Numerical Geodynamic Experiments of Continental Collision:
Past and Present

by

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Abstract

Research explores deep continental lithosphere (i.e., the continental lower crust and mantle lithosphere) deformation during continental collision.

I found that depending on the composition/rheology of the crust and the amount of radiogenic heat production in the crust, three dominant modes of mantle lithosphere deformation evolve under Neoarchean-like conditions: (1) a pure-shear thickening style; (2) an imbrication style; (3) and a “flat-subduction” style. The imbrication and the flat-subduction styles result in the emplacement of “plate-like” mantle lithosphere at depths between 200 km and 325 km. The imbrication style behavior shifts to the “flat-subduction” style behavior after a crustal inversion event.

I investigated mature Phanerozoic-style collision and found that it is sensitive to mantle lithosphere density, mantle lithosphere yield stress, lower-crustal strength and to the presence of phase change-related density changes in the lower crust. The early stages of collision are accommodated by subduction of lower crust and mantle lithosphere along a discrete shear zone beneath the overriding plate. Next, the subducting lower crust and mantle lithosphere retreat from the collision zone, permitting the sub-lithospheric mantle to upwell and intrude the overriding plate. Next, the lower crust and mantle lithosphere of the overriding plate delaminate from the overlying crust. This process produces plateau-
like uplift. These modeling results are interpreted in the context of available geological and geophysical observables for the Himalayan-Tibetan orogen.

I quantitatively investigated the effects that sediment deposition may have on continental lithosphere deformation during collision. In the absence of sedimentation, the early stages of collision are accommodated by subduction of lower crust and mantle lithosphere beneath the overriding plate. Next, the subducting lower crust and mantle lithosphere retreat from the collision zone. This permits the sub-lithospheric mantle to upwell and come into contact with the thickened upper crust. When sedimentation is imposed subduction-like consumption of the subducting plate remains stable.

Using numerical geodynamic models, I studied the influence of the pressure-dependence of viscosity on tectonic deformation during collision. At low activation volumes, high convergence rates, and low to moderate initial Moho temperatures the subduction style of mantle lithosphere deformation is dominant. At low activation volumes, high convergence rates, and high initial Moho temperatures distributed pure-shear style deformation occurs. At low activation volumes, low convergence rate, and moderate to high initial Moho temperatures the mantle lithosphere prefers a convective removal style of deformation. Increasing the activation volume of mantle material in either of these three cases changes the style of mantle lithosphere deformation because its viscosity increases non-linearly.
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For Kath and Gilles (the sidewinder sleeps tonight)
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Chapter 1

General introduction
1.1 Introduction

The theory of plate tectonics provides a kinematic framework for understanding the behavior of the rigid plates about the surface of the Earth and the dynamic evolution of the lithosphere at the plate boundaries (e.g., Wilson, 1965). Convergent plate boundaries comprise both subduction and collision zones. At subduction zones geological and geophysical observations suggest that convergence is accommodated by subduction of one plate (oceanic plate) beneath the other (continental or oceanic plate). At collisional plate boundaries the colliding plates are both continental in nature, or one is continental and the other carries a magmatic arc (reviewed in Schellart, 2010). The behavior of the deep continental lithosphere (i.e., the continental mantle lithosphere) during continental collision is difficult to determine because of the lack of seismic zones to delineate the plate at depth (e.g., as with a Wadati-Benioff zone) and the potential presence of a highly deformed thick continental crust can help obscure the nature of continental collision zones. Forward geodynamic numerical models have been used to attempt to gain insight into some of the possible tectonic processes that occur at depth during continental collision, investigating some of the fundamental styles of lithospheric deformation and/or thermal-mechanical parameters influencing these styles.

Numerical geodynamic models of continental collision suggest primary styles of deformation to accommodate shortening continental mantle lithosphere: 1) lithospheric “dripping”, where the thickened dense lower mantle lithosphere becomes gravitationally unstable and descends as a viscous Rayleigh-Taylor-gravitational instability (e.g., Houseman et al., 1981); 2) delamination, where the dense continental mantle lithosphere peels away from the overlying crust, directly exposing the crust to upwelling S.L.M. (e.g., Bird, 1979); or 3) continental subduction, where continental plate collision is accommodated by under-thrusting/subduction of a coherent, “plate-like” portion of one continental plate beneath the other along a discrete shear zone (e.g., Pysklywec et al., 2002) (Figure 1.1). Previous numerical and analogue geodynamic experiments have shown that Rayleigh-Taylor-type
instabilities of the lithosphere, lithosphere delamination and continental subduction are controlled by boundary conditions and the thermal, mechanical and rheological properties of the lithosphere (e.g., Beaumont et al., 1996; Billen and Houseman, 2004; Buck and Toksöz, 1983; Conrad and Molnar, 1997; Harig et al., 2008; Houseman and Molnar, 1997; Houseman et al., 2000; Lenardic and Kaula, 1995; Molnar et al., 1998; Molnar and Houseman, 2004; Morency and Doin, 2004; Neil and Houseman, 1999; Pysklywec et al., 2000; Pysklywec, 2001; Pysklywec et al., 2002; Pysklywec and Cruden, 2004; Pysklywec and Beaumont, 2004; Willett et al., 1993; Willett and Beaumont, 1994; Pysklywec et al., 2000; Pysklywec, 2001; Pysklywec et al., 2002). Despite this work, there is still debate about the fundamental style of lithospheric deformation at collision, yielding wide-ranging interpretations of tectonics at such plate boundaries. One example is the behavior of the continental mantle lithosphere when much shortening occurs (∼1800 km). Such shortening is characteristic of the Himalayan-Tibetan orogen (e.g., England and Houseman, 1986) where the fate of deep lithospheric material is not well constrained.

From a rheological perspective, it was recently shown that results for olivine determined at pressures below ∼0.5 GPa (e.g., Hirth and Kohlstedt (1996); Chopra and Paterson (1984)) cannot be assumed to be valid at pressures greater than several GPa because at these pressures the pressure-dependence of viscosity is large owing to the activation volume (e.g., Karato, 2010; Kawazoe et al., 2009). That said, recent technological developments (e.g., Kawazoe et al., 2009) have permitted experimentalists to accurately measure the rheological parameters of olivine to several GPa. This now enables geodynamic modelling to include this rheological factor and assess its significance in controlling the dynamics of plate boundary evolution.

The density of the continental mantle lithosphere is also an important physical parameter in controlling the behavior of the continental lithosphere during collision. For example, an important assumption of continental mantle lithosphere dripping and delamination is that the mantle lithosphere is denser than the underlying S.L.M. Recent geochemical
Figure 1.1: Conceptual illustration of the a) Rayleigh-Taylor gravitational instability, b) delamination and c) subduction styles of continental lithosphere deformation during collision. Frames a) and c) are modified from Pysklywec et al. (2002).
studies (e.g., Poudjom Djomani et al., 2001; O’Reilly et al., 2001) suggest that continental mantle lithosphere density has undergone a secular increase since the Archean. Poudjom Djomani et al. (2001) showed that since the Archean, continental mantle lithosphere has become progressively less depleted (i.e., more fertile) in terms of Al and Ca contents and in Mg# and Fe/Al. They also showed that continental mantle lithosphere density has increased from Archean (3310 kg/m$^3$), to Proterozoic (3330 kg/m$^3$), to Phanerozoic times (3360 kg/m$^3$). It is important to note that these studies imply that, given identical reference temperatures, Phanerozoic continental mantle lithosphere is less dense than the underlying S.L.M. (3390 kg/m$^3$). Given the temperature-dependence of density, the greater density comes from lower temperatures in the upper portion of the continental mantle lithosphere (e.g., Poudjom Djomani et al., 2001). That said, it is not unreasonable to assume that, under certain conditions, continental mantle lithosphere may be more dense than the S.L.M. For example, this may occur by melt intrusion and freezing (e.g., Elkins-Tanton and Hager, 2000; Jull and Kelemen, 2001). Jull and Kelemen (2001) showed that the densities of lower crustal compositions resulting from arc magmatism (e.g., olivine pyroxenite) may exceed that of the S.L.M. by 50-250 kg/m$^3$. Given that olivine pyroxenites are also found as cratonic and non-cratic mantle xenoliths it is not unreasonable to assume that, under certain conditions, continental mantle lithosphere may be denser than the S.L.M. Elkins-Tanton and Hager (2000) showed that melt intruding the continental mantle lithosphere as dikes, above an asthenospheric upwelling, may solidify as eclogite. Consequently, the density of the continental mantle lithosphere may rise to 3340 kg/m$^3$ while that of the asthenosphere may be lowered to 3280 kg/m$^3$. That said, how the evolution of mantle lithosphere density through geologic time has affected the dynamics of deep continental lithosphere deformation during plate collision has not been closely studied.

