally, this modelling study suggests that subduction initiation at passive margin is physically feasible under restricted conditions.

First, subduction initiation at a passive margin requires processes that regulate stress levels in the oceanic lithosphere. Our modelling results show that rheologic conditions and processes favouring distributed deformation within the ductile part of the oceanic lithosphere is key for reducing stress levels and de-localising deformation (Fig. 6.8b). As such localised deformation within the brittle layer, expressed as shear band type structures, fail to propagate into and through the ductile layer to form a subduction plate boundary. In contrast, rheological decoupling at the margin allows for the development of a long-lasting shear zone where stresses are relaxed through the formation of a decollement that propagate through the mantle. This behaviour is tied to low convergence rates, i.e. below 0.8 cm yr^{-1} (Fig. 6.8b). Similar results have been obtained by Gülcher et al. (2019), who observed in their numerical models, reactivation of multiple detachment faults upon their inversion but did not produce intra-oceanic subduction. Our results predict that even young (5 Myr old) oceanic lithospheres can support tectonic stress of up to 12 TN/m underlining its long-term stability. As such, intra-oceanic subduction initiation requires fast convergence (Fig. 6.8c) and/or the implementation of lithosphere-scale pre-existing weak zones (Maffione et al., 2015) to localize deformation.

Figure 6.8: Strain-rate dependent mechanisms for subduction initiation at passive margins. (a) Geodynamic processes that potentially contribute to the transfer of stress to a passive margin. (b) Subduction at passive margin for slowly converging ocean-continent systems: the formation of a subduction plate boundary at the passive margin critically depends on the stability of the mid-ocean ridge, which is controlled by the distributed style of deformation. 0-45 Myr after the onset of shortening: strain accumulation at passive margin under low convergence rate. >45 Myr: strain localization and subduction. (c) Subduction amid oceanic plates for fast converging ocean-continent systems: Rapid closure of the oceanic basin leads to localized deformation and underthrusting near the ridge crest. Black arrows represent stress arising from mid-oceanic ridge and far-field forcing.

Subduction initiation at a passive margin also entails the transfer of deformation from the ridge to the margin during the cooling of the oceanic lithosphere (Fig. 6.7 a). Previous studies have emphasized that deformation at the passive margin through spontaneous margin collapse is an unlikely mechanism for subduction initiation, because stresses acting on the passive margin lithosphere are never at yield (Cloetingh et al., 1984; Mueller & Phillips, 1991). It is thus more likely that the nucleation of a subduction zone at a passive margin occurs upon additional external forcing to reach stress levels of at least 8-10 TN/m (Zhong & Li, 2020) to induce failure of the passive margin lithosphere. Numerical modelling studies simulating the shortening of an oceanic basin (Auzemery et al., 2020; Candiotti et al., 2020; McCarthy et al., 2020; Zhong & Li, 2020) emphasise that such stress levels can only be supported by thick oceanic lithospheres, suggesting that subduction initiation is only feasible in case of shortening of an old oceanic basin. Therefore, subduction at a passive margin could only happen after a relatively long period of cooling at an extinct ridge prior to plate convergence (> 60 Myr, Candiotti, 2020; McCarthy et al., 2020). In comparison, our models simulating the effects of an extinct and cooling mid-oceanic ridge suggest that subduction initiation at passive margins is also possible in case of slow shortening of a young oceanic basin (< 30 Myr), following a short period of cooling (> 5 Myr). Although our approach assumes an initial short period (5 Myr) of cooling where no external forcing (convergence) is applied, the results predict a more reasonable time scale for the development of a subduction zone at a passive margin, which amounts to 45 Myr from the moment that shortening is applied.

Third, subduction initiation at continental margins does not only depend on the age of the oceanic lithosphere (e.g. Auzemery et al., 2020; Zhong & Li, 2020) but also on the thermo-tectonic age of the rifted continental lithosphere (Fig. 6.4). We show that subduction initiation is favoured for thermally thin (Palaeozoic or younger) continental lithospheres (< 160 km) over cratons (> 180 km) (Artemieva, 2009). This could explain why subduction does not exist along the north and south Atlantic margin, where the continental lithosphere is particularly thick (Tesauro et al., 2013) and might also be the reason for the proposed transference of subduction from the paleo-Tethys, to the neo-Tethys, along the margin of the thinned Cimmerian micro-continent (Wan et al., 2019).

6.5.2 Model limitations

In our study, a very thin oceanic lithosphere is used to justify the implementation of a relatively weak load bearing lithosphere simulating an extinct mid-ocean ridge. As such melt- and fluid related processes (Dymkova & Gerya, 2013), processes, which are yet poorly constrained because of scarce high resolution observations and rheology data from natural systems (Bickert et al., 2019) are not taken into account. These have been discussed in models deploying lithosphere extension (e.g. Dannberg et al., 2019; Ligi et al., 2008), which is different to our setup, which focusses on contraction of an oceanic basin and its continental margins. Moreover, our models do not include inherited structures such as detachment faults that may reduce the stress level needed for subduction initiation (Beaussier et al., 2019; Gülcher et al., 2019b; Maffione et al., 2015) or mantle upwelling at ridges that can push plates apart (Rowley et al., 2016). Therefore,

the stress levels obtained in our study are larger than plate tectonic forces and represent upper bounds. In addition, we note that the earlier mentioned 20 TN/m limit is only valid for a 4 layer-continental lithosphere. This limit strongly varies with rheological layering and thermal state at the margin (Auzemery et al., 2020) and explains differences with similar studies (e.g., Zhong & Li, 2019, 6 TN/m), in which the crust is weaker and thermal thickness at the margin is significantly smaller than in our models.

6.5.3 Subduction initiation in the Wilson cycle

Our modelling results together with previous modelling studies (Cloetingh et al., 1989; Gurnis et al., 2004; Hall, 2019; Nikolaeva et al., 2011; Stern & Gerya, 2018) suggest a mechanism that favors forced subduction initiation at passive margins. In particular, we find that slow convergence ($< 0.8 \text{ cm yr}^{-1}$) over a long period of time ($\sim 40 \text{ Myr}$) is critical in this context. Forces at play may include ridge push (e.g. Forsyth, 1973; Vlaar & Wortel, 1976) GPE from adjacent high areas (Ghosh et al., 2006; Marques et al., 2013; Pascal & Cloetingh, 2009), transference from an existing subduction zone (Baes et al., 2018; Duarte et al., 2013) or mantle flow (Candioti et al. 2020 and references therein). Associated with this far-field tectonics (Fig. 6.8a), the partial starvation of basaltic melt underneath an ultraslow spreading ridge can lead to the gradual increase in thermal thickness of the oceanic lithosphere by 40-50 km (Husson, 2012) resulting in a period where stresses transmitted from the extinct ridge affect the passive margin (Fig. 6.7).

In the frame of the above, the Alpine Tethys is a particular good analogue where the geological record demonstrates ultra-slow spreading (Lagabrielle & Cannat, 1990) and subduction initiated at the passive margin (Manzotti et al., 2014b; Marroni et al., 2017; Schmid et al., 2004b). The latter process was probably associated with low convergence rates affecting the Adriatic crust of Palaeozoic ($\sim 320 \text{ Ma}$) tectono-thermal age (Castellarin & Cantelli, 2010). The actual age of subduction initiation in the Alps is debated, but high pressure rocks found in the Adriatic margin constrain an age for underthrusting of ca 90-80 Ma (Manzotti et al., 2014a and references therein). However, several authors suggest an earlier subduction initiation at ca. 130 Ma, related to the change in Africa-Europe convergence controlled by the opening of the south Atlantic Ocean (Handy et al., 2010; van Hinsbergen et al., 2020). Based on our modelling results, we argue that subduction initiation in the Alps was a long-lasting (40-50 Myr) process that required a long period of very slow convergence allowing for strengthening of the ridge and the stress build-up at the ocean-continent transition. We note that the latter is achieved for times when Africa did not move head-on relative to Europe. During this period from 130 to 85 Ma, referred as “Cretaceous Quiet Zone” by van Hinsbergen et al. (2020), plate reorganisation leads to a quasi-absence of convergence between Africa and Europe (c.a. 0.6 cm yr^{-1} based on Fig. 6.9). We suggest that such configuration would favour a reduction of the spreading rate, the strengthening of the mid-oceanic ridge as well as strain accumulation at the margin, until subduction finally initiates due to an increase in plate convergence rate 90 Ma (Capitanio & Goes, 2006). This scenario implies that at the onset of slow convergence at 130 Ma, the Piemonte-Liguria ocean is

Figure 6.9: Plate motion history of Africa with respect to Europe from the Triassic to present day modified after van Hinsbergen et al. (2020). Red and green lines represents the motions path calculated by van Hinsbergen et al. (2020) and Handy et al. (2010), respectively. Numbers are ages in Ma. The yellow and blue lines represent the outline of Africa and illustrates its position at 200Ma and 130Ma (van Hinsbergen et al. 2020). The motion paths highlight a period of slow Africa-Europe convergence between 130 and 90 Ma, which is key for the future development of a subduction zone along the Adriatic passive margin.

30-40 Myr old, which falls within the range of “favourable” conditions for SI at passive margins (Fig. 6.6a). We furthermore remark that exhumed and serpentinitised mantle lithosphere of the Alpine Tethys provided additional favourable mechanical conditions for subduction initiation at stress levels that are compatible with plate tectonic forces (see chapter 5).

6.6 Conclusion

Subduction initiation at passive continental margins requires a sensitive balance between horizontal deviatoric stresses and the strength of the lithosphere, which acts as a stress guide. Both can vary as a function of the thermal thickness of the continental lithosphere at the margin (i.e. its age), the age of the oceanic lithosphere and the convergence rate. Strain accumulation at the passive margin during a long period of time with very slow convergence enables the development of a long-lasting shear-zone in the lower crust that reduces the stress level necessary for passive margin inversion. At the same time, this shear zone is critical for connecting localised deformation within the

brittle crust to deformation in the mantle lithosphere. This evolution leading to the formation of a subduction plate boundary at the passive margin critically depends on the stability of the mid-ocean ridge, which is controlled by the distributed style of deformation within the ductile oceanic lithosphere. This has a de-localising effect, which prevents the formation of a throughgoing shear zone and maintains a low level of stress in the lithosphere. Under these conditions, oceanic plate cooling together with gravitational stresses and far-field tectonic forces provide suitable driving forces for subduction nucleation at passive margins. In contrast, models with high convergence rate favour strain localization also within the ductile oceanic lithosphere, because shear heating mechanism are more effective, predicting the formation of a subduction plate boundary at the weakest spot of the system, the mid-oceanic ridge.

Chapter 7

Chapter 7

Concluding remarks

The aim of this thesis is to quantify mechanical conditions for induced subduction initiation at passive continental margins. For that purpose, a novel combination of physical analogue and numerical modelling has been employed to gain new insights in temporal and spatial scales of deformation during the formation and development of a subduction zones. Both methodologies were used to infer the influence of key parameters, such as the convergence rate, rheology and geometry of passive continental margins inherited from the rift phase, on processes controlling the localisation and style of passive margin deformation, together with propagation of deformation in the mantle lithosphere (chapters 3-6).

The modelling results highlight that subduction-initiation at passive continental margins is the result of a complex interaction between driving forces, rheological layering and localisation mechanisms. Furthermore, modelling predictions are consistent with observations from inverted passive continental margins (chapter 3) and the orogenic systems, such as the Alps and Dinarides/Hellenides (chapter 5), demonstrating that the mechanisms discussed in this study are relevant for natural cases of subduction initiation.

7.1 Mechanisms and evolution of induced subduction initiation at passive continental margins

The formation of a new subduction zone requires the development of a lithosphere-scale shear zone, representing the future subduction plate boundary (England & Wortel, 1980; Mueller & Phillips, 1991). Previously research has demonstrated that subduction initiation is physically difficult to simulate in the absence of a pre-existing lithospheric scale weak zone that promotes gravitational instability (e.g. Cloetingh et al., 1989; Gurnis et al., 2004; Stern & Gerya, 2018), suggesting that additional far-field forcing is needed for subduction initiation (Arcay et al., 2020; Candioti et al., 2020; Zhong & Li, 2020). This observation is in particular relevant for continental passive margins, where the strength of the oceanic and continental lithospheres increases with time, requiring progressively higher stress levels for failure (Cloetingh et al., 2008). Such failure is expected to occur at strength minima, which coincides with the ocean-continent transition where dense oceanic lithosphere is laterally replaced with buoyant continental lithosphere.

Similar to other studies (e.g. Nikolaeva et al., 2010; Zhong & Li, 2020), the results of chapter 3 confirm that critical factors for subduction initiation are the rheological stratification, and strength of the crust as well as that of the mantle lithosphere at passive margins. In addition, our results provide a detailed description on (1) the mechanics of subduction initiation and (2) characteristic length scale of deformation. In particular, the modelling results show that the early-stage deformation commences at the ocean-continent transition through the activation of thrust faults that

Figure 7.1: 3-stage tectonic evolution of subduction initiation at passive margins including (a) inversion at passive margin, (b) underthrusting, and (c) subduction including characteristic length-scales of deformation.

propagate continent-ward across the margin (Fig. 7.1a). Associated deformation involves flexure of the oceanic plate and folding of the passive continental margin over a characteristic length scale of 500-600 km. The subsequent shear localization within the ductile passive margin crust leads to the formation of a decollement, which facilitates underthrusting of the oceanic lithosphere (Fig. 7.1b). Localization of deformation reduces the zone affected by deformation to ~ 200 -300 km. Subsequently, the crustal shear zone propagates into the mantle lithosphere to accommodate the downward motion of the subducting plate. At this stage, the area affected by deformation remains constant over time (Fig. 7.1c).