Geodynamic models of collision also show that the behavior of the continental lithosphere may be strongly influenced by surface processes (i.e., erosion and sedimentation) (e.g., Beaumont et al., 1992; Willett, 1999a; Batt and Braun, 1997; Koons, 1990; Simpson,
The concepts of climate-tectonic interactions derived from previous numerical modelling efforts have been used to explain various enigmatic features of, specifically, collisional zones. One example of this is in the Southern Alps of New Zealand, where the asymmetric elevation profile of the orogen may be explained by the dominance of tectonic advection over fluvial incision in controlling topography Willett (1999a). As another example, Willett et al. (2006) suggest that the shift from orogenic construction to orogenic destruction of the European Alps at the end of the Miocene may be a response to a climate-driven increase in erosional flux. In this example the increase in erosional flux, reflected by an increase in sediment yield, is ascribed to a climatic shift to wetter conditions, possibly related to dessication of the Mediterranean Sea. Though most numerical geodynamic experiments of continental collision include surface erosion and sediment deposition (e.g., Burov and Toussaint, 2007; Yamato et al., 2008), how sediment deposition may modify the evolution of the continental lithosphere during plate collision is not well understood. As the conjugate to surface erosion, it may be expected that deposition of the removed mass may similarly influence the lithospheric tectonics.

The overall objective of this thesis is to investigate the styles of deep continental lithosphere (i.e., lower crust and mantle lithosphere) deformation during continental collision in both inter- and intra-plate settings using forward numerical geodynamic experiments. Specific studies related to the research themes outlined above were undertaken as separate projects written and published as individual journal articles. The thesis comprises six chapters.
1.2 Thesis structure

1.2.1 Chapter 2

In Chapter 2 I investigate the nature of continental lithosphere during collision under Neoarchean-like conditions as a potential process for creating thick SCLM (reviewed in Carlson et al., 2005), specifically considering the dynamical interaction between the continental crust and mantle. We conduct a series of computational geodynamic experiments that test the sensitivity of continental mantle lithosphere deformation to a buoyant mantle lithosphere, varying crustal compositions, reflective of crustal composition variability in Archean cratons, and degrees of radiogenic heat production (RHP) in the crust. As we demonstrate, tectonic behavior is significantly modified from present-day plate behavior owing to different thermal/compositional conditions during this time.

1.2.2 Chapter 3

In Chapter 3 I build on previous work to explore the styles of deep continental lithosphere deformation during mature collision. In particular, we investigate the development of deep continental lithosphere subduction, retreat and delamination during the collision process. A series of numerical geodynamic experiments are presented that test the sensitivity of mature collision (e.g., 1800 km of convergence) to varying mantle lithosphere density, mantle lithosphere yield stress and to the presence of phase change-related density changes in the lower crust (i.e., eclogitization of mafic lower crust). As a preliminary case study, the numerical results are considered in the context of the ~50 Ma Himalayan-Tibetan orogen where the process of deformation in the deep lithosphere is poorly understood (e.g., Willett and Beaumont, 1994; Houseman et al., 1981).
1.2.3 Chapter 4

Despite the presence of erosion/sedimentation in most numerical models of continental collision (e.g., Pysklywec, 2006; Burov and Toussaint, 2007), how, specifically, sediment deposition may modify the evolution of the continental lithosphere during plate collision is not well understood. As the conjugate to surface erosion, it may be expected that deposition of the removed mass may similarly influence the lithospheric tectonics. In Chapter 4 I quantitatively investigate this process using thermal-mechanical numerical experiments of the coupled processes of tectonic deformation and crustal mass flux. A free surface, prescribed erosional laws (e.g., empirically derived relief-dependent erosion) and sediment deposition dependent on the amount of material eroded make up the top boundary of the model domain and allow topography to develop self consistently with the underlying geodynamics. We conduct a series of experiments with varying convergence rates, initial MOHO temperatures and degrees of sediment deposition to show how the inclusion of the effects of sediment deposition produce a modified style of continental plate behavior during collision.

1.2.4 Chapter 5

Chapter 5 examines how the pressure-dependence of viscosity influences the style of mantle lithosphere deformation during continental collision. A series of numerical geodynamic experiments are presented that test the sensitivity of mantle lithosphere deformation to varying activation volumes of mantle material, convergence rates and initial Moho temperatures. Depending on where a numerical experiment is located in activation volume-convergence rate-initial Moho temperature space a range of mantle lithosphere deformation styles may develop (e.g., subduction, subduction-drip, ablative, ablative-drip, pure-shear thickening and convection removal).
1.3 Statement of authorship

Chapter 2 is co-authored by R. N. Pysklywec. R. Gray is the first author. R. Gray ran the thermal-mechanical numerical experiments, wrote the MATLAB algorithms for analyses, evaluated/plotted the data and wrote the paper. R. N Pysklywec revised the earlier versions of the manuscript. The interpretations of model results were discussed between R. Gray and R. N. Pysklywec. The paper is published as “Geodynamic models of Archean continental collision and the formation of mantle lithosphere keels” in Geophysical Research Letters, VOL. 37, L19301, 2010, doi:10.1029/2010GL043965. This study contributed to a small, but growing body of knowledge regarding lithospheric dynamics under Archean-like conditions. I believe this study has provided new insights into the poorly understood dynamical behavior of the Archean lithosphere during continental collision. More specifically, it has provided and alternative hypothesis to explain the anomalously thick mantle lithosphere beneath Archean terranes, e.g., the Superior Province of Canada.

Chapter 3 is co-authored by R. N. Pysklywec. R. Gray is the first author. R. Gray ran the thermal-mechanical numerical experiments, modified the source code to include mineralogical phase changes, wrote the MATLAB algorithms for analyses, evaluated/plotted the data and wrote the paper. R. N Pysklywec revised the earlier versions of the manuscript. The interpretations of model results were discussed between R. Gray and R. N. Pysklywec. The paper is published as “Geodynamic models of mature continental collision: Evolution of an orogen from lithospheric subduction to continental retreat/delamination” in Journal of Geophysical Research-Solid Earth, VOL. 117, B03408, 2012, doi:10.1029/2011JB008692. In this study I built on previous work to explore the styles of continental lithosphere deformation during mature collision where ~1800 km of convergence has been accommodated by horizontal shortening. I was the first to conduct a suite of numerical geodynamic experiments that tested the sensitivity of mature continental collision to varying mantle lithosphere density, mantle lithosphere yield stress, lower crustal strength and to the presence of phase change-related density changes in the lower
crust (i.e., eclogitization of mafic lower crust).

Chapter 4 is co-authored by R. N. Pysklywec. R. Gray is the first author. R. Gray ran the thermal-mechanical numerical experiments, modified the source code to include sedimentation, wrote the MATLAB algorithms for analyses, evaluated/plotted the data and wrote the paper. R. N Pysklywec revised the earlier versions of the manuscript. The interpretations of model results were discussed between R. Gray and R. N. Pysklywec. The paper is published as “Influence of sediment deposition on deep lithospheric tectonics” in Geophysical Research Letters, VOL. 39, L11312, 2012, doi:10.1029/2012GL051947. In this contribution I was the first to consider the effects that sedimentation may have on the style of deep continental lithosphere deformation during collision, and proposed that the presence of sediment deposition in numerical models of continental collision may lead to a greater degree of coupling between the colliding plates. The results of the study demonstrated the first quantitative insights into the feedback between surface deposition and deep lithospheric tectonics.

Chapter 5 is co-authored by R. N. Pysklywec. R. Gray is the first author. R. Gray ran the thermal-mechanical numerical experiments, modified the source code to include the pressure-dependence of viscosity, wrote the MATLAB algorithms for analyses, evaluated/plotted the data and wrote the paper. R. N Pysklywec revised the earlier versions of the manuscript. The interpretations of model results were discussed between R. Gray and R. N. Pysklywec. The paper is under review as “Influence of viscosity pressure-dependence on deep lithospheric tectonics during continental collision” in Journal of Geophysical Research- Solid Earth (manuscript #: 2012JB010001). Building on previous work, in this contribution I was the first to quantitatively investigate the influence of the pressure-dependence of viscosity on the process of deep lithosphere deformation during collision. The models demonstrate how the inclusion of viscosity pressure-dependence substantially alters the style of continental mantle lithosphere deformation.
Chapter 2

Geodynamic models of Archean continental collision and the formation of mantle lithosphere keels
Abstract

The processes responsible for the formation of thick, strong and cold Archean sub-continental lithospheric mantle (mantle keels) beneath Archean cratons remain elusive. Here, the dynamics of some such processes are studied by forward numerical modeling of the thermo-mechanical evolution of continental lithosphere undergoing collision and orogenesis under Neoarchean-like conditions. The numerical experiments illustrate that depending on the composition of the crust and the degree of radioactive heat production (R.H.P.) in the crust, three dominant modes of mantle lithosphere deformation evolve: (1) pure-shear thickening; (2) imbrication; (3) and a mode best described as underplating. All three modes of deformation result in the thickening and emplacement of plate-like mantle lithosphere to depths between 200 km and 350 km. The transition from pure-shear thickening to imbrication is largely dependent on the degree of R.H.P. in the crust, while the transition from the imbrication style to the underplating style is dependent on the composition of the lower crust.