The above suggests that observations documenting passive margin inversion can be interpreted as an expression of subduction initiation as suggested for the Brazilian (Marques et al., 2014) or the Korean margins (Kim et al., 2018). However, our modelling results show that the area of the inverted passive margin does not necessarily coincide with the ultimate locus of the subduction zone, because the latter is controlled by the strength of the lithospheric mantle in such a way that a weak or a strong mantle lithosphere promote migration of deformation to the oceanic or to the continental domain, respectively (chapter 3).

7.2 Strain localization and propagation of deformation in the mantle lithosphere

As deformation and failure of the mantle lithosphere is physically difficult to achieve, numerical and analogue models of subduction initiation often invoke pre-existing lithospheric weak zones (Mart et al., 2005; Goren et al., 2008; Nikolaeva et al., 2010). As shown also by Kiss et al. (2020), we furthermore demonstrate that subduction initiation is physically feasible in the absence of pre-defined mantle structures, by the evolution of major shear zones in the crust that propagates into the mantle where thermally induced shear heating processes within Peierl's creep domain facilitate strain localization and the formation of a subduction plate boundary (Duretz et al., 2014; Thielmann & Kaus, 2012). Our numerical models control this feature by the presence of

a decoupling horizon at the base of the continental crust that is inherent to the rheological stratification of the passive margin (Fig. 7.1b), which is different to thermal-induced strain localization (Crameri & Kaus, 2010; Thielmann & Kaus, 2012). Our modelling results also show that if no shear localization takes place at the base of the passive margin crust, subduction initiates either in the oceanic domain or in continental domain (Fig. 7.2; see also chapters 4 and 5).

After development of an incipient subduction zone, feedback mechanisms are required to maintain a weak subduction interface. Comparison between our analogue and numerical modelling results infers that shear heating (chapter 4), weak sediments, or a serpentinite mantle wedge (chapter 5), are required for a self-sustaining subduction zone. Therefore, depending on the presence or absence of above-described mechanisms, the localisation and mode of deformation of subduction initiation can vary greatly, including: intra-oceanic subduction, subduction at the passive margin, intra-continental subduction, pop-down subduction, double sided subduction or lithosphere folding (in both numerical and analogue models, Fig. 7.2).

Figure 7.2: Comparison between (a) numerical and (b) analogue modelling results of subduction initiation. The development of a long-lasting subduction fault is controlled by weakening mechanisms at the evolving plate interface. Note that oceanic lithosphere is represented by a monomineralic layer of olivine.

7.3 Favourable parameter combination for subduction initiation at passive margin

Our research presents new insights on the parameter space (Fig. 7.3) relevant for subduction initiation at passive continental margins and complements the findings of earlier studies (e.g. Candiotti et al., 2020; Faccenna et al., 1999; Kiss et al., 2020; Nikolaeva et al., 2010; Thielmann & Kaus, 2012; Zhong & Li, 2019).

First, our modelling results support the findings of Nikolaeva et al. (2010) and emphasise that crust-mantle decoupling at the passive continental margin is necessary

for the development of a shear zone facilitating underthrusting of the oceanic crust and the subsequent evolution of a subduction zone (Fig. 7.3c). In addition, chapter 5 emphasizes that subduction initiation is further controlled by the geometry and rheological layering of the continental crust at passive margins. Such features are inherent in the rheological characteristics of hyper-extended non-magmatic passive margins, where serpentinized mantle lithosphere is observed at the sea floor (e.g. Manatschal et al., 2001). In contrast, a magma-rich rifted margin (Geoffroy, 2005) is characterized by an increase in strength caused by the underplating and intrusion of mafic material into the continental crust. This leads to a rheological coupling between the continental crust and the mantle lithosphere and the early phases of margin deformation are followed by a transfer of deformation to the oceanic domain where an intra-oceanic subduction zone nucleates.

Second, the likelihood of subduction initiation at passive margins increases with a large strength contrast at the ocean-continent transition, facilitating localization of deformation (chapters 3 & 4). Suitable conditions include an intermediate age (ca. 50–100 Myr) of oceanic lithosphere (Fig. 7.3a, Auzemery et al., 2020; Zhong & Li, 2020) together with a continental lithosphere of Palaeozoic thermo-tectonic age (with a thermal thickness of less than ~170 km, Fig. 7.3b). This observation suggests that subduction is

Figure 7.3: Synthesis of the main driving factors for subduction initiation at passive margins. The locus of subduction is a function of a complex interplay between thermo-tectonic age of (a) the oceanic and (b) the continental lithosphere, (c) crust-mantle coupling at the passive margin and (d) convergence rate.

more likely to initiate along thinned micro-continents such as magmatic arc systems in the Pacific ocean (e.g. western Japan, Hall, 2019) than along cratonic passive continental margins, such as the South American or African continents (e.g. Niu et al., 2003).

Third, because subduction at passive margins requires stress levels much higher than ridge-push (e.g. Cloetingh et al., 1989; Kiss et al., 2020; Zhong & Li, 2019), it is critical to gain insight in the balance between the compressional stresses acting on the lithosphere and the strength of the oceanic lithosphere that varies through time by cooling. In this context we note that the strength of the mid-ocean ridge is higher when the melt concentration is low (e.g. for ultra-slow spreading ridge systems, Husson et al., 2015). In chapter 6, we show that a combination of oceanic plate cooling and slow convergence rates (~ 0.8 cm/yr, Fig. 7.3d) over a period of 40-60 Myr creates mechanical stability of the mid-ocean ridge and favours strain accumulation at the passive margin. Therefore, these conditions together with a transfer of stress from regions of high gravitational potential energy (Pascal & Cloetingh, 2009), from an existing subduction zone (Duarte et al., 2013), from another mid-ocean ridges (Forsyth, 1973) or from mantle convection (Candioti et al., 2020) may drive the nucleation of new subduction zones. In contrast, fast convergence of young or old oceanic lithosphere predicts intra-oceanic or intra-continental subduction, respectively.

7.4 Subduction initiation in the Alpine system

Our modelling results agree and further constrain observations of subduction initiation processes in the Alpine system indicating that subduction of the Alpine Tethys nucleated at the Adriatic passive continental margin (Schmid et al., 2020; Manzotti et al., 2014a; Marroni et al., 2017). The magma-poor, Adriatic passive continental margin with a reconstructed length of 150-200 km and an inherited Palaeozoic thermo-tectonic age favoured subduction initiation, further enhanced by the structural and mechanical grain related to the rifting phase. This grain was inherited from previously formed detachment faults exhuming and serpentinitising the mantle lithosphere (Mohn et al., 2011). In agreement with observations (Manzotti et al., 2014a; Mohn et al., 2011) our modelling documents the contractional reactivation of the serpentinitized mantle layer as detachment fault, which connects with a shear zone at the base of the continental crust to form a continuous shear structure that transfers deformation to the mantle lithosphere and initiates subduction in the area of the necking domain. Furthermore, the slow Africa – Europe convergence commencing 50 Myr before the nucleation of subduction coincide with the suitable boundary conditions (convergence rate < 0.8 cm/yr) for subduction initiation. Therefore, we speculate that the slow convergence rate starting at 130 Ma (van Hinsbergen et al., 2020) contributed to a reduction of the spreading rate in the Piemonte-Liguria ocean and thus to strengthening of the mid-oceanic ridge (Husson et al., 2015) ultimately leading to induced subduction 60-70 Myr after the rifting. This unique example contrasts with other modes of subduction recorded in the Dinarides and beyond (e.g. Oman, Anatolia), where rheological and kinematic conditions for subduction initiation at continental margins were not met.

7.5 Future perspectives

The results of this study provide insights into the mechanisms and favourable rheological and kinematic conditions for subduction initiation at passive continental margin where convergence is orthogonal to an ocean-continent transition. In particular, these results have shown that localisation of subduction initiation is controlled by the passive margin geological conditions and no “one fits all” model can be postulated. These results demonstrate that the complexity of subduction initiation processes require further refinement of the parameter space to obtain more control in the complex feedback mechanisms at play.

An important step forward will be the implementation of full 3D numerical models to better account for lateral variations of passive continental margin geometries rheologies and kinematics. A number of studies have shown promising results on the spatial evolution of subduction zones (e.g. Gülcher et al., 2019a; Malatesta et al., 2013; Zhou et al., 2020). Higher resolution models are required to better resolve deformation mechanisms and propagation of deformation laterally and vertically during subduction initiation. Additionally, larger-scale 3D models are required to study the effects of plate reorganizations and oblique convergence on subduction initiation. Such thermo-mechanical models need to be better linked with the geological record and plate tectonic reconstructions. Such improved coupling is needed for hypothesis testing and the analysis of subduction systems, such as in the Neotethys example, where subduction initiation was likely controlled by inherited structures, which are reactivated during a change in plate direction and convergence rate (e.g. van Hinsbergen et al., 2020). Furthermore, a stronger coupling of 3D geodynamic with geomorphological modelling allows better coupling of numerical and analogue models with the architecture of sedimentary basins (Balázs et al., 2021).

Another intriguing insight for subduction initiation at continental passive margins relates to the fact that most pre-existing weak structures are located within the oceanic domain, whereas most of the existing subduction systems are at continental margins. This apparent contradiction is not well understood, but it may relate to fundamental aspects of subduction initiation including the driving forces together with the strength and stability of plate boundaries. For instance, mid-ocean ridges are intrinsically weak features that should favours subduction initiation. Yet they are stable over geological time scales as testified by the Atlantic or Indian mid-oceanic ridge. Chapter 4 highlights that mid-oceanic ridges or other rheologically weak structures can accommodate deformation without necessarily creating a new subduction zone. This topic requires more investigation and integration of multi-scale processes. Of particular interest is the interaction between magmatic and hydro-thermomechanical processes at mid-oceanic ridges with external tectonic forcing (Yamato et al., 2013) and mantle flow (Funicello et al., 2003; Whittaker et al., 2015).

This study proposes mechanisms and conditions for the nucleation of subduction zones that respects the general physics of plates tectonics. Nevertheless, it is unclear if these mechanisms are also operating under different rheological configurations such as envisaged for early Earth or exo-planets (van Hunen & van den Berg, 2008). In

particular, thermally driven localization mechanisms in semi-brittle layers are thought to be key for the development of subduction faults (Thielmann & Kaus, 2012). In addition, the effects of sediments have been inferred to be critical for setting up the plate tectonic cycle during the early stages of Earth history (Sobolev and Brown (2019), when subduction was likely an unstable process associated with sinking of material by pop-down style deformation (chapter 3), or double-sided deformation (Fig. 7.2b).

ACKNOWLEDGEMENTS

Acknowledgements

This thesis is the results of 4 years of hard work but also of very joyful and festive moments that gave me the opportunity to learn and grow, both as a researcher and as an individual. The completion of this project could not have been accomplished without the support of the people around me.

Firstly, I would like to thank my supervisors . Ernst for his fantastic support, his devotion and his kindness. Thank you for your time, guidance, and for the confidence you gave me. I am also happy that you finally accepted that croissants are French. Dimitrios, “le comte”, for your elegant guidance, and the numerous coffee breaks used to discuss your tastes for art, history, philosophy. Fred, Liviu and Sierd for being always enthusiastic about my research and for your very important suggestions in my manuscripts.

I am very grateful to my dear colleague, co-author and friend J.P. Brun. You have been an exemplary and visionary mentor to me. Thank you for guiding me on the right path until your last breath.

A big thanks to Thibault and Pipo. I am very grateful for everything you provided during the last four years. You have helped a great deal in my research. Thank you so much for always answering my stupid questions and for the time you spent on correcting my mistakes. I also had such a wonderful time in Rennes thanks to you and your pizza, fondues and galettes.

I am also indebted to the reading committee for their valuable comments and suggestions. My sincere thanks to Francesca, Taras, Douwe, Jeroen, Stefan.

Dear Tecto-friends, I spent with you ones of the best years of my life. Many thanks to Damien, Elisa, Fauzi, Inge, István, Jeroen, Jon, Lukman, Manfred, Ylona, for the good company and the moral support that contributed to the success of my work. Thank you Atilla for your last refreshment whisky. Thank you Anouk, for your never ending smile and your passion for carnival. Thanks Nevena and Marcus for introducing Oude Pothuis to me. Thanks Sahar for your fantastic Kebab. Thank you also to Tessa, Yola, Job, Mike and Sjaak for your support in the Lab and the good laughs. A big thanks to my closest friends in Utrecht, Simon, Nemanja, Kristof, Ester and Anna. I will always remember the fantastic activities we had together. I will always remember our vegetarian dinners and the good spirit. I will always remember our crew.

Of course, I am not forgetting the non-Tecto friends who were so important during the pandemic period. Thanks to my Hungarian friends for our excellent discussions and for the good laughs. Thanks to the expat-group for the funny memes that I used as a distraction. Thanks Andrea, Boris, Cassandre, Phiphou for being such inspiring friends and for supporting me in the difficult moments. In particular, I would like to thank the Carpathian “cigognes” for being on my side. My dear Subitops, It was a great chance for me to be part of such an inspiring and fun group. Especially, the time spent together in the field made unforgettable memories.

I also want to thank my girlfriend Tracy who was stuck with me during the long months of writing. I would like to express my deep and sincere gratitude to you for being

so supportive, for accepting my cooking, for entertaining me in the hard time and also for being there to celebrate my successes.

My dear Family. Tout ceci n'aurait pas été possible sans vous. C'est vous qui m'avez construit et guidé. C'est vous qui m'avez inspiré. Merci à ma chère maman, à Luc qui m'ont toujours accepté tel que je suis. Merci à mon papa, à Valérie, qui ont toujours soutenu mes projets. Merci à mes grand parents qui m'ont appris la vie. Merci à mes frères qui ont tout vécu à mes côtés.

References

- Afonso, J., Ranalli, G., & Fernandez, M. (2007). Density structure and buoyancy of the oceanic lithosphere revisited. *Geophysical Research Letters*, 34(10).
- Agard, P., Yamato, P., Soret, M., Prigent, C., Guillot, S., Plunder, A., Dubacq, B., Chauvet, A., & Monié, P. (2016). Plate interface rheological switches during subduction infancy: Control on slab penetration and metamorphic sole formation. *Earth and Planetary Science Letters*, 451, 208-220.
- Aharonov, E., & Scholz, C. H. (2019). The brittle-ductile transition predicted by a physics-based friction law. *Journal of Geophysical Research: Solid Earth*, 124(3), 2721-2737.
- Alvarez-Marron, J., Rubio, E., & Torne, M. (1997). Subduction-related structures in the North Iberian Margin. *Journal of Geophysical Research: Solid Earth*, 102(B10), 22497-22511. doi:<https://doi.org/10.1029/97JB01425>
- Arcay, D., Lallemand, S., Abecassis, S., & Garel, F. (2020). Can subduction initiation at a transform fault be spontaneous? *Solid Earth*, 11(1), 37-62.
- Armitage, J., Collier, J., Minshull, T., & Henstock, T. (2011). Thin oceanic crust and flood basalts: India-Seychelles breakup. *Geochemistry, Geophysics, Geosystems*, 12(5).
- Artemieva, I. (2011). *Lithosphere: an interdisciplinary approach*: Cambridge University Press.
- Artemieva, I. M. (2009). The continental lithosphere: reconciling thermal, seismic, and petrologic data. *Lithos*, 109(1-2), 23-46.
- Assumpcao, M. (1992). The regional intraplate stress field in South America. *Journal of Geophysical Research: Solid Earth*, 97(B8), 11889-11903.
- Assumpção, M. (1998). Seismicity and stresses in the Brazilian passive margin. *Bulletin of the Seismological Society of America*, 88(1), 160-169.
- Assumpção, M., & Sacek, V. (2013). Intra-plate seismicity and flexural stresses in central Brazil. *Geophysical Research Letters*, 40(3), 487-491. doi:<https://doi.org/10.1002/grl.50142>
- Auzemery, A., Willingshofer, E., Yamato, P., Duretz, T., & Sokoutis, D. (2020). Strain localization mechanisms for subduction initiation at passive margins *Global and Planetary Change*. doi:<https://doi.org/10.1016/j.gloplacha.2020.103323>
- Babić, L., Hochuli, P. A., & Zupanec, J. (2002). The Jurassic ophiolitic mélange in the NE Dinarides: Dating, internal structure and geotectonic implications. *Eclogae Geologicae Helveticae*, 263-275.
- Babist, J., Handy, M., Konrad-Schmolke, M., & Hammerschmidt, K. (2006). Precollisional, multistage exhumation of subducted continental crust: The Sesia Zone, western Alps. *Tectonics*, 25(6).
- Baes, M., & Sobolev, S. V. (2017). Mantle flow as a trigger for subduction initiation: A missing element of the Wilson Cycle concept. *Geochemistry, Geophysics, Geosystems*, 18(12), 4469-4486.
- Baes, M., Sobolev, S. V., & Quinteros, J. (2018). Subduction initiation in mid-ocean induced by mantle suction flow. *Geophysical Journal International*, 215(3), 1515-1522.

- Balázs, A., Mañenco, L., Granjeon, D., Alms, K., François, T., & Sztanó, O. (2021). Towards stratigraphic-thermo-mechanical numerical modelling: Integrated analysis of asymmetric extensional basins. *Global and Planetary Change*, 196, 103386.
- Barros, L. V., Assumpção, M., Quintero, R., & Caixeta, D. (2009). The intraplate Porto dos Gaúchos seismic zone in the Amazon craton—Brazil. *Tectonophysics*, 469(1-4), 37-47.
- Bauer, K., Neben, S., Schreckenberger, B., Emmermann, R., Hinz, K., Fechner, N., Gohl, K., Schulze, A., Trumbull, R. B., & Weber, K. (2000). Deep structure of the Namibia continental margin as derived from integrated geophysical studies. *Journal of Geophysical Research: Solid Earth*, 105(B11), 25829-25853.
- Bayrakci, G., Minshull, T., Sawyer, D., Reston, T. J., Klaeschen, D., Papenberg, C., Ranero, C., Bull, J., Davy, R., & Shillington, D. (2016). Fault-controlled hydration of the upper mantle during continental rifting. *Nature Geoscience*, 9(5), 384.
- Beaussier, S. J., Gerya, T. V., & Burg, J.-P. (2019). Near-ridge initiation of intraoceanic subduction: Effects of inheritance in 3D numerical models of the Wilson Cycle. *Tectonophysics*, 763, 1-13.
- Beekman, F., Bull, J. M., Cloetingh, S., & Scrutton, R. (1996). Crustal fault reactivation facilitating lithospheric folding/buckling in the central Indian Ocean. *Geological Society, London, Special Publications*, 99(1), 251-263.
- Beltrán-Triviño, A., Winkler, W., von Quadt, A., & Gallhofer, D. (2016). Triassic magmatism on the transition from Variscan to Alpine cycles: evidence from U–Pb, Hf, and geochemistry of detrital minerals. *Swiss Journal of Geosciences*, 109(3), 309-328.
- Beltrando, M., Rubatto, D., & Manatschal, G. (2010). From passive margins to orogens: The link between ocean-continent transition zones and (ultra) high-pressure metamorphism. *Geology*, 38(6), 559-562.
- Benhenni, L., Quesnel, Y., Berguig, M., Samai, S., & Hamoudi, M. (2019). Joint modeling of potential-field data and geodynamic interpretation for northeast Algeria. *Journal of African Earth Sciences*, 159, 103566.
- Beniest, A., Koptev, A., Sassi, W., Guichet, X., & Leroy, S. (2018). *The thermal evolution of marginal basins: Results of thermo-mechanical numerical modelling*. Paper presented at the EGU General Assembly Conference Abstracts.
- Bickert, M., Cannat, M., Jammes, S., Tommasi, A., & Lavier, L. (2019). *How do detachment faults root into the deep lithosphere at magma-starved mid-ocean ridges?* Paper presented at the AGU Fall Meeting Abstracts.
- Billen, M. I., & Hirth, G. (2005). Newtonian versus non-Newtonian upper mantle viscosity: Implications for subduction initiation. *Geophysical Research Letters*, 32(19).
- Bills, B. G., Currey, D. R., & Marshall, G. A. (1994). Viscosity estimates for the crust and upper mantle from patterns of lacustrine shoreline deformation in the Eastern Great Basin. *Journal of Geophysical Research: Solid Earth*, 99(B11), 22059-22086.
- Bird, P. (2017). Stress field models from Maxwell stress functions: Southern California. *Geophysical Journal International*, 210(2), 951-963.
- Bonnet, S., Guillocheau, F., Brun, J. P., & Van Den Driessche, J. (2000). Large-scale relief development related to Quaternary tectonic uplift of a Proterozoic-Paleozoic basement: The Armorican Massif, NW France. *Journal of Geophysical Research: Solid Earth*, 105(B8), 19273-19288.

- Boonma, K., Kumar, A., García-Castellanos, D., Jiménez-Munt, I., & Fernández, M. (2019). Lithospheric mantle buoyancy: the role of tectonic convergence and mantle composition. *Scientific reports*, 9(1), 1-8.
- Bortolotti, V., Chiari, M., Marroni, M., Pandolfi, L., Principi, G., & Saccani, E. (2013). Geodynamic evolution of ophiolites from Albania and Greece (Dinaric-Hellenic belt): one, two, or more oceanic basins? *International Journal of Earth Sciences*, 102(3), 783-811.
- Bortolotti, V., Marroni, M., Pandolfi, L., & Principi, G. (2005). Mesozoic to Tertiary tectonic history of the Mirdita ophiolites, northern Albania. *Island Arc*, 14(4), 471-493.
- Bourgeois, O., Ford, M., Diraison, M., De Veslud, C. L. C., Gerbault, M., Pik, R., Ruby, N., & Bonnet, S. (2007). Separation of rifting and lithospheric folding signatures in the NW-Alpine foreland. *International Journal of Earth Sciences*, 96(6), 1003-1031.
- Brace, W., & Kohlstedt, D. (1980). Limits on lithospheric stress imposed by laboratory experiments. *Journal of Geophysical Research: Solid Earth*, 85(B11), 6248-6252.
- Bradley, D. C. (2008). Passive margins through earth history. *Earth-Science Reviews*, 91(1-4), 1-26.
- Broerse, T., Norder, B., Govers, R., Sokoutis, D., Willingshofer, E., & Picken, S. J. (2019). New analogue materials for nonlinear lithosphere rheology, with an application to slab break-off. *Tectonophysics*, 756, 73-96.
- Brun, J.-P. (2002). Deformation of the continental lithosphere: Insights from brittle-ductile models. *Geological Society, London, Special Publications*, 200(1), 355-370.
- Brun, J., & Beslier, M. (1996). Mantle exhumation at passive margins. *Earth and Planetary Science Letters*, 142(1-2), 161-173.
- Brun, J., & Cobbold, P. (1980). Strain heating and thermal softening in continental shear zones: a review. *Journal of Structural Geology*, 2(1-2), 149-158.
- Brun, J. P. (1999). Narrow rifts versus wide rifts: inferences for the mechanics of rifting from laboratory experiments. *Philosophical Transactions of the Royal Society of London. Series A: Mathematical, Physical and Engineering Sciences*, 357(1753), 695-712.
- Buck, W. R. (1991). Modes of continental lithospheric extension. *Journal of Geophysical Research: Solid Earth*, 96(B12), 20161-20178. doi: <https://doi.org/10.1029/91JB01485>
- Burke, K. (1976). The Chad Basin: an active intra-continental basin. In *Developments in Geotectonics* (Vol. 12, pp. 197-206): Elsevier.
- Burke, K., Dewey, J., & Kidd, W. (1977). World distribution of sutures—the sites of former oceans. *Tectonophysics*, 40(1-2), 69-99.
- Burov, E., & Cloetingh, S. (2010). Plume-like upper mantle instabilities drive subduction initiation. *Geophysical Research Letters*, 37(3).
- Burov, E., & Watts, A. (2006). The long-term strength of continental lithosphere: "jelly sandwich" or "crème brûlée"? *GSA today*, 16(1), 4.
- Burov, E. B. (2011). Rheology and strength of the lithosphere. *Marine and Petroleum Geology*, 28(8), 1402-1443.
- Burov, E. B., & Diament, M. (1995). The effective elastic thickness (T_e) of continental lithosphere: what does it really mean? *Journal of Geophysical Research: Solid Earth*, 100(B3), 3905-3927.

- Byerlee, J. (1978). Friction of rocks. In *Rock friction and earthquake prediction* (pp. 615-626): Springer.
- Calignano, E., Sokoutis, D., Willingshofer, E., Gueydan, F., & Cloetingh, S. (2015). Asymmetric vs. symmetric deep lithospheric architecture of intra-plate continental orogens. *Earth and Planetary Science Letters*, 424, 38-50.
- Candioti, L. G., Schmalholz, S. M., & Duretz, T. (2020). Impact of upper mantle convection on lithosphere hyperextension and subsequent horizontally forced subduction initiation. *Solid Earth*, 11(6), 2327-2357.
- Capitanio, F. A., & Goes, S. (2006). Mesozoic spreading kinematics: consequences for Cenozoic Central and Western Mediterranean subduction. *Geophysical Journal International*, 165(3), 804-816.
- Cassinis, G., Cortesogno, L., Gaggero, L., Perotti, C., & Buzzi, L. (2008). Permian to Triassic geodynamic and magmatic evolution of the Brescian Prealps (eastern Lombardy, Italy). *Bollettino della Società geologica italiana*, 127(3), 501-518.
- Castellarin, A., & Cantelli, L. (2010). Geology and evolution of the Northern Adriatic structural triangle between Alps and Apennine. *Rendiconti Lincei*, 21(1), 3-14.
- Célérier, J., Sandiford, M., Hansen, D. L., & Quigley, M. (2005). Modes of active intraplate deformation, Flinders Ranges, Australia. *Tectonics*, 24(6).
- Chapman, D. S., & Pollack, H. N. (1977). Regional geotherms and lithospheric thickness. *Geology*, 5(5), 265-268.
- Chemenda, A. I., Mattauer, M., & Bokun, A. N. (1996). Continental subduction and a mechanism for exhumation of high-pressure metamorphic rocks: new modelling and field data from Oman. *Earth and Planetary Science Letters*, 143(1-4), 173-182.
- Cheng, Q. (2016). Fractal density and singularity analysis of heat flow over ocean ridges. *Sci Rep* 6: 19167.
- Chenin, P., Picazo, S., Jammes, S., Manatschal, G., Müntener, O., & Karner, G. (2019). Potential role of lithospheric mantle composition in the Wilson cycle: a North Atlantic perspective. *Geological Society, London, Special Publications*, 470(1), 157-172.
- Chimpliganond, C., Assumpção, M., Von Huelsen, M., & França, G. S. (2010). The intracratonic Caraíbas–Itacarambi earthquake of December 09, 2007 (4.9 mb), Minas Gerais State, Brazil. *Tectonophysics*, 480(1-4), 48-56.
- Chulick, G. S., Detweiler, S., & Mooney, W. D. (2013). Seismic structure of the crust and uppermost mantle of South America and surrounding oceanic basins. *Journal of South American Earth Sciences*, 42, 260-276.
- Clerc, C., Jolivet, L., & Ringenbach, J.-C. (2015). Ductile extensional shear zones in the lower crust of a passive margin. *Earth and Planetary Science Letters*, 431, 1-7.
- Clerc, C., Ringenbach, J.-C., Jolivet, L., & Ballard, J.-F. (2018). Rifted margins: Ductile deformation, boudinage, continentward-dipping normal faults and the role of the weak lower crust. *Gondwana Research*, 53, 20-40.
- Cloetingh, S., Beekman, F., Ziegler, P. A., van Wees, J.-D., & Sokoutis, D. (2008). Post-rift compressional reactivation potential of passive margins and extensional basins. *Geological Society, London, Special Publications*, 306(1), 27-70.