2.1 Introduction

Archean sub-continental lithospheric mantle (SCLM) is composed of highly refractory peridotitic mantle that exists as a keel of thickened lithosphere beneath most Archean cratons (Carlson et al., 2005). Models of SCLM formation are poorly resolved, but generally fall into two categories (reviewed in Lee, 2006). The plume model suggests that partial melting in a plume head is the mechanism responsible for the formation of Archean SCLM. However, this hypothesis is problematic because the protoliths of cratonic peridotites underwent partial melting at pressures less than their equilibration pressures (Lee, 2006). In the second model, imbrication of buoyant oceanic lithosphere during lateral tectonic accretion is responsible for forming SCLM (e.g., Helmstaedt and Schulze, 1986). However, such a lateral tectonic accretion model implies a greater amount of eclogite in
SCLM than is implied by mantle-derived xenoliths (Schulze, 1989). A third, though less popular, model envisages SCLM as a result of continental collision (e.g., Jordan, 1978). However, previous studies on Archean continental collision have focused largely on lateral gravitationally-driven flow of lithosphere in response to a warmer continental geotherm and a buoyant sub-continental lithospheric mantle (e.g., Rey and Houseman, 2006) and the dynamic interaction between the mantle and lithosphere (C. M. Cooper and Moresi, 2006).

In this study, we investigate the nature of continental lithosphere during collision under Neoarchean-like (2.8-2.5 Ga) conditions as a potential process for creating thick SCLM, specifically considering the dynamical interaction between the continental crust and mantle. We conduct a series of computational geodynamic experiments that test the sensitivity of continental mantle lithosphere deformation to a buoyant mantle lithosphere; varying crustal compositions, reflective of crustal composition variability in Archean cratons, and degrees of radiogenic heat production (R.H.P.) in the crust. As we demonstrate, these tectonics are significantly modified from present-day plate behavior owing to different thermal/compositional conditions during this time. We do not preclude the possibility of other processes (e.g., plume-related), but rather focus on horizontal tectonics as the most widely cited.

2.2 Modeling Neoarchean continental collision

In the numerical experiments we used SOPALE, a plane-strain visco-plastic finite-element code (Fullsack, 1995). The experiments model an idealized collision between continental Neoarchean plates (Fig. 2.1). Plate velocities during the Archean are very unconstrained. For example, van Hunen and van den Berg (2008) showed that subduction of Archean oceanic lithosphere occurred at rates of ∼2-60 cm a⁻¹, whereas Korenaga (2006) showed that Archean plate tectonics was more sluggish than today by a factor of ∼2. Given this apparent contradiction and that the purpose of this chapter is to investigate mantle
lithosphere deformation during collision with modification only to crustal stratification and
R.H.P. I decided to model Neoarchean continental collision at a “moderate” rate of 3 cm a\(^{-1}\). Several aspects of the configuration tailor the model to Neoarchean conditions: (1) the initial geotherm in the experiments is significantly elevated from present-day estimates; (2) the internal heat production in the crust is based on the interpreted thermal state of Archean cratons \(\sim 2.6\) Ga (e.g., Pollack, 1997)); (3) the crustal configurations are chosen based on variability that exists in Archean cratons (Arndt, 1999; Musacchio et al., 2004); and (4) the densities of the mantle lithosphere and the sub-lithospheric mantle are based on mantle xenolith/xenocryst studies that have placed constraints on the secular variation in the density of the SCLM (Poudjom Djomani et al., 2001). The more depleted mantle lithosphere in the Archean and assumed paleo-geotherms lead to more buoyant cratonic mantle lithosphere than the underlying mantle. For the experiments, we vary the first three factors to consider how they control the behavior of the convergent mantle lithosphere.

**Figure 2.1:** Physical properties and initial configuration of numerical experiments. Continental convergence is modeled by introducing new lithosphere at the right margin of the box with velocity \(v_c = 3\) cm a\(^{-1}\). Left margin of the lithosphere is held fixed while outward flux \(v_o = 0.5\) cm a\(^{-1}\) is distributed evenly along sides of the sub-lithospheric mantle to balance the injected lithosphere. A weak zone \(\eta_{\text{eff.}} = 5 \times 10^{19}\) Pa-s is inserted in the upper mantle lithosphere to seed the onset of deformation. Its location can be interpreted as the plate boundary the moment collision begins; the assumption is that heterogeneities and weak zones can exist at plate boundaries. Density is a function of temperature and composition (using \(\alpha = 3.0 \times 10^{-5}\) and \(2.8 \times 10^{-5}\) K\(^{-1}\) for crust and mantle materials, respectively). Rheological parameters for crust and mantle materials are listed in Table 2.1.
**Table 2.1:** Physical parameters for reference experiment

<table>
<thead>
<tr>
<th>Mechanical Parameters</th>
<th>Upper crust</th>
<th>Lower crust</th>
<th>Mantle lithosphere</th>
<th>Sub-lithospheric mantle</th>
<th>Weak zone</th>
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<tr>
<td>Ref. density (kg m(^{-3}))</td>
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<td>3390</td>
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<td>(\phi_1) (deg.)</td>
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<td>15</td>
<td>15</td>
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<td>-</td>
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<tr>
<td>(\phi_2) (deg.)</td>
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<td>2</td>
<td>-</td>
<td>-</td>
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<tr>
<td>((I'<em>2)</em>{1/2})</td>
<td>1.5</td>
<td>1.5</td>
<td>1.5</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>((I'<em>2)</em>{1/2})</td>
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<td>0.5</td>
<td>0.5</td>
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<td>-</td>
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<td>Cohesion (MPa)</td>
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<td>-</td>
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<td>Wet</td>
<td>Wet olivine</td>
<td>Wet olivine</td>
<td>(\eta = 8 \times 10^{19}) Pa(\cdot)s</td>
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<tr>
<td></td>
<td>quartzite/</td>
<td>quartzite/</td>
<td>Diabase/</td>
<td>Felsic</td>
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<td></td>
<td>Dry</td>
<td>Granulite/</td>
<td>granite</td>
<td>Mafic</td>
<td></td>
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<tr>
<td>(f)</td>
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<td>1</td>
<td>140</td>
<td>1</td>
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<td>(2.00951 \times 10^{-21}/-/-/-)</td>
<td>(8.8334 \times 10^{-22}/-/-/-)</td>
<td>(2.00951 \times 10^{-21}/-/-/-)</td>
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</tr>
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<td>485/243/445</td>
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<td>515</td>
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</tr>
<tr>
<td>Heat capacity (J kg(^{-1}) K(^{-1}))</td>
<td>750</td>
<td>750</td>
<td>750</td>
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</tr>
<tr>
<td>Thermal conductivity (W m(^{-1}) K(^{-1}))</td>
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<td>2.25</td>
<td>2.25</td>
<td>2.25</td>
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<td>Thermal expansion (K(^{-1}))</td>
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<td>(3.0 \times 10^{-5})</td>
<td>(2.8 \times 10^{-5})</td>
<td>(2.8 \times 10^{-5})</td>
<td>(3.0 \times 10^{-5})</td>
</tr>
</tbody>
</table>

\(^a\) 1) Gleason and Tullis (1995); 2) Ranalli (1997); 3) Hirth and Kohlstedt (1996)
2.3 Experimental Results

Figure 2.2a shows results of Run1 after 111 Myr of imposed convergence, corresponding to 3330 km (i.e., 70%) of lithospheric shortening. Precambrian accretionary orogens typically underwent ∼50% (∼80 Myr) shortening (reviewed in Chardon et al., 2009). These experiments are not intended to recreate mantle lithosphere structures beneath Archean cratons, but to investigate the nature of Neoarchean continental lithosphere during collision; the results presented in Figures 2, 3 and 4 are still valid when 50% shortening has occurred. At this stage, five shear zones have developed in the mantle lithosphere (labeled sequentially in the order of their development). Four of the shear zones (SZ1,3-5) dip towards the right margin of the box, while SZ2 dips towards the left. The buoyancy of the mantle lithosphere compared to the underlying sub-lithospheric mantle, is preventing it from subducting into the sub-lithospheric mantle as shortening progresses and allowing it to thicken significantly. Rather, imposed convergence is causing the mantle lithosphere to imbricate. Plate-like mantle lithosphere (log(η_{eff.}) ≥ 23) is present at depths of ∼200 - 390 km below all shear zones. Some of the shortening has been accommodated in the mantle lithosphere by pure-shear thickening (e.g., to the left of SZ4) and folding (e.g., to the right of SZ1), but this is minor compared to the imbrication. Imposed convergence is dominantly accommodated in the crust by pure shear-thickening. The disjointed grid lines in the sub-lithospheric mantle are the result of extensive mantle convection-driven deformation of the Lagrangian grid.