- Cloetingh, S., & Burov, E. (2011). Lithospheric folding and sedimentary basin evolution: a review and analysis of formation mechanisms. *Basin Research*, 23(3), 257-290. doi:10.1111/j.1365-2117.2010.00490.x
- Cloetingh, S., Burov, E., & Poliakov, A. (1999). Lithosphere folding: Primary response to compression? (from central Asia to Paris basin). *Tectonics*, 18(6), 1064-1083. doi:10.1029/1999tc900040
- Cloetingh, S., & Burov, E. B. (1996). Thermomechanical structure of European continental lithosphere; constraints from rheological profiles and EET estimates. *Geophysical Journal International*, 124(3), 695-723. doi:<https://doi.org/10.1111/j.1365-246X.1996.tb05633.x>
- Cloetingh, S., Koptev, A., Kovács, I., Gerya, T., Beniest, A., Willingshofer, E., Ehlers, T. A., Andrić-Tomašević, N., Botsyun, S., & Eizenhöfer, P. R. (2021). Plume-Induced Sinking of Intracontinental Lithospheric Mantle: An Overlooked Mechanism of Subduction Initiation? *Geochemistry, Geophysics, Geosystems*, 22(2), e2020GC009482.
- Cloetingh, S., Wortel, M., & Vlaar, N. (1984). Passive margin evolution, initiation of subduction and the Wilson cycle. In H. Zwart, H. Behr, & O. JE (Eds.), *Tectonophysics* (Vol. 109, pp. 147-163).
- Cloetingh, S., & Wortel, R. (1986). Stress in the Indo-Australian plate. *Tectonophysics*, 132(1-3), 49-67.
- Cloetingh, S., Wortel, R., & Vlaar, N. (1989). On the initiation of subduction zones. In *Subduction Zones Part II*, (pp. 7-25): Birkhauser Verlag.
- Cloetingh, S. A. P. L., Wortel, M., & Vlaar, N. (1982). Evolution of passive continental margins and initiation of subduction zones. *Nature*, 297(5862), 139.
- Coblentz, D. D., & Richardson, R. M. (1996). Analysis of the South American intraplate stress field. *Journal of Geophysical Research: Solid Earth*, 101(B4), 8643-8657.
- Cogné, N., Gallagher, K., Cobbold, P. R., Riccomini, C., & Gautheron, C. (2012). Post-breakup tectonics in southeast Brazil from thermochronological data and combined inverse-forward thermal history modeling. *Journal of Geophysical Research: Solid Earth*, 117(B11).
- Corti, G. (2009). Continental rift evolution: from rift initiation to incipient break-up in the Main Ethiopian Rift, East Africa. *Earth-Science Reviews*, 96(1-2), 1-53.
- Crameri, F., & Kaus, B. J. (2010). Parameters that control lithospheric-scale thermal localization on terrestrial planets. *Geophysical Research Letters*, 37(9).
- Crameri, F., Magni, V., Domeier, M., Shephard, G. E., Chotalia, K., Cooper, G., Eakin, C. M., Grima, A. G., Gürer, D., & Király, Á. (2020). A transdisciplinary and community-driven database to unravel subduction zone initiation. *Nature communications*, 11(1), 1-14.
- Crameri, F., & Tackley, P. J. (2016). Subduction initiation from a stagnant lid and global overturn: new insights from numerical models with a free surface. *Progress in Earth and Planetary Science*, 3(1), 1-19.
- Crisci, C., Ferrara, G., Mazzuoli, R., & Rossi, P. (1984). Geochemical and geochronological data on Triassic volcanism of the Southern Alps of Lombardy (Italy): genetic implications. *Geologische Rundschau*, 73(1), 279-292.

- Csontos, L., & Vörös, A. (2004). Mesozoic plate tectonic reconstruction of the Carpathian region. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 210(1), 1-56.
- Dannberg, J., Gassmöller, R., Grove, R., & Heister, T. (2019). A new formulation for coupled magma/mantle dynamics. *Geophysical Journal International*, 219(1), 94-107.
- Davy, P., & Cobbold, P. (1991). Experiments on shortening of a 4-layer model of the continental lithosphere. *Tectonophysics*, 188(1-2), 1-25.
- de Lima, C. C. (2000). Ongoing compression across intraplate South America: observations and some implications for petroleum exploitation and exploration. *Revista Brasileira de Geociências*, 30(1), 203-207.
- De Min, A., Velicogna, M., Ziberna, L., Chiaradia, M., Alberti, A., & Marzoli, A. (2020). Triassic magmatism in the European Southern Alps as an early phase of Pangea break-up. *Geological Magazine*, 1-23.
- De Vicente, G., & Vegas, R. (2009). Large-scale distributed deformation controlled topography along the western Africa–Eurasia limit: tectonic constraints. *Tectonophysics*, 474(1-2), 124-143.
- Dewey, J. (1969). Continental margins: a model for conversion of Atlantic type to Andean type. *Earth and Planetary Science Letters*, 6(3), 189-197.
- Dick, H. J., Lin, J., & Schouten, H. (2003). An ultraslow-spreading class of ocean ridge. *Nature*, 426(6965), 405-412.
- Dimitrijevic, M., & Dimitrijevic, M. (1991). Triassic carbonate platform of the Drina-Ivanjica element (Dinarides). *Acta Geologica Hungarica*, 34(1-2), 15-44.
- Dixon, J. E., Dixon, T. H., Bell, D., & Malservisi, R. (2004). Lateral variation in upper mantle viscosity: role of water. *Earth and Planetary Science Letters*, 222(2), 451-467.
- Dombrádi, E., Sokoutis, D., Bada, G., Cloetingh, S., & Horváth, F. (2010). Modelling recent deformation of the Pannonian lithosphere: lithospheric folding and tectonic topography. *Tectonophysics*, 484(1-4), 103-118.
- Doré, A., & Lundin, E. (1996). Cenozoic compressional structures on the NE Atlantic margin; nature, origin and potential significance for hydrocarbon exploration. *Petroleum Geoscience*, 2(4), 299-311.
- Drury, M. R. (2005). Dynamic recrystallization and strain softening of olivine aggregates in the laboratory and the lithosphere. *Geological Society, London, Special Publications*, 243(1), 143-158.
- Duarte, J. C., Rosas, F. M., Terrinha, P., Schellart, W. P., Boutelier, D., Gutscher, M.-A., & Ribeiro, A. (2013). Are subduction zones invading the Atlantic? Evidence from the southwest Iberia margin. *Geology*, 41(8), 839-842.
- Ducea, M. N., Saleeby, J. B., & Bergantz, G. (2015). The architecture, chemistry, and evolution of continental magmatic arcs. *Annual Review of Earth and Planetary Sciences*, 43, 299-331.
- Duretz, T., Agard, P., Yamato, P., Ducassou, C., Burov, E. B., & Gerya, T. V. (2016a). Thermo-mechanical modeling of the obduction process based on the Oman ophiolite case. *Gondwana Research*, 32, 1-10.
- Duretz, T., Petri, B., Mohn, G., Schmalholz, S., Schenker, F., & Müntener, O. (2016b). The importance of structural softening for the evolution and architecture of passive margins. *Scientific reports*, 6(1), 1-7.

- Duretz, T., Schmalholz, S., Podladchikov, Y., & Yuen, D. (2014). Physics-controlled thickness of shear zones caused by viscous heating: Implications for crustal shear localization. *Geophysical Research Letters*, 41(14), 4904-4911.
- Dymkova, D., & Gerya, T. (2013). Porous fluid flow enables oceanic subduction initiation on Earth. *Geophysical Research Letters*, 40(21), 5671-5676.
- England, P., & Wortel, R. (1980). Some consequences of the subduction of young slabs. *Earth and Planetary Science Letters*, 47(3), 403-415.
- Erickson, S. G. (1993). Sedimentary loading, lithospheric flexure, and subduction initiation at passive margins. *Geology*, 21(2), 125-128.
- Evans, B., & Goetze, C. (1979). The temperature variation of hardness of olivine and its implication for polycrystalline yield stress. *Journal of Geophysical Research: Solid Earth*, 84(B10), 5505-5524.
- Faccenna, C., Giardini, D., Davy, P., & Argentieri, A. (1999). Initiation of subduction at Atlantic-type margins: Insights from laboratory experiments. *Journal of Geophysical Research: Solid Earth*, 104(B2), 2749-2766.
- Ferreira, J. M., Oliveira, T., Takeya, M. K., & Assumpção, M. (1998). Superposition of local and regional stresses in northeast Brazil: evidence from focal mechanisms around the Potiguar marginal basin. *Geophysical Journal International*, 134(2), 341-355.
- Flesch, L. M., Haines, A. J., & Holt, W. E. (2001). Dynamics of the India-Eurasia collision zone. *Journal of Geophysical Research: Solid Earth*, 106(B8), 16435-16460.
- Forsyth, D. (1973). Compressive stress between two mid-ocean ridges. *Nature*, 243(5402), 78-79.
- Franke, D. (2013). Rifting, lithosphere breakup and volcanism: Comparison of magma-poor and volcanic rifted margins. *Marine and Petroleum Geology*, 43, 63-87.
- Funiciello, F., Faccenna, C., Giardini, D., & Regenauer-Lieb, K. (2003). Dynamics of retreating slabs: 2. Insights from three-dimensional laboratory experiments. *Journal of Geophysical Research: Solid Earth*, 108(B4).
- Gawlick, H.-J., Sudar, M., Suzuki, H., Deric, N., Missoni, S., Lein, R., & Jovanović, D. (2009). Upper Triassic and Middle Jurassic radiolarians from the ophiolitic mélange of the Dinaridic Ophiolite Belt, SW Serbia. *Neues Jahrbuch für Geologie und Paläontologie-Abhandlungen*, 253(2-3), 293-311.
- Geoffroy, L. (2005). Volcanic passive margins. *Comptes Rendus Geoscience*, 337(16), 1395-1408.
- Gerbault, M. (2000). At what stress level is the central Indian Ocean lithosphere buckling? *Earth and Planetary Science Letters*, 178(1-2), 165-181.
- Gerya, T. V., Connolly, J. A., & Yuen, D. A. (2008). Why is terrestrial subduction one-sided? *Geology*, 36(1), 43-46.
- Gerya, T. V., Stern, R. J., Baes, M., Sobolev, S. V., & Whattam, S. A. (2015). Plate tectonics on the Earth triggered by plume-induced subduction initiation. *Nature*, 527(7577), 221-225.
- Gerya, T. V., & Yuen, D. A. (2003). Characteristics-based marker-in-cell method with conservative finite-differences schemes for modeling geological flows with strongly variable transport properties. *Physics of the Earth and Planetary Interiors*, 140(4), 293-318.

- Ghosh, A., Holt, W. E., Flesch, L. M., & Haines, A. J. (2006). Gravitational potential energy of the Tibetan Plateau and the forces driving the Indian plate. *Geology*, 34(5), 321-324.
- Ghosh, A., Holt, W. E., & Wen, L. (2013). Predicting the lithospheric stress field and plate motions by joint modeling of lithosphere and mantle dynamics. *Journal of Geophysical Research: Solid Earth*, 118(1), 346-368. doi:<https://doi.org/10.1029/2012JB009516>
- Göğüş, O. H., & Pysklywec, R. N. (2008). Near-surface diagnostics of dripping or delaminating lithosphere. *Journal of Geophysical Research: Solid Earth*, 113(B11).
- Goren, L., Aharonov, E., Mulugeta, G., Koyi, H. A., & Mart, Y. (2008). Ductile deformation of passive margins: A new mechanism for subduction initiation. *Journal of Geophysical Research: Solid Earth*, 113(B8). doi:<https://doi.org/10.1029/2005JB004179>
- Grollmund, B., & Zoback, M. D. (2003). Impact of glacially induced stress changes on fault-seal integrity offshore Norway. *AAPG bulletin*, 87(3), 493-506.
- Guillot, S., Schwartz, S., Reynard, B., Agard, P., & Prigent, C. (2015). Tectonic significance of serpentinites. *Tectonophysics*, 646, 1-19.
- Guilmette, C., Smit, M. A., van Hinsbergen, D. J., Gürer, D., Corfu, F., Charette, B., Maffione, M., Rabeau, O., & Savard, D. (2018). Forced subduction initiation recorded in the sole and crust of the Semail Ophiolite of Oman. *Nature Geoscience*, 11(9), 688-695.
- Gülcher, A., Beaussier, S., & Gerya, T. (2019a). On the formation of oceanic detachment faults and their influence on intra-oceanic subduction initiation: 3D thermomechanical modeling, *Earth Planet. Sc. Lett.*, 506, 195–208.
- Gülcher, A. J., Beaussier, S. J., & Gerya, T. V. (2019b). On the formation of oceanic detachment faults and their influence on intra-oceanic subduction initiation: 3D thermomechanical modeling. *Earth and Planetary Science Letters*, 506, 195-208.
- Guo, P., Niu, Y., Ye, L., Liu, J., Sun, P., Cui, H., Zhang, Y., Gao, J., Su, L., & Zhao, J. (2014). Lithosphere thinning beneath west North China Craton: evidence from geochemical and Sr–Nd–Hf isotope compositions of Jining basalts. *Lithos*, 202, 37-54.
- Gurnis, M., Hall, C., & Lavier, L. (2004). Evolving force balance during incipient subduction. *Geochemistry, Geophysics, Geosystems*, 5(7).
- Hall, R. (2019). The subduction initiation stage of the Wilson cycle. *Geological Society, London, Special Publications*, 470(1), 415-437.
- Hamai, L., Petit, C., Abtout, A., Yelles-Chaouche, A., & Déverchère, J. (2015). Flexural behaviour of the north Algerian margin and tectonic implications. *Geophysical Journal International*, 201(3), 1426-1436.
- Hamai, L., Petit, C., Le Pourhiet, L., Yelles-Chaouche, A., Déverchère, J., Beslier, M.-O., & Abtout, A. (2018). Towards subduction inception along the inverted North African margin of Algeria? Insights from thermo-mechanical models. *Earth and Planetary Science Letters*, 501, 13-23.
- Handy, M. R., Schmid, S. M., Bousquet, R., Kissling, E., & Bernoulli, D. (2010). Reconciling plate-tectonic reconstructions of Alpine Tethys with the geological–geophysical record of spreading and subduction in the Alps. *Earth-Science Reviews*, 102(3-4), 121-158.

- Hansen, F., & Carter, N. (1983). *Semibrittle creep of dry and wet Westerly granite at 1000 MPa*. Paper presented at the The 24th US Symposium on Rock Mechanics (USRMS).
- Hilaireret, N., Reynard, B., Wang, Y., Daniel, I., Merkel, S., Nishiyama, N., & Petitgirard, S. (2007). High-pressure creep of serpentine, interseismic deformation, and initiation of subduction. *Science*, 318(5858), 1910-1913.
- Hirauchi, K.-i., Fukushima, K., Kido, M., Muto, J., & Okamoto, A. (2016). Reaction-induced rheological weakening enables oceanic plate subduction. *Nature communications*, 7, 12550.
- Hirth, G., & Kohlstedt, D. (2003a). Rheology of the upper mantle and the mantle wedge: A view from the experimentalists. *Inside the subduction Factory*, 83-105.
- Hirth, G., & Kohlstedt, D. (2003b). Rheology of the upper mantle and the mantle wedge: A view from the experimentalists. *GEOPHYSICAL MONOGRAPH-AMERICAN GEOPHYSICAL UNION*, 138, 83-106.
- Holford, S., Hillis, R., Duddy, I., Green, P., Stoker, M., Tuitt, A., Backé, G., Tassone, D., & MacDonald, J. (2011). Cenozoic post-breakup compressional deformation and exhumation of the southern Australian margin. *The APPEA Journal*, 51(1), 613-638.
- Holford, S. P., Green, P. F., Duddy, I. R., Turner, J. P., Hillis, R. R., & Stoker, M. S. (2009). Regional intraplate exhumation episodes related to plate-boundary deformation. *Geological Society of America Bulletin*, 121(11-12), 1611-1628.
- House, M. A., Gurnis, M., Kamp, P. J., & Sutherland, R. (2002). Uplift in the Fiordland region, New Zealand: Implications for incipient subduction. *Science*, 297(5589), 2038-2041.
- Houseman, G. A., McKenzie, D. P., & Molnar, P. (1981). Convective instability of a thickened boundary layer and its relevance for the thermal evolution of continental convergent belts. *Journal of Geophysical Research: Solid Earth*, 86(B7), 6115-6132.
- Hu, J., Liu, L., Faccenda, M., Zhou, Q., Fischer, K. M., Marshak, S., & Lundstrom, C. (2018). Modification of the Western Gondwana craton by plume–lithosphere interaction. *Nature Geoscience*, 11(3), 203-210.
- Hubbert, M. K. (1937). Theory of scale models as applied to the study of geologic structures. *Bulletin of the Geological Society of America*, 48(10), 1459-1520.
- Hürlimann, N., Müntener, O., Ulmer, P., Nandedkar, R., Chiaradia, M., & Ovtcharova, M. (2016). Primary magmas in continental arcs and their differentiated products: petrology of a post-plutonic dyke suite in the Tertiary Adamello batholith (Alps). *Journal of Petrology*, 57(3), 495-534.
- Husson, L. (2012). The dynamics of plate boundaries over a convecting mantle. *Physics of the Earth and Planetary Interiors*, 212, 32-43.
- Husson, L., Yamato, P., & Bezous, A. (2015). Ultraslow, slow, or fast spreading ridges: Arm wrestling between mantle convection and far-field tectonics. *Earth and Planetary Science Letters*, 429, 205-215.
- Jaeger, J. C., Cook, N. G., & Zimmerman, R. (2009). *Fundamentals of rock mechanics*: John Wiley & Sons.
- Jammes, S., & Lavier, L. L. (2016). The effect of biminerale composition on extensional processes at lithospheric scale. *Geochemistry, Geophysics, Geosystems*, 17(8), 3375-3392.

- Jaquet, Y., & Schmalholz, S. M. (2018). Spontaneous ductile crustal shear zone formation by thermal softening and related stress, temperature and strain rate evolution. *Tectonophysics*, 746, 384-397.
- Johnson, H. (2008). *The nature and origin of compression in passive margins*.
- Jolivet, L., Faccenna, C., Huet, B., Labrousse, L., Le Pourhiet, L., Lacombe, O., Lecomte, E., Burov, E., Denèle, Y., & Brun, J.-P. (2013). Aegean tectonics: Strain localisation, slab tearing and trench retreat. *Tectonophysics*, 597, 1-33.
- Kameyama, M., Yuen, D. A., & Karato, S.-I. (1999). Thermal-mechanical effects of low-temperature plasticity (the Peierls mechanism) on the deformation of a viscoelastic shear zone. *Earth and Planetary Science Letters*, 168(1-2), 159-172.
- Kapferer, N., Mercolli, I., Berger, A., Ovtcharova, M., & Fügenschuh, B. (2012). Dating emplacement and evolution of the orogenic magmatism in the internal Western Alps: 2. The Biella Volcanic Suite. *Swiss Journal of Geosciences*, 105(1), 67-84.
- Karig, D. E. (1982). Initiation of subduction zones: implications for arc evolution and ophiolite development. *Geological Society, London, Special Publications*, 10(1), 563-576.
- Karner, G., & Watts, A. (1983). Gravity anomalies and flexure of the lithosphere at mountain ranges. *Journal of Geophysical Research: Solid Earth*, 88(B12), 10449-10477.
- Kaufmann, G., & Amelung, F. (2000). Reservoir-induced deformation and continental rheology in vicinity of Lake Mead, Nevada. *Journal of Geophysical Research: Solid Earth*, 105(B7), 16341-16358.
- Keep, M., & Harrowfield, M. (2008). Elastic flexure and distributed deformation along Australia's North West Shelf: Neogene tectonics of the Bonaparte and Browse basins. *Geological Society, London, Special Publications*, 306(1), 185-200.
- Kelemen, P. B., & Holbrook, W. S. (1995). Origin of thick, high-velocity igneous crust along the US East Coast Margin. *Journal of Geophysical Research: Solid Earth*, 100(B6), 10077-10094.
- Kendall, J.-M., & Lithgow-Bertelloni, C. (2016). Why is Africa rifting? *Geological Society, London, Special Publications*, 420(1), 11-30.
- Kim, G.-B., Yoon, S.-H., Kim, S.-S., & So, B.-D. (2018). Transition from buckling to subduction on strike-slip continental margins: Evidence from the East Sea (Japan Sea). *Geology*, 46(7), 603-606. doi:10.1130/g40305.1
- Kirby, S. H. (1985). Rock mechanics observations pertinent to the rheology of the continental lithosphere and the localization of strain along shear zones. *Tectonophysics*, 119(1-4), 1-27.
- Kiss, D., Candiotti, L. G., Duretz, T., & Schmalholz, S. M. (2020). Thermal softening induced subduction initiation at a passive margin. *Geophysical Journal International*, 220(3), 2068-2073.
- Klemperer, S. L. (2003). An overview of the Izu-Bonin-Mariana subduction factory (Vol. 138, pp. 175-222): American Geophysical Union Washington, DC.
- Koptev, A., Beniest, A., Gerya, T., Ehlers, T. A., Jolivet, L., & Leroy, S. (2019). Plume-induced breakup of a subducting plate: Microcontinent formation without cessation of the subduction process. *Geophysical Research Letters*, 46(7), 3663-3675.
- Koptev, A., Cloetingh, S., & Ehlers, T. A. (2021). Longevity of small-scale ("baby") plumes and their role in lithospheric break-up. *Geophysical Journal International*.

- Korenaga, T., & Korenaga, J. (2016). Evolution of young oceanic lithosphere and the meaning of seafloor subsidence rate. *Journal of Geophysical Research: Solid Earth*, 121(9), 6315-6332.
- Kronenberg, A. K., Kirby, S. H., & Pinkston, J. (1990). Basal slip and mechanical anisotropy of biotite. *Journal of Geophysical Research: Solid Earth*, 95(B12), 19257-19278.
- Kruse, S. E., & Royden, L. H. (1994). Bending and unbending of an elastic lithosphere: The Cenozoic history of the Apennine and Dinaride foredeep basins. *Tectonics*, 13(2), 278-302.
- Krystopowicz, N. J., & Currie, C. A. (2013). Crustal eclogitization and lithosphere delamination in orogens. *Earth and Planetary Science Letters*, 361, 195-207.
- Lagabriele, Y., & Cannat, M. (1990). Alpine Jurassic ophiolites resemble the modern central Atlantic basement. *Geology*, 18(4), 319-322.
- Lagabriele, Y., Chauvet, A., Ulrich, M., & Guillot, S. (2013). Passive obduction and gravity-driven emplacement of large ophiolitic sheets: The New Caledonia ophiolite (SW Pacific) as a case study? *Bulletin de la Société géologique de France*, 184(6), 545-556.
- Le Breton, E., Brune, S., Ustaszewski, K., Zahirovic, S., Seton, M., & Müller, R. D. (2020). Kinematics and extent of the Piemonte-Liguria Basin—implications for subduction processes in the Alps. *Solid Earth Discussions*, 1-42.
- Leprêtre, A., Klingelhoefer, F., Graindorge, D., Schnurle, P., Beslier, M. O., Yelles, K., Déverchère, J., & Bracene, R. (2013). Multiphased tectonic evolution of the Central Algerian margin from combined wide-angle and reflection seismic data off Tipaza, Algeria. *Journal of Geophysical Research: Solid Earth*, 118(8), 3899-3916. doi:10.1002/jgrb.50318
- Leroy, M., Dauteuil, O., & Cobbold, P. R. (2004). Incipient shortening of a passive margin: the mechanical roles of continental and oceanic lithospheres. *Geophysical Journal International*, 159(1), 400-411.
- Lévy, F., & Jaupart, C. (2012). The initiation of subduction by crustal extension at a continental margin. *Geophysical Journal International*, 188(3), 779-797.
- Li, L., Stephenson, R., & Clift, P. D. (2016a). The Canada Basin compared to the southwest South China Sea: Two marginal ocean basins with hyper-extended continent-ocean transitions. *Tectonophysics*, 691, 171-184.
- Li, Z. H., Liu, M., & Gerya, T. (2016b). Lithosphere delamination in continental collisional orogens: A systematic numerical study. *Journal of Geophysical Research: Solid Earth*, 121(7), 5186-5211.
- Liati, A., Froitzheim, N., & Fanning, C. M. (2005). Jurassic ophiolites within the Valais domain of the Western and Central Alps: geochronological evidence for re-rifting of oceanic crust. *Contributions to Mineralogy and Petrology*, 149(4), 446-461.
- Ligi, M., Cuffaro, M., Chierici, F., & Calafato, A. (2008). Three-dimensional passive mantle flow beneath mid-ocean ridges: an analytical approach. *Geophysical Journal International*, 175(2), 783-805.
- Lohrmann, J., Kukowski, N., Adam, J., & Oncken, O. (2003). The impact of analogue material properties on the geometry, kinematics, and dynamics of convergent sand wedges. *Journal of Structural Geology*, 25(10), 1691-1711.

- Løseth, H., & Henriksen, S. (2005). *A Middle to Late Miocene compression phase along the Norwegian passive margin*. Paper presented at the Geological Society, London, Petroleum Geology Conference series.
- Lundin, E., & Doré, A. (2002). Mid-Cenozoic post-breakup deformation in the 'passive' margins bordering the Norwegian–Greenland Sea. *Marine and Petroleum Geology*, 19(1), 79-93.
- Lustrino, M., Abbas, H., Agostini, S., Caggiati, M., Carminati, E., & Gianolla, P. (2019). Origin of Triassic magmatism of the Southern Alps (Italy): Constraints from geochemistry and Sr-Nd-Pb isotopic ratios. *Gondwana Research*, 75, 218-238.
- Luth, S., Willingshofer, E., Sokoutis, D., & Cloetingh, S. (2010). Analogue modelling of continental collision: Influence of plate coupling on mantle lithosphere subduction, crustal deformation and surface topography. *Tectonophysics*, 484(1-4), 87-102.
- Luttrell, K., & Sandwell, D. (2012). Constraints on 3-D stress in the crust from support of mid-ocean ridge topography. *Journal of Geophysical Research: Solid Earth*, 117(B4).
- Mackwell, S., Zimmerman, M., & Kohlstedt, D. (1998). High-temperature deformation of dry diabase with application to tectonics on Venus. *Journal of Geophysical Research: Solid Earth*, 103(B1), 975-984.
- Maffione, M., Thieulot, C., Van Hinsbergen, D. J., Morris, A., Plümper, O., & Spakman, W. (2015). Dynamics of intraoceanic subduction initiation: 1. Oceanic detachment fault inversion and the formation of supra-subduction zone ophiolites. *Geochemistry, Geophysics, Geosystems*, 16(6), 1753-1770.
- Maffione, M., & van Hinsbergen, D. J. (2018). Reconstructing plate boundaries in the Jurassic neo-Tethys from the east and west Vardar ophiolites (Greece and Serbia). *Tectonics*, 37(3), 858-887.
- Maffione, M., van Hinsbergen, D. J., de Gelder, G. I., van der Goes, F. C., & Morris, A. (2017). Kinematics of Late Cretaceous subduction initiation in the Neo-Tethys Ocean reconstructed from ophiolites of Turkey, Cyprus, and Syria. *Journal of Geophysical Research: Solid Earth*, 122(5), 3953-3976.
- Mahatsente, R. (2017). Global models of ridge-push force, geoid, and lithospheric strength of oceanic plates. *Pure and Applied Geophysics*, 174(12), 4395-4406.
- Mahatsente, R., & Coblentz, D. (2015). Ridge-push force and the state of stress in the Nubia-Somalia plate system. *Lithosphere*, 7(5), 503-510.
- Malatesta, C., Gerya, T., Crispini, L., Federico, L., & Capponi, G. (2013). Oblique subduction modelling indicates along-trench tectonic transport of sediments. *Nature communications*, 4(1), 1-6.
- Malatesta, C., Gerya, T., Scambelluri, M., Federico, L., Crispini, L., & Capponi, G. (2011). *Serpentine channel and the role of serpentine buoyancy for exhumation of HP rocks (Voltri Massif, Western Alps)*. Paper presented at the Goldschmidt conference abstracts.
- Malavieille, J. (2010). Impact of erosion, sedimentation, and structural heritage on the structure and kinematics of orogenic wedges: Analog models and case studies. *GSA today*, 20(1), 4-10.
- Manatschal, G., & Bernoulli, D. (1999). Architecture and tectonic evolution of nonvolcanic margins: Present-day Galicia and ancient Adria. *Tectonics*, 18(6), 1099-1119.