Except for the composition of its crust, the set-up of experiment Run2 is identical to Run1. In Run2, a diabase rheology (Ranalli, 1997) (to simulate wet mafic volcanics) is used for the upper-crust and a felsic granulite rheology (Ranalli, 1997) is used for the lower crust. The motivation for this crustal set-up is to replicate Neoarchean crust prior to a possible crustal inversion event (e.g., the ∼2.65 Ga Late Archean Magmatic Event (Ahmad, 2009)). Figure 2.2b shows the results of the numerical experiment after 111 Myr of imposed convergence. Run2 shows a similar imbricating mantle lithosphere to
Run1 - even the relative timing of the shear zones is the same - but the vergence of the structures is more uniform in Run2. However, the vergence of the shear zones (e.g., the opposite polarity of SZ2 in Run1) may be a rather arbitrary event in the model where small-scale perturbations lead to high strain in the strain-softening materials. The wet quartzite upper-crust in Run1 is weaker than the wet diabase upper-crust in Run2, and as a result, imposed convergence in Run2 is accommodated in the upper crust by pure-shear thickening as well as folding. The difference suggests that the crust in Run1 is at least partially coupled to the mantle lithosphere whereas there is a more complete decoupling in Run 2, as evidenced by the highly deformed Lagrangian grid. Below SZ1-5, plate-like mantle lithosphere is present at depths of ~200 to ~390 km.

Experiment Run3 is identical to Run1, except now a granite rheology (Ranalli, 1997) is used for the upper crust and a mafic granulite rheology (Ranalli, 1997) is used for the lower crust. The motivation for this crustal set-up is to replicate Neoarchean crust after a possible crustal inversion event. After 111 Myr of imposed convergence, pro-side (Fig. 2.1) mantle lithosphere has slid beneath retro-side mantle lithosphere along SZ1 for a distance of ~950 km (Figure 2.2c). This style of deformation is best described as “underplating”. This behavior occurs because the mafic granulite lower crust is stronger than either a wet quartzite or a felsic granulite lower crust. Consequently, there is a greater degree of coupling between the lower crust and mantle lithosphere, resulting in a stronger lithosphere. As a consequence of this enhanced coupling, localized thrusting and exhumation of pro-side mafic granulite lower crust onto adjacent retro-side mafic granulite lower crust has occurred above SZ1. Plate-like mantle lithosphere is present at depths of ~350 km.

Experiment Run4 (Fig. 2.3a) is identical to the reference model Run1, but R.H.P. in the crust is increased to $6.76 \times 10^{-10}$ W kg$^{-1}$. Initially, imposed convergence is accommodated in the mantle lithosphere by the development of a shear zone (SZ1). Increased R.H.P. in the crust results in increased lower crustal temperatures, which in turn warms
Figure 2.2: Evolution of the thermo-mechanical models with varying crustal compositions and R.H.P. = 4.2 × 10^{-10} W kg^{-1} (1.7x that of today, (Mareschal and Jaupart, 2006)). We assign all the heat-producing elements to the felsic portions of the crust. Inset frames show filled contours of plate-like (log(\eta_{eff.}) ≥ 23) material. Material with log(\eta_{eff.}) < 23 is omitted from the plot (white regions). Viscous flow law of \dot{\varepsilon} = A\sigma^n \exp(-Q/RT) is used, where \dot{\varepsilon} is strain rate, \sigma is differential stress and T is temperature. Variables A, n and Q are material parameter, power exponent and activation energy, respectively. Due to computational limitations, viscosity range of 5 × 10^{19} to 10^{27} Pa-s is imposed in experiments, thus ignoring the effects of partial melting.
the mantle lithosphere. Higher mantle lithosphere temperatures lower its effective viscosity. Consequently, after the development of SZ1, imposed convergence is no longer accommodated in the mantle lithosphere by the development of shear zones but by pure-shear thickening and minor folding. Plate-like mantle lithosphere extends to depths of $\sim 200-300$ km. As in Run1, the overlying crust accommodates imposed convergence by undergoing pure-shear thickening.

Experiment Run5 (Fig. 2.3b) is identical to Run2, but the R.H.P. in the felsic granulite lower crust is increased to $6.76 \times 10^{-10}$ W kg$^{-1}$. The main difference between the models is that in Run5 the shear zones do not develop as readily as they did in Run2. For example, there are fewer shear zones in Run5 and these are not as well developed as in Run2 (viz., SZ4). This is owing to the increased R.H.P. in the lower crust that causes higher mantle lithosphere temperatures and a shift to more ductile deformation, mostly in the form of pure-shear thickening combined with the imbrication behavior. As in Run2, plate-like mantle lithosphere extends to depths between $\sim 200$ and 350 km. Similarly, the crust is strongly decoupled from the mantle lithosphere and undergoes folding and pure-shear thickening.

The setup of experiment Run6 (Fig. 2.3c) is identical to Run3, but the R.H.P. in the granite upper crust is increased to $6.76 \times 10^{-10}$ W kg$^{-1}$. The most notable difference between Run3 and Run6 is the presence of a second shear zone (SZ2) that has begun to form, of opposite polarity from SZ1. Below SZ1-2, plate-like mantle lithosphere extends to depths greater than $\sim 400$ km. Again, the crust is well coupled to the mantle lithosphere; leading to localized deformation of the crust at the mantle shear zones and exhumation of the lower crustal layer to the surface.

The results shown represent a small subset of a series of numerical experiments. When the log of effective viscosity of the lower crust (calculated near the retro-margin (Fig. 2.1) of the box) is plotted versus the degree of R.H.P. in the crust for each of the experiments, the modes of mantle lithosphere behavior group in separate domains (Fig. 2.4). The
Figure 2.3: Evolution of the thermo-mechanical models with wet quartzite crust (A), diabase upper crust and felsic granulite lower crust (B) and granite upper crust with mafic granulite lower crust (C) when R.H.P. = $6.76 \times 10^{-10} \text{ W kg}^{-1}$ (2.7x that of today, (Mareschal and Jaupart, 2006)). Otherwise model set-up is as experiments described for Figure 2.2. Inset frames show filled contours of plate-like ($\log(\eta_{\text{eff.}}) \geq 23$) material. As in Fig. 2.2, material with $\log(\eta_{\text{eff.}}) < 23$ is omitted from the plot (white regions).
pure-shear thickening style occurs when the degree of R.H.P. in the crust is sufficiently elevated to weaken the lower crust. This style transitions to the imbrication style as the degree of R.H.P. is lowered, allowing the lower crust to strengthen and more effectively couple to the mantle lithosphere than in the pure-shear thickening style. This transition is in contrast to the one that marks the shift from the imbrication style to the underplating style. This transition is less dependent on the degree of R.H.P. in the crust than on the composition of the lower crust. Transitioning from a felsic granulite to a mafic granulite lower crust (possibly accomplished by vertical crustal tectonics) results in a stronger lower crust. This leads to a greater degree of coupling between the lower crust and the mantle lithosphere, in turn increasing the strength of the mantle lithosphere and lithosphere.

![Figure 2.4: Illustration of the relationship between the three mantle lithosphere deformation styles (i.e., pure-shear thickening, imbrication and underplating) identified in the suite of numerical experiments carried out for this study.](image)

2.4 Conclusions and discussion

We tested how the behavior of mantle lithosphere changed with modification only to the composition of the lower crust and the degree of R.H.P. in the crust. Though we recognize
that other factors may be important (e.g., varying chemical buoyancy) we consider various
first-order interpretations of the state of the Archean lithosphere (i.e., the degree of R.H.P.
in the crust and the configuration of the crust). The numerical experiments identify three
dominant styles of model Neoarchean-like mantle lithosphere deformation during collision:
(1) dominantly pure-shear thickening (Figure 2.3a); (2) imbrication (Figure 2.2a, 2.2b,
2.3b); (3) and underplating with localized exhumation of the crust (Figure 2.2c, 2.3c).

The pure-shear-thickening style of deformation occurs when the temperature of the
lower crust is sufficiently raised by R.H.P.. This leads to a decrease in the degree of coupling
between the lower crust and the mantle lithosphere, in turn enabling a dominantly ductile
distribution of deformation in the mantle lithosphere. The imbrication style of deformation
occurs in a mantle lithosphere that is overlain by a rheologically weak lower crust (e.g.,
wet quartzite or felsic granulite) while the degree of R.H.P. in the crust is sufficiently
low. This allows the strong upper portion of the mantle lithosphere to progressively
underthrust adjacent mantle lithosphere along a weak/decoupling crust-mantle interface.
Imposed convergence is accommodated in the mantle lithosphere by underplating when
the lower crust is rheologically stronger, e.g., mafic granulite. The stronger lower crust
does not decouple from the underlying mantle lithosphere to the extent that it would
if it were more felsic. As a result, the lithosphere deforms as a single plate, more akin
to oceanic lithosphere. These three styles of deformation result in the emplacement of
plate-like mantle lithosphere at depths of ∼200-400 km.