- Manatschal, G., Froitzheim, N., Rubenach, M., & Turrin, B. (2001). The role of detachment faulting in the formation of an ocean-continent transition: insights from the Iberia Abyssal Plain. *Geological Society, London, Special Publications*, 187(1), 405-428.
- Manatschal, G., & Müntener, O. (2009). A type sequence across an ancient magma-poor ocean–continent transition: the example of the western Alpine Tethys ophiolites. *Tectonophysics*, 473(1-2), 4-19.
- Manzotti, P., Balleve, M., Zucali, M., Robyr, M., & Engi, M. (2014a). The tectonometamorphic evolution of the Sesia–Dent Blanche nappes (internal Western Alps): review and synthesis. *Swiss Journal of Geosciences*, 107(2-3), 309-336.
- Manzotti, P., Le Carlier De Veslud, C., Le Bayon, B., & Balleve, M. (2014b). Petro-structural map of the Money Unit (Gran Paradiso Massif, Valnontey valley, Western Alps). *Journal of Maps*, 10(2), 324-340.
- Marques, F., Cabral, F., Gerya, T., Zhu, G., & May, D. (2014). Subduction initiates at straight passive margins. *Geology*, 42(4), 331-334.
- Marques, F. O., Nikolaeva, K., Assumpção, M., Gerya, T. V., Bezerra, F. H., do Nascimento, A. F., & Ferreira, J. M. (2013). Testing the influence of far-field topographic forcing on subduction initiation at a passive margin. *Tectonophysics*, 608, 517-524.
- Marroni, M., Meneghini, F., & Pandolfi, L. (2017). A revised subduction inception model to explain the Late Cretaceous, double-vergent orogen in the precollisional western Tethys: Evidence from the Northern Apennines. *Tectonics*, 36(10), 2227-2249.
- Mart, Y., Aharonov, E., Mulugeta, G., Ryan, W., Tentler, T., & Goren, L. (2005). Analogue modelling of the initiation of subduction. *Geophysical Journal International*, 160(3), 1081-1091. doi:10.1111/j.1365-246X.2005.02544.x
- Martinod, J., & Davy, P. (1992). Periodic instabilities during compression or extension of the lithosphere 1. Deformation modes from an analytical perturbation method. *Journal of Geophysical Research: Solid Earth*, 97(B2), 1999-2014.
- Masini, E., Manatschal, G., & Mohn, G. (2013). The Alpine Tethys rifted margins: Reconciling old and new ideas to understand the stratigraphic architecture of magma-poor rifted margins. *Sedimentology*, 60(1), 174-196.
- Masson, D., Cartwright, J., Pinheiro, L., Whitmarsh, R., Beslier, M.-O., & Roeser, H. (1994). Compressional deformation at the ocean–continent transition in the NE Atlantic. *Journal of the Geological Society*, 151(4), 607-613.
- Maystrenko, Y., & Scheck-Wenderoth, M. (2009). Density contrasts in the upper mantle and lower crust across the continent–ocean transition: constraints from 3-D gravity modelling at the Norwegian margin. *Geophysical Journal International*, 179(1), 536-548.
- Maystrenko, Y. P., Scheck-Wenderoth, M., Hartwig, A., Anka, Z., Watts, A. B., Hirsch, K. K., & Fishwick, S. (2013). Structural features of the Southwest African continental margin according to results of lithosphere-scale 3D gravity and thermal modelling. *Tectonophysics*, 604, 104-121. doi:<https://doi.org/10.1016/j.tecto.2013.04.014>
- McAdoo, D. C., & Sandwell, D. T. (1985). Folding of oceanic lithosphere. *Journal of Geophysical Research: Solid Earth*, 90(B10), 8563-8569.

- McCarthy, A., Chelle-Michou, C., Müntener, O., Arculus, R., & Blundy, J. (2018). Subduction initiation without magmatism: The case of the missing Alpine magmatic arc. *Geology*, 46(12), 1059-1062.
- McCarthy, A., Tugend, J., Mohn, G., Candioti, L., Chelle-Michou, C., Arculus, R., Schmalholz, S. M., & Müntener, O. (2020). A case of Ampferer-type subduction and consequences for the Alps and the Pyrenees. *American Journal of Science*, 320(4), 313-372.
- McKenzie, D. (1977). The initiation of trenches: a finite amplitude instability. *Island arcs, deep sea trenches and back-arc basins*, 1, 57-61.
- Meffan-Main, S., Cliff, R., Barnicoat, A., Lombardo, B., & Compagnoni, R. (2004). A Tertiary age for Alpine high-pressure metamorphism in the Gran Paradiso massif, Western Alps: A Rb–Sr microsampling study. *Journal of Metamorphic Geology*, 22(4), 267-281.
- Midtkandal, I., Brun, J.-P., Gabrielsen, R. H., & Huismans, R. S. (2013). Control of lithosphere rheology on subduction polarity at initiation: Insights from 3D analogue modelling. *Earth and Planetary Science Letters*, 361, 219-228.
- Mikes, T., Christ, D., Petri, R., Dunkl, I., Frei, D., Báldi-Beke, M., Reitner, J., Wemmer, K., Hrvatović, H., & von Eynatten, H. (2008). Provenance of the bosnian Flysch. *Swiss Journal of Geosciences*, 101(1), 31-54.
- Milani, E., & Zalán, P. (1998). *Brazilian Geology Part I: The geology of Paleozoic cratonic basins and Mesozoic interior rifts of Brazil*. Paper presented at the 1998 AAPG International Conference and Exhibition, Rio de Janeiro, Brazil, Short Course Notes.
- Mixon, R. B., & Newell, W. L. (1977). Stafford fault system: Structures documenting Cretaceous and Tertiary deformation along the Fall Line in northeastern Virginia. *Geology*, 5(7), 437-440.
- Modenesi-Gauttieri, M. C., De Toledo, M. C. M., Hiruma, S. T., Taioli, F., & Shimada, H. (2011). Deep weathering and landscape evolution in a tropical plateau. *Catena*, 85(3), 221-230.
- Mohn, G., Manatschal, G., Beltrando, M., Masini, E., & Kuszniir, N. (2012). Necking of continental crust in magma-poor rifted margins: Evidence from the fossil Alpine Tethys margins. *Tectonics*, 31(1).
- Mohn, G., Manatschal, G., Masini, E., & Müntener, O. (2011). Rift-related inheritance in orogens: a case study from the Austroalpine nappes in Central Alps (SE-Switzerland and N-Italy). *International Journal of Earth Sciences*, 100(5), 937-961.
- Molnar, P., & Gray, D. (1979). Subduction of continental lithosphere: Some constraints and uncertainties. *Geology*, 7(1), 58-62.
- Monjoie, P., Bussy, F., Schaltegger, U., Mulch, A., Lapierre, H., & Pfeifer, H.-R. (2007). Contrasting magma types and timing of intrusion in the Permian layered mafic complex of Mont Collon (Western Alps, Valais, Switzerland): evidence from U/Pb zircon and 40 Ar/39 Ar amphibole dating. *Swiss Journal of Geosciences*, 100(1), 125-135.
- Monnereau, M., Rabinowicz, M., & Arquis, E. (1993). Mechanical erosion and reheating of the lithosphere: a numerical model for hotspot swells. *Journal of Geophysical Research: Solid Earth*, 98(B1), 809-823.

- Most, T. (2003). Geodynamic evolution of the Eastern Pelagonian Zone in northwestern Greece and the Republic of Macedonia. *Implications from U/Pb, Rb/Sr, K/Ar, Ar/Ar, geochronology and fission track thermochronology*. Tübingen, Phd, Germany, 1-170.
- Moyen, J.-F., Paquette, J.-L., Ionov, D., Gannoun, A., Korsakov, A., Golovin, A., & Moine, B. (2017). Paleoproterozoic rejuvenation and replacement of Archaean lithosphere: Evidence from zircon U–Pb dating and Hf isotopes in crustal xenoliths at Udachnaya, Siberian craton. *Earth and Planetary Science Letters*, 457, 149-159.
- Mueller, S., & Phillips, R. J. (1991). On The initiation of subduction. *Journal of Geophysical Research: Solid Earth*, 96(B1), 651-665. doi:<https://doi.org/10.1029/90JB02237>
- Müller, R. D., Sdrolias, M., Gaina, C., & Roest, W. R. (2008). Age, spreading rates, and spreading asymmetry of the world's ocean crust. *Geochemistry, Geophysics, Geosystems*, 9(4).
- Munch, J., Gerya, T., & Ueda, K. (2020). Oceanic crust recycling controlled by weakening at slab edges. *Nature communications*, 11(1), 1-6.
- Müntener, O., Hermann, J., & Trommsdorff, V. (2000). Cooling history and exhumation of lower-crustal granulite and upper mantle (Malenco, Eastern Central Alps). *Journal of Petrology*, 41(2), 175-200.
- Müntener, O., Pettke, T., Desmurs, L., Meier, M., & Schaltegger, U. (2004). Refertilization of mantle peridotite in embryonic ocean basins: trace element and Nd isotopic evidence and implications for crust–mantle relationships. *Earth and Planetary Science Letters*, 221(1-4), 293-308.
- Naliboff, J., Conrad, C., & Lithgow-Bertelloni, C. (2009). Modification of the lithospheric stress field by lateral variations in plate-mantle coupling. *Geophysical Research Letters*, 36(22).
- Nalpas, T., & Brun, J.-P. (1993). Salt flow and diapirism related to extension at crustal scale. *Tectonophysics*, 228(3-4), 349-362.
- Neto, H. C. L., Ferreira, J. M., Bezerra, F. H. R., Assumpção, M. S., do Nascimento, A. F., Sousa, M. O., & Menezes, E. A. (2013). Upper crustal earthquake swarms in São Caetano: Reactivation of the Pernambuco shear zone and trending branches in intraplate Brazil. *Tectonophysics*, 608, 804-811.
- Nikolaeva, K., Gerya, T. V., & Marques, F. O. (2010). Subduction initiation at passive margins: Numerical modeling. *Journal of Geophysical Research: Solid Earth*, 115(B3). doi: <https://doi.org/10.1029/2009JB006549>
- Nikolaeva, K., Gerya, T. V., & Marques, F. O. (2011). Numerical analysis of subduction initiation risk along the Atlantic American passive margins. *Geology*, 39(5), 463-466.
- Niu, Y., O'HARA, M. J., & Pearce, J. A. (2003). Initiation of subduction zones as a consequence of lateral compositional buoyancy contrast within the lithosphere: a petrological perspective. *Journal of Petrology*, 44(5), 851-866.
- Osmundsen, P. T., & Redfield, T. F. (2011). Crustal taper and topography at passive continental margins. *Terra Nova*, 23(6), 349-361. doi:<https://doi.org/10.1111/j.1365-3121.2011.01014.x>

- Oxburgh, E., & Parmentier, E. (1977). Compositional and density stratification in oceanic lithosphere-causes and consequences. *Journal of the Geological Society*, 133(4), 343-355.
- Oxburgh, E., & Turcotte, D. (1976). The physico-chemical behaviour of the descending lithosphere. *Tectonophysics*, 32(1-2), 107-128.
- Pamic, J. J. (1984). Triassic magmatism of the Dinarides in Yugoslavia. *Tectonophysics*, 109(3-4), 273-307.
- Parsons, B., & McKenzie, D. (1978). Mantle convection and the thermal structure of the plates. *Journal of Geophysical Research: Solid Earth*, 83(B9), 4485-4496.
- Pascal, C., & Cloetingh, S. A. (2009). Gravitational potential stresses and stress field of passive continental margins: Insights from the south-Norway shelf. *Earth and Planetary Science Letters*, 277(3-4), 464-473.
- Pe-Piper, G. (1998). The nature of Triassic extension-related magmatism in Greece: evidence from Nd and Pb isotope geochemistry. *Geological Magazine*, 135(3), 331-348.
- Pedoja, K., Husson, L., Regard, V., Cobbold, P. R., Ostanciaux, E., Johnson, M. E., Kershaw, S., Saillard, M., Martinod, J., & Furgerot, L. (2011). Relative sea-level fall since the last interglacial stage: are coasts uplifting worldwide? *Earth-Science Reviews*, 108(1-2), 1-15.
- Pedreira, D., Pulgar, J., Gallart, J., & Torné, M. (2007). Three-dimensional gravity and magnetic modeling of crustal indentation and wedging in the western Pyrenees-Cantabrian Mountains. *Journal of Geophysical Research: Solid Earth*, 112(B12).
- Pereira, R., Alves, T. M., & Cartwright, J. (2011). Post-rift compression on the SW Iberian margin (eastern North Atlantic): a case for prolonged inversion in the ocean-continent transition zone. *Journal of the Geological Society*, 168(6), 1249-1263.
- Péron-Pinvidic, G., & Manatschal, G. (2009). The final rifting evolution at deep magma-poor passive margins from Iberia-Newfoundland: a new point of view. *International Journal of Earth Sciences*, 98(7), 1581-1597.
- Photiades, A., Saccani, E., & Tassinari, R. (2003). Petrogenesis and tectonic setting of volcanic rocks from the Subpelagonian ophiolitic mélange in the Agorani area (Othrys, Greece). *Ophioliti*, 28(2), 121-135.
- Pinto, L. G. R., de Pádua, M. B., Ussami, N., Vitorello, Í., Padilha, A. L., & Braitenberg, C. (2010). Magnetotelluric deep soundings, gravity and geoid in the south São Francisco craton: Geophysical indicators of cratonic lithosphere rejuvenation and crustal underplating. *Earth and Planetary Science Letters*, 297(3-4), 423-434.
- Poh, J., Yamato, P., Duretz, T., Gapais, D., & Ledru, P. (2020). Precambrian deformation belts in compressive tectonic regimes: A numerical perspective. *Tectonophysics*, 777, 228350.
- Popov, A., & Sobolev, S. V. (2008). SLIM3D: A tool for three-dimensional thermomechanical modeling of lithospheric deformation with elasto-visco-plastic rheology. *Physics of the Earth and Planetary Interiors*, 171(1-4), 55-75.
- Quigley, M. C., Clark, D., & Sandiford, M. (2010). Tectonic geomorphology of Australia. *Geological Society, London, Special Publications*, 346(1), 243-265.