Recent papers (e.g., Rey and Houseman, 2006; G. Duclaux and Menot, 2007) demon-
strate that two-dimensional (2D) experiments cannot adequately study Archean continen
tal collision because three-dimensional (3D) warm and buoyant lithosphere will undergo
orogen-parallel ductile flow during orogenesis. In some of the experiments the lower crust
accommodates convergence by undergoing pure-shear thickening and if given the opportu-
nity will likely flow somewhat in the out-of-plane direction. Furthermore, our experiments
illustrate that continental collision may be dominantly accommodated in the mantle litho-
sphere along brittle shear zones. We have chosen to focus on the detailed crust-mantle evolution and consequently have chosen a high resolution 2D model while neglecting (recognized) 3D orogenic processes.
Chapter 3

Geodynamic models of mature continental collision: Evolution of an orogen from lithospheric subduction to continental retreat/delamination
Abstract

The behavior of deep lithospheric processes in continental collision are frequently speculated on, but poorly understood. In this study we build on previous work to explore the styles of continental lithosphere deformation during mature collision where \(\sim 1800\) km of convergence has been accommodated by horizontal shortening. We conducted a suite of numerical geodynamic experiments that test the sensitivity of mature continent collision to varying mantle lithosphere density, mantle lithosphere yield stress, lower crustal strength and to the presence of phase change-related density changes in the lower crust (i.e., eclogitization of mafic lower crust). The models suggest that the early stages of collision are accommodated by subduction of lower crust and mantle lithosphere along a discrete shear zone beneath the overriding plate. Following this initial stage of subduction, the subducting lower crust and mantle lithosphere can retreat from the collision zone, permitting the sub-lithospheric mantle to upwell and intrude the overriding plate. As a result, the lower crust and mantle lithosphere of the overriding plate delaminate from the overlying crust. With local isostatic adjustment, subduction- and delamination-driven crustal processes plateau-like uplift occurs. The sub-crustal evolution also causes bands of syn-convergent crustal extension to develop. In models with a rheologically weaker lower crust, surface crustal response to the deep lithosphere dynamics becomes more diffuse. As an example, the numerical experiments satisfy a number of surface observables of the Himalayan-Tibetan orogen, namely: 1) the current general mantle lithosphere architecture as defined by seismic analyses; 2) the long-wavelength plateau uplift of the western Tibetan Plateau; 3) the surface heat flow measurements in India, the Tethyan Himalaya, Qiangtang and in Qaidam basin; and 4) the anomalous syn-convergent extension in the southern portion of the orogen.
3.1 Introduction

The behavior of the deep continental lithosphere (i.e., the continental mantle lithosphere) during periods of mature continental collision is difficult to determine because of the difficulty of studying a remote region through the thickened continental crust, its possibly heterogeneous nature and because the collisions are “less seismic” compared to other plate boundaries (i.e., no Wadati-Benioff zone to delineate the plate). That said, several hypotheses have been proposed to explain how the continental mantle lithosphere fundamentally responds during periods of mature collision (e.g., Willett and Beaumont, 1994; Houseman et al., 1981).

Houseman et al. (1981) proposed that during collision the lower mantle lithosphere deforms by distributed pure-shear thickening followed by convective removal to the (sub-lithospheric mantle) S.L.M. In their concept, the thickened dense lower mantle lithosphere can become gravitationally unstable and under certain conditions develop as a Rayleigh-Taylor-type mantle downwelling. Subsequent studies of Rayleigh-Taylor instabilities of the lithosphere have shown that the process is controlled by the viscous rheology of the lithosphere (e.g., Buck and Toksöz, 1983; Lenardic and Kaula, 1995; Houseman and Molnar, 1997; Molnar et al., 1998), horizontal shortening of the lithosphere (e.g., Conrad and Molnar, 1997; Molnar et al., 1998) and rheological stratification (e.g., Neil and Houseman, 1999; Pysklywec and Cruden, 2004). The explicit assumptions of these models are that: (1) there exists a suitable wavelength perturbation at the base of the lithosphere that initiates Rayleigh-Taylor instability; and (2) the growth rate of the instability outpaces thermal diffusion of the descending cold mantle lithosphere into the underlying S.L.M.

An alternate hypothesis to the distributed pure-shear thickening model is the subduction model (e.g., Beaumont et al., 2001). It implies that during collision shortening is accommodated by underthrusting/subduction of one plate beneath the other along a discrete shear zone. The concept of mantle lithosphere subduction has been incorporated as a basal boundary condition to model crustal deformation in compressional orogens (e.g.,
Willett et al., 1993; Willett and Beaumont, 1994; Beaumont et al., 1996, 2001, 2004). These studies demonstrate that during collision a mantle subduction basal boundary condition is favored over a distributed Rayleigh-Taylor-type instability of the mantle lithosphere because it produces distinctly asymmetrical crustal deformation, a feature observed in many collisional orogens. Willett and Beaumont (1994) showed that 1000 km of shortening results in diffuse deformation of the crust overlying the subducting plate (pro-plate) while the crust of the overriding plate (retro-plate) experienced more localized deformation. In their experiments the crustal suture migrated continuously towards the retro-side of the box (i.e., in the direction of subduction), producing an orogen composed largely of crust derived from the subducting plate (pro-plate). More recent crustal scale studies of continental collision (e.g., Beaumont et al., 2001, 2004, 2006) illustrate that ∼2000 km of shortening may be accommodated by crustal-scale channel flows (gravitationally driven channel flows of low-viscosity, melt-weakened, middle crust that tunnels outward from beneath the model plateau toward its flanks). These channel flows may be permitted to be exhumed/exposed by denudation focused on the high-relief transition between the model plateau and foreland.

Recent results of numerical experiments of continent collision and geophysical constraints have been interpreted to show that small-scale orogens (e.g., the South Island of New Zealand) accommodate ∼100-150 km of convergence by subduction/underthrusting of the upper mantle lithosphere while the lower mantle lithosphere deforms by distributed pure-shear thickening and Rayleigh-Taylor viscous instability (e.g., Pysklywec et al., 2000, 2002). More recent numerical and analogue experiments have focused on the tectonic consequences of previously subducting oceanic lithosphere (e.g., slab break-off and delamination) (e.g., Faccenda et al., 2008, 2009; Warren et al., 2008a,b) and the dynamics of trench motion (e.g., Faccenna et al., 2007). Other studies explore the syn- and post-collisional styles of mantle lithosphere delamination (Gogus and Pysklywec, 2008b,a), i.e., where the mantle lithosphere peels away from the overlying crust in the manner defined
by Bird (1979). While these works provide insight into the styles of mantle lithosphere deformation during collision they are restricted to the early stages of collision.

In this study we build on previous work to explore the styles of deep continental lithosphere deformation during mature collision. In particular, we investigate the development of deep continental lithosphere subduction, retreat and delamination during the collision process. A series of geodynamic experiments are presented that test the sensitivity of mature collision (e.g., 1800 km of convergence) to varying mantle lithosphere density, mantle lithosphere yield stress and to the presence of phase change-related density changes in the lower crust (i.e., eclogitization of mafic lower crust). As a preliminary case study, the numerical results are considered in the context of the ∼50 Ma Himalayan-Tibetan orogen. Though convergence between the Indian and Eurasian plates is responsible for the development of the Himalaya and the Tibetan Plateau, the process of deformation in the deep lithosphere is poorly understood (e.g., Willett and Beaumont, 1994; Houseman et al., 1981). By considering the Himalayan-Tibetan orogen, the study can take advantage of a wealth of geological and geophysical information across the well studied region. That said, portions of the research will be relevant to the development of other mature orogens as well as younger, smaller-scale orogens.

3.2 Modeling mature continental collision

3.2.1 Basic principals

In the numerical experiments we used SOPALE. The plane-strain numerical code employs the Boussinesq approximation and solves for the deformation of high Prandtl number incompressible viscous-plastic media. The governing hydrodynamic equations for the numerical models include the conservation equations of mass, momentum and internal energy:

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

(3.1)
\[ \nabla \cdot \sigma + \rho \mathbf{g} = 0 \quad (3.2) \]
\[ \rho c_p \left( \frac{\partial T}{\partial t} + \mathbf{u} \cdot \nabla T \right) = k \nabla^2 T + H. \quad (3.3) \]

In these equations \( \mathbf{u} \) (m s\(^{-1}\)), \( \sigma \) (Pa), \( \rho \) (kg m\(^{-3}\)), \( \mathbf{g} \) (m s\(^{-2}\)), \( c_p \) (J kg\(^{-1}\) K\(^{-1}\)), \( T \) (K), \( k \) (W m\(^{-1}\) K\(^{-1}\)), \( H \) (W m\(^{-3}\)) and \( t \) (s) are the velocity, stress tensor, density, gravitational acceleration, specific heat capacity, temperature, thermal conductivity, volumetric rate of internal heat production and time. Equations (1) - (3) are solved using the Arbitrary Lagrangian Eulerian finite element method (Fullsack, 1995). In this method, creeping flows are calculated on a Eulerian grid (resolution = 101x401) that is restricted to small vertical dilations (corresponding to the development of topography on the top free surface). At each time step, the vertical component of velocity computed along the top surface of the computational domain is used to incrementally update the position of the top surface of the Eulerian grid. The rest of the Eulerian grid is then evenly dilated or compressed to accommodate the motion of the top surface (Pysklywec and Shahnas, 2003). The Lagrangian grid (resolution = 301x1201) is advected with the computed velocity field and in turn is used to update the evolving material and thermal property distributions on the Eulerian grid.