- Radulescu, I. G., Rubatto, D., Gregory, C., & Compagnoni, R. (2009). The age of HP metamorphism in the Gran Paradiso Massif, Western Alps: a petrological and geochronological study of “silvery micaschists”. *Lithos*, 110(1-4), 95-108.
- Ramberg, H. (1981). *Gravity, deformation and the earth's crust: in theory, experiments and geological application*: Academic press.
- Ranalli, G. (1995). *Rheology of the Earth*: Springer Science & Business Media.
- Räss, L., Duretz, T., Podladchikov, Y. Y., & Schmalholz, S. M. (2017). M2Di: Concise and efficient MATLAB 2-DS tokens solvers using the Finite Difference Method. *Geochemistry, Geophysics, Geosystems*, 18(2), 755-768.
- Regenauer-Lieb, K., Yuen, D. A., & Branlund, J. (2001). The initiation of subduction: criticality by addition of water? *Science*, 294(5542), 578-580.
- Regis, D., Rubatto, D., Darling, J., Cenki-Tok, B., Zucali, M., & Engi, M. (2014). Multiple metamorphic stages within an eclogite-facies terrane (Sesia Zone, Western Alps) revealed by Th–U–Pb petrochronology. *Journal of Petrology*, 55(7), 1429-1456.
- Reston, T., & Manatschal, G. (2011). Rifted margins: Building blocks of later collision. In *Arc-continent collision* (pp. 3-21): Springer.
- Ricard, Y., & Bercovici, D. (2009). A continuum theory of grain size evolution and damage. *Journal of Geophysical Research: Solid Earth*, 114(B1).
- Riccomini, C., Sant’Anna, L. G., & Ferrari, A. L. (2004). Evolução geológica do rift continental do sudeste do Brasil. *Geologia do continente Sul-Americano: evolução da obra de Fernando Flávio Marques de Almeida*, 383-405.
- Richardson, R. M., Solomon, S. C., & Sleep, N. H. (1979). Tectonic stress in the plates. *Reviews of Geophysics*, 17(5), 981-1019.
- Rioux, M., Garber, J., Bauer, A., Bowring, S., Searle, M., Kelemen, P., & Hacker, B. (2016). Synchronous formation of the metamorphic sole and igneous crust of the Semail ophiolite: New constraints on the tectonic evolution during ophiolite formation from high-precision U–Pb zircon geochronology. *Earth and Planetary Science Letters*, 451, 185-195.
- Ritter, M. C., Leever, K., Rosenau, M., & Oncken, O. (2016). Scaling the sandbox—Mechanical (dis) similarities of granular materials and brittle rock. *Journal of Geophysical Research: Solid Earth*, 121(9), 6863-6879.
- Roberts, D., & Bally, A. (2012). From rifts to passive margins: a continuum of extension. In *Regional Geology and Tectonics: Phanerozoic Rift Systems and Sedimentary Basins* (pp. 18-31): Elsevier.
- Robertson, A., Karamata, S., & Šarić, K. (2009). Overview of ophiolites and related units in the Late Palaeozoic–Early Cenozoic magmatic and tectonic development of Tethys in the northern part of the Balkan region. *Lithos*, 108(1-4), 1-36.
- Robertson, A., & Stillman, C. (1979). Late Mesozoic sedimentary rocks of Fuerteventura, Canary Islands: implications for West African continental margin evolution. *Journal of the Geological Society*, 136(1), 47-60.
- Rocha, M. P., de Azevedo, P. A., Marotta, G. S. A., Schimmel, M., & Fuck, R. (2016). Causes of intraplate seismicity in central Brazil from travel time seismic tomography. *Tectonophysics*, 680, 1-7.

- Rolf, T., & Tackley, P. (2011). Focussing of stress by continents in 3D spherical mantle convection with self-consistent plate tectonics. *Geophysical Research Letters*, 38(18).
- Rolland, Y., Galoyan, G., Bosch, D., Sosson, M., Corsini, M., Fornari, M., & Verati, C. (2009). Jurassic back-arc and Cretaceous hot-spot series In the Armenian ophiolites—Implications for the obduction process. *Lithos*, 112(3-4), 163-187.
- Roure, F., Casero, P., & Addoum, B. (2012). Alpine inversion of the North African margin and delamination of its continental lithosphere. *Tectonics*, 31(3). doi:<https://doi.org/10.1029/2011TC002989>
- Rowley, D. B., Forte, A. M., Rowan, C. J., Glišović, P., Moucha, R., Grand, S. P., & Simmons, N. A. (2016). Kinematics and dynamics of the East Pacific Rise linked to a stable, deep-mantle upwelling. *Science advances*, 2(12), e1601107.
- Rybacki, E., & Dresen, G. (2004). Deformation mechanism maps for feldspar rocks. *Tectonophysics*, 382(3-4), 173-187.
- Saccani, E., Beccaluva, L., Photiades, A., & Zeda, O. (2011). Petrogenesis and tectono-magmatic significance of basalts and mantle peridotites from the Albanian–Greek ophiolites and sub-ophiolitic mélanges. New constraints for the Triassic–Jurassic evolution of the Neo-Tethys in the Dinaride sector. *Lithos*, 124(3-4), 227-242.
- Saddiqi, O., Michard, A., Goffe, B., Poupeau, G., & Oberhänsli, R. (2006). Fission-track thermochronology of the Oman Mountains continental windows, and current problems of tectonic interpretation. *Bulletin de la Société géologique de France*, 177(3), 127-134.
- Santimano, T., & Pysklywec, R. (2020). The influence of lithospheric mantle scars and rheology on intraplate deformation and orogenesis: Insights from tectonic analog models. *Tectonics*, 39(5), e2019TC005841.
- Schaltegger, U., Ulianov, A., Müntener, O., Ovtcharova, M., Peytcheva, I., Vonlanthen, P., Vennemann, T., Antognini, M., & Girlanda, F. (2015). Megacrystic zircon with planar fractures in miaskite-type nepheline pegmatites formed at high pressures in the lower crust (Ivrea Zone, southern Alps, Switzerland). *American Mineralogist*, 100(1), 83-94.
- Schellart, W. P., & Strak, V. (2016). A review of analogue modelling of geodynamic processes: Approaches, scaling, materials and quantification, with an application to subduction experiments. *Journal of Geodynamics*, 100, 7-32.
- Scherreiks, R., Meléndez, G., BouDagher-Fadel, M., Fermeli, G., & Bosence, D. (2014). Stratigraphy and tectonics of a time-transgressive ophiolite obduction onto the eastern margin of the Pelagonian platform from Late Bathonian until Valanginian time, exemplified in northern Evvoia, Greece. *International Journal of Earth Sciences*, 103(8), 2191-2216.
- Schmalholz, S., & Duretz, T. (2015). Shear zone and nappe formation by thermal softening, related stress and temperature evolution, and application to the Alps. *Journal of Metamorphic Geology*, 33(8), 887-908.
- Schmalholz, S. M., & Duretz, T. (2017). Impact of grain size evolution on necking in calcite layers deforming by combined diffusion and dislocation creep. *Journal of Structural Geology*, 103, 37-56.

- Schmalholz, S. M., Duretz, T., Hetényi, G., & Medvedev, S. (2019). Distribution and magnitude of stress due to lateral variation of gravitational potential energy between Indian lowland and Tibetan plateau. *Geophysical Journal International*, 216(2), 1313-1333.
- Schmalholz, S. M., & Fletcher, R. C. (2011). The exponential flow law applied to necking and folding of a ductile layer. *Geophysical Journal International*, 184(1), 83-89.
- Schmalholz, S. M., Kaus, B. J., & Burg, J.-P. (2009). Stress-strength relationship in the lithosphere during continental collision. *Geology*, 37(9), 775-778.
- Schmid, S., Fügenschuh, B., Kissling, E., & Schuster, R. (2004a). *Three lithospheric transects across the Alps and their forelands*. Paper presented at the Swiss Geoscience Meeting, Lausanne.
- Schmid, S. M., Fügenschuh, B., Kissling, E., & Schuster, R. (2004b). Tectonic map and overall architecture of the Alpine orogen. *Eclogae Geologicae Helvetiae*, 97(1), 93-117.
- Schmid, S. M., Fügenschuh, B., Kounov, A., Maženco, L., Nievergelt, P., Oberhänsli, R., Pleuger, J., Schefer, S., Schuster, R., & Tomljenović, B. (2020). Tectonic units of the Alpine collision zone between Eastern Alps and western Turkey. *Gondwana Research*, 78, 308-374.
- Schubert, G., Turcotte, D. L., & Olson, P. (2001). *Mantle convection in the Earth and planets*: Cambridge University Press.
- Sengör, A. C., & Burke, K. (1978). Relative timing of rifting and volcanism on Earth and its tectonic implications. *Geophysical Research Letters*, 5(6), 419-421.
- Shemenda, A. I. (1992). Horizontal lithosphere compression and subduction: constraints provided by physical modeling. *Journal of Geophysical Research: Solid Earth*, 97(B7), 11097-11116.
- Shervais, J. W., Reagan, M., Haugen, E., Almeev, R. R., Pearce, J. A., Prytulak, J., Ryan, J. G., Whattam, S. A., Godard, M., & Chapman, T. (2019). Magmatic response to subduction initiation: Part 1. Fore-arc basalts of the Izu-Bonin arc from IODP Expedition 352. *Geochemistry, Geophysics, Geosystems*, 20(1), 314-338.
- Smit, J., Brun, J., & Sokoutis, D. (2003). Deformation of brittle-ductile thrust wedges in experiments and nature. *Journal of Geophysical Research: Solid Earth*, 108(B10).
- Sobolev, S. V., & Brown, M. (2019). Surface erosion events controlled the evolution of plate tectonics on Earth. *Nature*, 570(7759), 52-57.
- Sokoutis, D., Bonini, M., Medvedev, S., Boccaletti, M., Talbot, C. J., & Koyi, H. (2000). Indentation of a continent with a built-in thickness change: experiment and nature. *Tectonophysics*, 320(3-4), 243-270.
- Sokoutis, D., Burg, J.-P., Bonini, M., Corti, G., & Cloetingh, S. (2005). Lithospheric-scale structures from the perspective of analogue continental collision. *Tectonophysics*, 406(1-2), 1-15.
- Sokoutis, D., & Willingshofer, E. (2011). Decoupling during continental collision and intra-plate deformation. *Earth and Planetary Science Letters*, 305(3-4), 435-444.
- Solomatov, V. (2004). Initiation of subduction by small-scale convection. *Journal of Geophysical Research: Solid Earth*, 109(B1).
- Šoštarić, S. B., Palinkaš, A., Neubauer, F., Cvetković, V., Bernroider, M., & Genser, J. (2014). The origin and age of the metamorphic sole from the Rogozna Mts., Western

- Vardar Belt: New evidence for the one-ocean model for the Balkan ophiolites. *Lithos*, 192, 39-55.
- Stampfli, G. M., & Hochard, C. (2009). Plate tectonics of the Alpine realm. *Geological Society, London, Special Publications*, 327(1), 89-111.
- Stanton, N., Kusznir, N., Gordon, A., & Schmitt, R. (2019). Architecture and Tectono-magmatic evolution of the Campos Rifted Margin: Control of OCT structure by basement inheritance. *Marine and Petroleum Geology*, 100, 43-59.
- Starostenko, V., Janik, T., Lysynchuk, D., Środa, P., Czuba, W., Kolomiyets, K., Aleksandrowski, P., Gintov, O., Omelchenko, V., Komminaho, K., Guterch, A., Tiira, T., Gryn, D., Legostaeva, O., Thybo, H., & Tolkunov, A. (2013). Mesozoic(?) lithosphere-scale buckling of the East European Craton in southern Ukraine: DOBRE-4 deep seismic profile. *Geophysical Journal International*, 195(2), 740-766. doi:10.1093/gji/ggt292
- Stein, C. A., Cloetingh, S., & Wortel, R. (1989a). Seasat-derived gravity constraints on stress and deformation in the northeastern Indian Ocean. *Geophysical Research Letters*, 16(8), 823-826.
- Stein, C. A., & Stein, S. (1993). Constraints on Pacific midplate swells from global depth-age and heat flow-age models. *The Mesozoic Pacific: Geology, Tectonics, and Volcanism, Geophys. Monogr. Ser.*, 77, 53-76.
- Stein, S., Cloetingh, S., Sleep, N. H., & Wortel, R. (1989b). Passive margin earthquakes, stresses and rheology. In *Earthquakes at North-Atlantic Passive Margins: Neotectonics and Postglacial Rebound* (pp. 231-259): Springer.
- Stephenson, R., & Cloetingh, S. (1991). Some examples and mechanical aspects of continental lithospheric folding. *Tectonophysics*, 188(1-2), 27-37.
- Stern, R. J. (2004). Subduction initiation: spontaneous and induced. *Earth and Planetary Science Letters*, 226(3-4), 275-292.
- Stern, R. J., & Gerya, T. (2018). Subduction initiation in nature and models: A review. *Tectonophysics*, 746, 173-198.
- Stern, R. J., Reagan, M., Ishizuka, O., Ohara, Y., & Whattam, S. (2012). To understand subduction initiation, study forearc crust: To understand forearc crust, study ophiolites. *Lithosphere*, 4(6), 469-483.
- Strak, V., & Schellart, W. P. (2018). A subduction and mantle plume origin for Samoan volcanism. *Scientific reports*, 8(1), 1-12.
- Sutherland, R., King, P., & Wood, R. (2001). Tectonic evolution of Cretaceous rift basins in south-eastern Australia and New Zealand: implications for exploration risk assessment.
- Sutra, E., Manatschal, G., Mohn, G., & Unternehr, P. (2013). Quantification and restoration of extensional deformation along the Western Iberia and Newfoundland rifted margins. *Geochemistry, Geophysics, Geosystems*, 14(8), 2575-2597.
- Swedan, N. (2015). Ridge push engine of plate tectonics. *Geotectonics*, 49(4), 342-359.
- Teixell, A., Labaume, P., Ayarza, P., Espurt, N., de Saint Blanquat, M., & Lagabriele, Y. (2018). Crustal structure and evolution of the Pyrenean-Cantabrian belt: A review and new interpretations from recent concepts and data. *Tectonophysics*, 724, 146-170.
- Tesauro, M., Kaban, M. K., & Cloetingh, S. A. (2013). Global model for the lithospheric strength and effective elastic thickness. *Tectonophysics*, 602, 78-86.

- Thielicke, W., & Stamhuis, E. (2014). PIVlab—towards user-friendly, affordable and accurate digital particle image velocimetry in MATLAB. *Journal of open research software*, 2(1).
- Thielmann, M., & Kaus, B. J. (2012). Shear heating induced lithospheric-scale localization: Does it result in subduction? *Earth and Planetary Science Letters*, 359, 1-13.
- Thöni, M., & Miller, C. (2000). Permo-Triassic pegmatites in the eo-Alpine eclogite-facies Koralpe complex, Austria: age and magma source constraints from mineral chemical, Rb-Sr and Sm-Nd isotope data. *Schweizerische Mineralogische und Petrographische Mitteilungen*, 80(2), 169-186.
- Toth, J., & Gurnis, M. (1998). Dynamics of subduction initiation at preexisting fault zones. *Journal of Geophysical Research: Solid Earth*, 103(B8), 18053-18067.
- Trubelja, F., Burgath, K.-P., & Marchig, V. (2004). Triassic magmatism in the area of the Central Dinarides (Bosnia and Herzegovina): geochemical resolving of tectonic setting. *Geologia Croatica*, 57(2), 159-170.
- Tsikouras, B., Pe-Piper, G., Piper, D. J., & Hatzipanagiotou, K. (2008). Triassic rift-related komatiite, picrite and basalt, Pelagonian continental margin, Greece. *Lithos*, 104(1-4), 199-215.
- Tugend, J., Gillard, M., Manatschal, G., Nirrengarten, M., Harkin, C., Epin, M.-E., Sauter, D., Autin, J., Kuszniir, N., & Mcdermott, K. (2020). Reappraisal of the magma-rich versus magma-poor rifted margin archetypes. *Geological Society, London, Special Publications*, 476(1), 23-47.
- Turcotte, D., & Schubert, G. (2014). *Geodynamics*: Cambridge university press.
- Ueda, K., Gerya, T., & Sobolev, S. V. (2008). Subduction initiation by thermal–chemical plumes: numerical studies. *Physics of the Earth and Planetary Interiors*, 171(1-4), 296-312.
- Ulvrova, M. M., Coltice, N., Williams, S., & Tackley, P. J. (2019). Where does subduction initiate and cease? A global scale perspective. *Earth and Planetary Science Letters*, 528, 115836.
- Unternehr, P., Péron-Pinvidic, G., Manatschal, G., & Sutra, E. (2010). Hyper-extended crust in the South Atlantic: in search of a model. *Petroleum Geoscience*, 16(3), 207-215. doi:10.1144/1354-079309-904
- Ussami, N., Shiraiwa, S., & Dominguez, J. M. L. (1999). Basement reactivation in a sub-Andean foreland flexural bulge: The Pantanal wetland, SW Brazil. *Tectonics*, 18(1), 25-39.
- Ustaszewski, K., Schmid, S. M., Lugović, B., Schuster, R., Schaltegger, U., Bernoulli, D., Hottinger, L., Kounov, A., Fügenschuh, B., & Schefer, S. (2009). Late Cretaceous intra-oceanic magmatism in the internal Dinarides (northern Bosnia and Herzegovina): Implications for the collision of the Adriatic and European plates. *Lithos*, 108(1-4), 106-125.
- van de Lagemaat, S. H., Swart, M. L., Vaes, B., Kusters, M. E., Boschman, L. M., Burton-Johnson, A., Bijl, P. K., Spakman, W., & van Hinsbergen, D. J. (2021). Subduction initiation in the Scotia Sea region and opening of the Drake Passage: When and why? *Earth-Science Reviews*, 103551.
- Van der Meijde, M., Julià, J., & Assumpção, M. (2013). Gravity derived Moho for South America. *Tectonophysics*, 609, 456-467.

- van Hinsbergen, D. J., Maffione, M., Plunder, A., Kaymakçı, N., Ganerød, M., Hendriks, B. W., Corfu, F., Gürer, D., de Gelder, G. I., & Peters, K. (2016). Tectonic evolution and paleogeography of the Kırşehir Block and the Central Anatolian Ophiolites, Turkey. *Tectonics*, 35(4), 983-1014.
- van Hinsbergen, D. J., Torsvik, T. H., Schmid, S. M., Mañenco, L. C., Maffione, M., Vissers, R. L., Gürer, D., & Spakman, W. (2020). Orogenic architecture of the Mediterranean region and kinematic reconstruction of its tectonic evolution since the Triassic. *Gondwana Research*, 81, 79-229.
- van Hunen, J., & van den Berg, A. P. (2008). Plate tectonics on the early Earth: limitations imposed by strength and buoyancy of subducted lithosphere. *Lithos*, 103(1-2), 217-235.
- Vavra, G., Schmid, R., & Gebauer, D. (1999). Internal morphology, habit and U-Th-Pb microanalysis of amphibolite-to-granulite facies zircons: geochronology of the Ivrea Zone (Southern Alps). *Contributions to Mineralogy and Petrology*, 134(4), 380-404.
- Vergnolle, M., Pollitz, F., & Calais, E. (2003). Constraints on the viscosity of the continental crust and mantle from GPS measurements and postseismic deformation models in western Mongolia. *Journal of Geophysical Research: Solid Earth*, 108(B10).
- Viejo, G. F., & Gallastegui, J. (2005a). The ESCI-N Project after a decade: A síntesis of the results and open questions. *Trabajos de geología*, 25(25), 9-27.
- Viejo, G. F., & Gallastegui, J. (2005b). The ESCI-N Project after a decade: a synthesis of the results and open questions. *Trabajos de geología*(25), 9-27.
- Visonà, D., & Zanferrari, A. (2000). Some constraints on geochemical features in the Triassic mantle of the easternmost Austroalpine-Southalpine domain: evidence from the Karawanken pluton (Carinthia, Austria). *International Journal of Earth Sciences: Geologische Rundschau*, 89(1), 40.
- Vlaar, N., & Wortel, M. (1976). Lithospheric aging, instability and subduction. *Tectonophysics*, 32(3-4), 331-351.
- Wada, I., Wang, K., He, J., & Hyndman, R. D. (2008). Weakening of the subduction interface and its effects on surface heat flow, slab dehydration, and mantle wedge serpentinization. *Journal of Geophysical Research: Solid Earth*, 113(B4).
- Waldron, J. W., Schofield, D. I., Murphy, J. B., & Thomas, C. W. (2014). How was the Iapetus Ocean infected with subduction? *Geology*, 42(12), 1095-1098.
- Wan, B., Wu, F., Chen, L., Zhao, L., Liang, X., Xiao, W., & Zhu, R. (2019). Cyclical one-way continental rupture-drift in the Tethyan evolution: Subduction-driven plate tectonics. *Science China Earth Sciences*, 62(12), 2005-2016.
- Watts, A., & Burov, E. (2003). Lithospheric strength and its relationship to the elastic and seismogenic layer thickness. *Earth and Planetary Science Letters*, 213(1-2), 113-131.
- Watts, A., Zhong, S., & Hunter, J. (2013). The behavior of the lithosphere on seismic to geologic timescales. *Annual Review of Earth and Planetary Sciences*, 41, 443-468.
- Weertman, J. (1978). Creep laws for the mantle of the Earth. *Philosophical Transactions of the Royal Society of London. Series A, Mathematical and Physical Sciences*, 288(1350), 9-26.

- Weijermars, R., & Schmeling, H. (1986). Scaling of Newtonian and non-Newtonian fluid dynamics without inertia for quantitative modelling of rock flow due to gravity (including the concept of rheological similarity). *Physics of the Earth and Planetary Interiors*, 43(4), 316-330.
- Whattam, S. A., & Stern, R. J. (2015). Late Cretaceous plume-induced subduction initiation along the southern margin of the Caribbean and NW South America: The first documented example with implications for the onset of plate tectonics. *Gondwana Research*, 27(1), 38-63.
- White, R. S., Spence, G. D., Fowler, S. R., McKenzie, D. P., Westbrook, G. K., & Bowen, A. N. (1987). Magmatism at rifted continental margins. *Nature*, 330(6147), 439-444.
- Whittaker, J., Afonso, J., Masterton, S., Müller, R., Wessel, P., Williams, S., & Seton, M. (2015). Long-term interaction between mid-ocean ridges and mantle plumes. *Nature Geoscience*, 8(6), 479-483.
- Willett, S., Beaumont, C., & Fullsack, P. (1993). Mechanical model for the tectonics of doubly vergent compressional orogens. *Geology*, 21(4), 371-374.
- Willingshofer, E., Sokoutis, D., Beekman, F., Schönebeck, J.-M., Warsitzka, M., & Rosenau, M. (2018). Ring shear test data of feldspar sand and quartz sand used in the Tectonic Laboratory (TecLab) at Utrecht University for experimental Earth Science applications. *GFZ Data Services*, V. 1. *GFZ Data Services*. doi:<http://doi.org/10.5880/fidgeo.2018.072>
- Willingshofer, E., Sokoutis, D., Luth, S., Beekman, F., & Cloetingh, S. (2013). Subduction and deformation of the continental lithosphere in response to plate and crust-mantle coupling. *Geology*, 41(12), 1239-1242.
- Wilson, J. T. (1968). Static or mobile earth: the current scientific revolution. *Proceedings of the American Philosophical Society*, 112(5), 309-320.
- Wyssession, M. E., Okal, E. A., & Miller, K. L. (1991). Intraplate seismicity of the Pacific Basin, 1913–1988. *Pure and Applied Geophysics*, 135(2), 261-359.
- Yamato, P., Husson, L., Becker, T. W., & Pedoja, K. (2013). Passive margins getting squeezed in the mantle convection vice. *Tectonics*, 32(6), 1559-1570. doi: <https://doi.org/10.1002/2013TC003375>
- Yan, Q., & Shi, X. (2014). Petrologic perspectives on tectonic evolution of a nascent basin (Okinawa Trough) behind Ryukyu Arc: A review. *Acta Oceanologica Sinica*, 33(4), 1-12.
- Yumul Jr, G. P., Dimalanta, C. B., Tamayo Jr, R. A., & Maury, R. C. (2003). Collision, subduction and accretion events in the Philippines: A synthesis. *Island Arc*, 12(2), 77-91.
- Zhong, X., & Li, Z. H. (2019). Forced Subduction Initiation at Passive Continental Margins: Velocity-Driven Versus Stress-Driven. *Geophysical Research Letters*, 46(20), 11054-11064.
- Zhong, X., & Li, Z. H. (2020). Subduction Initiation During Collision-Induced Subduction Transference: Numerical Modeling and Implications for the Tethyan Evolution. *Journal of Geophysical Research: Solid Earth*, 125(2), e2019JB019288.
- Zhou, X., Li, Z.-H., Gerya, T. V., & Stern, R. J. (2020). Lateral propagation-induced subduction initiation at passive continental margins controlled by preexisting lithospheric weakness. *Science advances*, 6(10), eaaz1048.

- Ziegler, P. A., & Cloetingh, S. (2004). Dynamic processes controlling evolution of rifted basins. *Earth-Science Reviews*, 64(1-2), 1-50.
- Ziegler, P. A., Cloetingh, S., & van Wees, J.-D. (1995). Dynamics of intra-plate compressional deformation: the Alpine foreland and other examples. *Tectonophysics*, 252(1-4), 7-59.

About the author

Antoine Auzemery was born on the 12th of May 1991, in Limoges, France. After high school, he studied at the Engineer school UniLasalle Beauvais, and the University of Barcelona. In 2015, he earned his Master degree in Earth Sciences. In September 2016, he started his PhD project in the Tectonics Group at the Utrecht University.