Equations (3.1) - (3.3) are complemented by an associated linearized equation of state that relates density (\( \rho \)) to temperature (\( T \)):

\[ \rho(T) = \rho_o [1 - \alpha (T - T_o)] \quad (3.4) \]

where \( \alpha \) is the coefficient of thermal expansion, \( \rho_o \) is the reference material density and \( T_o \) is the reference temperature.

In the numerical code plastic deformation is modeled with a pressure-dependent Drucker-Prager criterion. The yield stress is

\[ (J_f')^{1/2} = P \sin \phi_{eff} + C \cos \phi_{eff} \quad (3.5) \]
where $J'_2$ is the second invariant of the deviatoric stress tensor, $C$ is the cohesion, $P$ is the dynamic pressure and $C \cos \phi_{eff} \approx 1$. The effective angle of internal friction, $\phi_{eff}$, is defined to implicitly include the effects of pore fluid pressure through the expression

$$\phi_{eff} = \frac{(P - P_f) \sin \phi}{P}$$

(3.6)

where $P_f$ is the pore fluid pressure and $\phi$ is the angle of internal friction. For dry frictional sliding conditions (i.e., $P_f = 0$) $\phi = 30^\circ$.

For materials undergoing viscous flow deformation a thermally activated power-law creep relation is used (e.g., Gerya, 2009) where the effective viscosity, $\eta_v^{\phi}$, is

$$\eta_v^{\phi} = (3^{-(1+n)/2n}2^{(1-n)/n})A^{(-1/n)}(\dot{I}_2'^{(1-n)/2n})\exp(\frac{Q}{nRT}).$$

(3.7)

and $\dot{I}_2'$ is the second invariant of the deviatoric strain rate tensor, $T$ is the temperature and the variables $A$, $n$ and $Q$ are the material parameter, power exponent and activation energy, respectively.

The nonlinear solution for each time-step is solved using $\eta_v^{\phi}$ when the local flow stress is less than the Drucker-Prager yield stress and $\eta_p^{\phi}$ when the Drucker-Prager yield stress is exceeded. Setting

$$\eta_p^{\phi} = \frac{(J'_2)^{1/2}}{2(\dot{I}_2')^{1/2}}$$

(3.8)

in regions that are on frictional plastic yield (i.e., yield stress equals $(J'_2)^{1/2}$) and deforming at strain rate $(\dot{I}_2')^{1/2}$ allows the velocity field to be computed from the finite element solution for creeping viscous flows (Willett, 1999b; Fullsack, 1995).

### 3.2.2 Model design, material properties and initial conditions

The initial configuration of the model (Figure 3.1) is constructed as an idealized cross-section of continental lithosphere underlain by S.L.M. In the models, a 120 km thick
continental lithosphere- made up of a 21.6 km thick upper-crust, a 14.4 km thick lower-crust and an 84 km thick mantle lithosphere- overlies 480 km of S.L.M. We assume that any previously subducting oceanic lithosphere has detached from adjacent continental lithosphere (e.g., Van der Voo et al., 1999).

**Figure 3.1:** Illustration of physical properties and initial configuration of numerical experiments modeling mature collision. Continental convergence is modeled by introducing new lithosphere at the right margin of the box with velocity 5.0 cm a\(^{-1}\), while left margin of the lithosphere is held fixed and outward flux of 0.64 cm a\(^{-1}\) is distributed evenly along sides of the S.L.M. to balance the injected lithosphere. A weak zone (\(\eta = 1.25 \times 10^{20}\) Pa s) is inserted into the upper mantle lithosphere to seed the onset of deformation in the numerical models. An empirical law for erosion rate, \(E\), in tectonically active regions (Montgomery and Brandon, 2002) is used: \(E = E_0 + \frac{4S}{(1 - (\frac{S}{S_c})^2)}\), where \(K = 0.6\) mm a\(^{-1}\) is a rate constant, \(S_c = 37^\circ\) is a limiting hillslope gradient, \(E_0 = 0.05\) mm a\(^{-1}\) is the background erosion rate due to weathering and \(S\) is mean surface slope.

In the models, convergence of continental lithosphere is imposed as a boundary condition by introducing new lithosphere into the right edge of the box at a horizontal velocity of 5.0 cm a\(^{-1}\) while lithosphere at the left edge of the box is held fixed. To balance the injection of lithosphere into the solution space, a small outward flux of 0.64 cm a\(^{-1}\) is distributed evenly along the sides of the S.L.M. We recognize that the convergence rate between the Indian and Eurasian plates decreased by a factor of \(\sim 3\) when the Indian continental margin initially collided with Eurasia. Furthermore, recent numerical experiments demonstrate that subduction of the dense Indian slab may not stall the process of continental collision, providing an explanation for the current northward advance of the Indian plate (Capitanio et al., 2010). However, we emphasize that the goal of this work is not to reproduce the tectonic evolution of the Himalayan-Tibetan orogen but to
understand the most basic physics behind the development of an evolved orogen. Consequently, the velocity boundary conditions were chosen to be uniform approximations of the Himalayan-Tibetan orogen. The model has a free top surface, allowing for the development of topography as the model evolves, and has zero tangential stress (free slip) at the other three sides. The effects of surface erosion are taken into consideration by employing a slope-dependent erosion law (Montgomery and Brandon, 2002) (Fig. 3.1). Sedimentation is not included in the experiments.

The power-law viscous parameters for the upper crust are based on the experimental results for wet quartzite (Gleason and Tullis, 1995), while those for the lower crust are based on the experimental results of dry Maryland diabase (Mackwell et al., 1998) (Table 3.1). In the models we scale down the pre-exponential term of the dry Maryland diabase lower crust (i.e., $3^{-(1+n)/2n} 2^{(1-n)/n}$) by a factor of 0.1 to reproduce weaker (i.e., more hydrous) conditions. A comparison demonstrates that the power-law flow properties of dry Maryland diabase scaled down by a factor of 0.1 corresponds to the power-law flow properties of intermediate composition granulite (reviewed in Beaumont et al., 2006). Given the uncertainties in the composition of the lower crust, Beaumont et al. (2006) argue that the power-law flow properties of the lower crust can be adequately modelled by scaling those of dry Maryland diabase down between 1 (the strongest end-member) and 0.05 (the weakest end-member). Consequently, we choose a factor of 0.1 (representative of an intermediate scaling factor) for the dry Maryland diabase. The power-law viscous parameters for the mantle lithosphere and the S.L.M. are based on experimental results for wet olivine (Hirth and Kohlstedt, 1996). In the numerical experiments the pre-exponential term of the mantle lithosphere is scaled up by a factor of 100 (Hirth and Kohlstedt, 1996) to model the viscous response of more refractory/stronger olivine. A plastic yield stress of 300 MPa is used for the mantle lithosphere (e.g., Pysklywec, 2006). Using a defined yield stress for the mantle lithosphere, as opposed to employing a yield stress dependent on pressure, $\phi_{eff}$ and cohesion, ensures that the subducting portion of the mantle lithosphere
remains coherent. This coherence occurs because the viscous \((J'_1)^{1/2}\) exceeds the plastic \((J'_2)^{1/2}\) in the upper portion of subducting mantle lithosphere. The coherence is required for the subducting portion to delaminate from the overlying crust in a manner defined by Bird (1979). The frictional plastic rheology of the lower crust and mantle lithosphere are permitted to soften by implementing a linear decrease in \(\phi_{eff}\) and plastic yield stress, respectively, with accumulated strain, \((I'_2)^{1/2}\). Through these strain softening mechanisms, the frictional strength of the lower crust is reduced from \(\phi_{eff} = 15^\circ\) to \(2^\circ\) while the plastic yield stress of the mantle lithosphere is reduced from 300 MPa to 50 MPa through the strain range \(0.5 < (I'_2)^{1/2} < 1.5\). We introduce these as arbitrary mechanisms of strain softening in the lower crust and mantle lithosphere (e.g., Poirier, 1980; Vissers et al., 1995; Jin et al., 1996a; Braun et al., 1999), respectively. Incorporating the ability of the model to self-consistently strain soften and to localize strain may be an important agent for initializing and sustaining deformation during the collision of rigid lithospheric plates (e.g., Pysklywec et al., 2000).

Because the numerical code solves for the deformation of high Prandtl number incompressible viscous-plastic media it neglects the effects of elasticity. That said, initiation of subduction is a process that may be facilitated by the effects of elasticity (e.g., Kaus and Podladchikov, 2006). In most numerical geodynamic experiments, subduction initiation or faulting/shearing is initiated by reducing the cohesion or angle of internal friction over a given interval of strain (e.g., Pysklywec et al., 2000). As Kaus and Becker (2007) point out, elasticity is helpful in these cases because it releases elastically stored energy in small areas and this may lead to further weakening. Less understood is the effect that elasticity may have on lithospheric gravitational instabilities. Kaus and Becker (2007) demonstrated that, under certain conditions, elasticity may speed up the growth-rate of Rayleigh-Taylor instabilities because it reduces the fraction of viscous deformation at timescales shorter than the Maxwell relaxation time. They also showed that for plate tectonics on Earth the effects of elasticity on Rayleigh-Taylor instability are negligible and do not change the
Table 3.1: Physical parameters for reference experiment

<table>
<thead>
<tr>
<th>Mechanical Parameters</th>
<th>Upper crust</th>
<th>Lower crust</th>
<th>Mantle lithosphere</th>
<th>Sub-lithospheric mantle</th>
<th>Weak zone</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ref. density (kg m$^{-3}$)</td>
<td>2800</td>
<td>3000</td>
<td>3360</td>
<td>3310</td>
<td>3360</td>
</tr>
<tr>
<td>Ref. density T (K)</td>
<td>293</td>
<td>293</td>
<td>293</td>
<td>293</td>
<td>293</td>
</tr>
<tr>
<td>Density HP (kg m$^{-3}$)</td>
<td>3000</td>
<td>3400</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>$\phi_1$ (deg.)</td>
<td>15</td>
<td>15</td>
<td>-</td>
<td>15</td>
<td>-</td>
</tr>
<tr>
<td>$\phi_2$ (deg.)</td>
<td>15</td>
<td>2</td>
<td>-</td>
<td>2</td>
<td>-</td>
</tr>
<tr>
<td>$(I'_1)^{1/2}$</td>
<td>0.5</td>
<td>0.5</td>
<td>-</td>
<td>0.5</td>
<td>-</td>
</tr>
<tr>
<td>$(I'_2)^{1/2}$</td>
<td>1.5</td>
<td>1.5</td>
<td>-</td>
<td>1.5</td>
<td>-</td>
</tr>
<tr>
<td>Cohesion (MPa)</td>
<td>10</td>
<td>10</td>
<td>300</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Flow law</td>
<td>Wet quartzite</td>
<td>Dry Maryland diabase</td>
<td>Wet olivine</td>
<td>Wet olivine</td>
<td>$\eta = 1.5 \times 10^{20}$ Pa·s</td>
</tr>
<tr>
<td>Reference$^a$</td>
<td>1</td>
<td>2</td>
<td>3</td>
<td>3</td>
<td>-</td>
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<tr>
<td>$f$</td>
<td>1</td>
<td>0.1</td>
<td>100</td>
<td>1</td>
<td>-</td>
</tr>
<tr>
<td>$A$ (Pa$^{-n}$s$^{-1}$)</td>
<td>$1.1 \times 10^{-28}$</td>
<td>$5.04 \times 10^{-28}$</td>
<td>$4.89 \times 10^{-15}$</td>
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</tr>
<tr>
<td>$n$</td>
<td>4</td>
<td>4.7</td>
<td>3.5</td>
<td>3.5</td>
<td>-</td>
</tr>
<tr>
<td>$Q$, (kJ mol$^{-1}$)</td>
<td>223</td>
<td>485</td>
<td>515</td>
<td>515</td>
<td>-</td>
</tr>
<tr>
<td>Heat capacity (J kg$^{-1}$ K$^{-1}$)</td>
<td>750</td>
<td>750</td>
<td>750</td>
<td>750</td>
<td>750</td>
</tr>
<tr>
<td>Thermal conductivity (W m$^{-1}$ K$^{-1}$)</td>
<td>2.25</td>
<td>2.25</td>
<td>2.25</td>
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<td>Thermal expansion (K$^{-1}$)</td>
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<td>$3.0 \times 10^{-5}$</td>
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<td>Radioactive heat production ($\mu$W m$^{-3}$)</td>
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<td>0</td>
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<td>0</td>
</tr>
</tbody>
</table>

$^a$ 1) Gleason and Tullis (1995); 2) Mackwell et al. (1998); 3) Hirth and Kohlstedt (1996)

timescales for mantle lithosphere detachment.

The reference densities of the mantle lithosphere and S.L.M. are 3360 and 3310 kg m$^{-3}$, respectively. Unless otherwise stated, all numerical experiments include a reference density change in the upper crust ($\rho_o = 2800$ to 3000 kg m$^{-3}$) and lower crust ($\rho_o = 3000$ to 3400 kg m$^{-3}$) when the pressure-temperature conditions reach eclogite facies conditions ($P \geq 1.2$ GPa and $T \geq 500$ °) (Hacker, 1996). Though the density change is reversible, we assume that complete transformation to eclogite facies occurs. The density change is accompanied by a volume change not included in the incompressible calculations. However, we emphasize that the associated volume change is small in these models and affects the velocity field only when the phase change occurs. Latent heat is not included in the energy equation.

The initial geotherm for the experiments is laterally uniform and is defined by a surface temperature of 20°C, an increase to 550°C at the MOHO, an increase to 1350°C at the
base of the mantle lithosphere and an increase to 1575°C at the bottom of the model (Fig. 3.1). The surface and bottom temperatures are held constant throughout the experiments. Owing to the fact that sediments in continental margins accumulate heavy minerals (e.g., zircon) that concentrate U and Th, we set the rate of crustal R.H.P. at 5.0 µW m\(^{-3}\) (e.g., Faccenda et al., 2008; Jimenez-Munt et al., 2008) in a model continental margin in the upper crust of the retro-plate (Fig. 3.1). The remainder of the upper crust has a R.H.P. rate of 2.0 µW m\(^{-3}\) (e.g., Beaumont et al., 2004). Because the density of heat producing elements is much higher in felsic rocks than in mafic rocks, we concentrate all the heat-producing elements in the upper crust.

Finally, a low-viscosity weak zone with viscosity 1.5 \times 10^{20} \text{ Pa\cdot s} and of dimension 20 km by 20 km, is inserted into the top 20 km of the mantle lithosphere to localize the initial deformation in the model (Figure 3.1). The weak zone only influences the initial position of plastic failure planes, not the finite deformation that ensues.

### 3.3 Model results

#### 3.3.1 Mantle lithosphere deformation

**The reference experiment**

Figure 3.2 shows the results of the reference experiment after several stages of imposed convergence. After 317 km of imposed convergence (∆x = 317 km) (Fig. 3.2a), pro-side mantle lithosphere, lower crust and a minor component of upper crust are subducting beneath retro-side lithosphere along a discrete shear zone. The shear strain rate \(\dot{\gamma}_{xy}\) plot shows that strain rate is highest along the mantle lithosphere/S.L.M. interface retro-ward of the shear zone, along the shear zone and along the lower/upper crustal interface within ~300 km of the shear zone. At the surface, this is mainly contractional, although a small zone of extension is active in the center of the orogen.

By ∆x = 951 km (Fig. 2b), the subducting pro-side material has retreated ~100 km...
Figure 3.2: Evolution of the reference model at cumulative convergence of a) 317 km, b) 951 km, c) 1268 km and d) 1800 km. For each primary frame the Lagrangian mesh (initially composed of rectangular cells) is superimposed on the material regions to show internal deformation. Inset frames show filled contours of shear strain rate ($\dot{\gamma}_{xy}$) to 300 km depth and the locations of surface extension (1) ($\dot{\gamma}_{xx} > 0$), compression (-1) ($\dot{\gamma}_{xx} < 0$) or neither (0) ($\dot{\gamma}_{xx} = 0$) to illustrate the rate of deformation of the different components of the solution space.
owing to the density difference between the subducting material (i.e., mantle lithosphere and lower crust) and the underlying S.L.M. and to the decoupling between the upper and lower crust in the area adjacent to the shear zone. In response, the S.L.M. is upwelling to replace the retreating material. The upwelling S.L.M. is also beginning to intrude the retro-plate along the lower/upper crustal interface. The intrusion of S.L.M. is causing the dense retro-side mantle lithosphere and eclogitized lower crust to delaminate from the overlying crust into the less dense underlying S.L.M. In essence, the system is evolving into a type of “double delamination”. The $\dot{\gamma}_{xy}$ plot shows highest strain rate: 1) along the pro-plate lower/upper crustal interface; 2) along the interface between the subducting pro-side lower crust and the upwelling S.L.M.; 3) in the head of the upwelling S.L.M.; 4) along the retro-plate lower/upper crustal interface; and 5) in the entire thickness of the thickened crust. The thickened crust above the subduction zone and above the S.L.M. upwelling is undergoing extension within the overall shortening regime.

By $\Delta x = 1268$ km (Fig. 3.2c), pro-side lower crust and mantle lithosphere have stopped retreating but the S.L.M. continues to intrude the retro-plate along the lower/upper crustal interface while retro-side mantle lithosphere and lower crust delaminate from the overlying crust in a manner first described by Bird (1979). The delaminating retro-side material is dipping shallowly into the S.L.M. but remains intact as a coherent plate; i.e., as opposed to a viscous dripping-type removal. The $\dot{\gamma}_{xy}$ plot illustrates that strain rate is concentrated in the same regions as after 951 km of convergence; the main difference being that at $\Delta x = 1268$ km the high strain rate zones are more developed.

At the latest stage ($\Delta x = 1800$ km) (Fig. 3.1d), retro-side lower crust and mantle lithosphere continue to delaminate from the overlying crust, while pro-side lower crust and mantle lithosphere continue to subduct into the underlying mantle. The delamination hinge has migrated $\sim 600$ km retro-ward of its initial position. Convergence is accommodated in the upper crust dominantly by pure-shear thickening above the zone of S.L.M. upwelling. However, this deformation is characterized by zones of crustal contraction and
extension that correspond to the evolving mantle lithosphere; contraction (at the plateau margins) and extension (within the plateau) are the result of mantle lithosphere delamination (Gogus and Pysklywec, 2008b). It is important to note that eclogitization of the lower crust occurs without kinetic hindrance in the reaction. Consequently, retreat and delamination rates observed in these experiments may be viewed as upper bounds.

**Experiment with a sub-lithospheric mantle reference density of 3360 kg m\(^{-3}\)**

Save for the fact that the initial material density \(\rho_o\) of the S.L.M. is 3360 kg m\(^{-3}\), the setup of RUN1 is identical to the reference experiment. As such, in RUN1, the mantle lithosphere is neutrally buoyant with respect to the S.L.M. (rather than being denser). By \(\Delta x = 1800 \text{ km} \) (Fig. 3.3) ~1800 km of pro-side mantle lithosphere and lower crust have subducted beneath the retro-plate into the underlying S.L.M. along a single discrete shear zone. Above the shear zone, the retro-side mantle lithosphere is deflected upward while mantle lithosphere material retro-ward of this deflection is slightly folded. Imposed convergence is accommodated in the upper crust by pure-shear thickening above and ~750 km retro-ward of the shear-zone. Surface extension occurs above the subduction zone despite the absence of delamination. This extension may be caused by gravitational collapse of the thickened crust at/near the main shear zone. The absence of pro-side mantle lithosphere retreat is a direct result of the mantle lithosphere and S.L.M. having identical reference material densities. The absence of retreat prohibits the underlying S.L.M. from upwelling and eventually intruding the retro-plate along the lower/upper crustal interface, thereby preventing delamination of retro-side mantle lithosphere and lower crust.

**Experiment with an “unscaled” lower crust rheology**

Save for the fact that the power-law flow parameters of the dry Maryland diabase lower crust are not scaled down by a factor of 0.1, RUN2 is identical to the reference model. As such the lower crust in RUN2 is stronger than in the reference experiment. By \(\Delta x = 1800\)
Figure 3.3: Evolution of thermo-mechanical model RUN1 at cumulative convergence of 1800 km. Save for the fact that the reference material density ($\rho_0$) of the sub-lithospheric mantle is 3360 kg m$^{-3}$, RUN1 is identical to the reference model. For each primary frame the Lagrangian mesh (initially composed of rectangular cells) is superimposed on the material regions to show internal deformation. Inset frames show filled contours of shear strain rate ($\dot{I}_{xy}^r$) to 300 km depth and the locations of surface extension (1) ($\dot{I}_{xx}^r > 0$), compression (-1) ($\dot{I}_{xx}^r < 0$) or neither (0) ($\dot{I}_{xx}^r = 0$) to illustrate the rate of deformation of the different components of the solution space.

km (Fig. 3.4) $\sim$1800 km of pro-side mantle lithosphere and lower crust have subducted beneath the retro-plate into the underlying S.L.M. along a single discrete shear zone in a manner similar to that observed in RUN1. Owing to the density difference between the subducting material (i.e., mantle lithosphere and lower crust) and the underlying S.L.M. and to the decoupling between the upper and lower crust in the area adjacent to the shear zone, the subducting pro-side material has retreated $\sim$20 km from the shear zone. The relative absence of material retreat and delamination owes itself to the “unscaled” dry Maryland diabase lower crust. Above the shear zone, the retro-side mantle lithosphere is slightly deflected upward while mantle lithosphere material retro-ward of this deflection is slightly deflected downward. Imposed convergence is accommodated in the upper crust by pure-shear thickening above and $\sim$750 km retro-ward of the shear-zone. Surface extension occurs above the subduction zone and within the model plateau despite the relative absence of material retreat and delamination. As in RUN1, this extension may be caused by gravitational collapse of the thickened crust at/near the main shear zone.
Figure 3.4: Evolution of thermo-mechanical model RUN2 at cumulative convergence of 1800 km. Save for the fact that the power-law flow parameters of the dry Maryland diabase lower crust are not scaled down by a factor of 0.1, RUN2 is identical to the reference model. For each primary frame the Lagrangian mesh (initially composed of rectangular cells) is superimposed on the material regions to show internal deformation. Inset frames show filled contours of shear strain rate ($I''_{xy}$) to 300 km depth and the locations of surface extension (1) ($I'_{xx} > 0$), compression (-1) ($I'_{xx} < 0$) or neither (0) ($I'_{xx} = 0$) to illustrate the rate of deformation of the different components of the solution space.

Experiment with a wet quartzite lower crust

RUN3 is identical to the reference experiment except that the lower crust is modeled after a wet quartzite rheology and undergoes a reference density change from 2800 to 3000 kg m$^{-3}$ when the pressure-temperature conditions reach eclogite facies conditions ($P \geq 1.2$ GPa and $T \geq 500 ^\circ$). By $\Delta x = 951$ km (Fig. 5a), ~951 km of pro-side mantle lithosphere and a minor component of lower crust have subducted beneath the retro-plate along a single discrete shear zone. Notably, the absence of a reference density change to 3400 kg m$^{-3}$ in the lower crust prohibits the bulk of the pro-side lower crust from subducting with the underlying mantle lithosphere and leads to a thicker crust, deforming by pure-shear, above the shear zone. The $I''_{xy}$ plot shows that the regions of highest strain rate are in the lower crust, indicating decoupling between the upper crust and mantle lithosphere; and in the S.L.M. beneath the shear zone.

Owing to the reference density difference between the mantle lithosphere and the S.L.M., after 1800 km of imposed convergence (Fig. 3.5b) the pro-side mantle lithosphere has retreated ~50 km and the underlying S.L.M. has upwelled to replace it. The absence of a strong lower crust (i.e., mafic granulite) leads to more diffuse deformation.
of the crust in the area peripheral to the shear zone than in the reference experiment. The $I_{xy}'$ plot shows the regions of highest strain rate to be in the lower crust between: 1) the upwelling S.L.M., 2) the base of stationary retro-side mantle lithosphere and 3) the top of subducting pro-side mantle lithosphere. The surface strain rate of RUN3 is unique among all the other experiments. Due to the decoupling between the mantle lithosphere and the upper crust, the localized dynamics of the mantle lithosphere are “smeared out” at the surface and crustal deformation is diffuse. The absence of the high reference density change in the lower crust reduces the cumulative density difference between the pro-side mantle lithosphere/lower crust and the underlying S.L.M., retarding retreat of subducting pro-side material in RUN3. As a result, the upwelling S.L.M. does not upwell enough to intrude the retro-plate in order to initiate delamination of retroside mantle lithosphere from the overlying crust.

Figure 3.5: Evolution of thermo-mechanical model RUN3 at cumulative convergence of a) 951 km and b) 1800 km. Save for the fact that the lower crust in RUN3 is modeled after a wet quartzite rheology and undergoes a reference density change from 2800 to 3000 kg m$^{-3}$ when the pressure-temperature conditions reach eclogite facies conditions (Hacker, 1996), RUN3 is identical to the reference model. Inset frames show filled contours of shear strain rate ($I_{xy}'$) to 300 km depth and the locations of surface extension (1) ($I_{xx}' > 0$), compression (-1) ($I_{xx}' < 0$) or neither (0) ($I_{xx}' = 0$) to illustrate the rate of deformation of the different components of the solution space.
Experiment with a mantle lithosphere yield stress of 120 MPa

RUN4 is identical to the reference experiment except that the yield stress of the mantle lithosphere is reduced to 120 MPa (Gogus and Pysklywec, 2008b). After 317 km of imposed convergence (Fig. 3.6a), pro-side mantle lithosphere, lower crust and a minor component of upper crust are subducting beneath the retro-plate along a discrete shear zone. Due to the mantle lithosphere’s reduced yield stress and the weight of the subducting mantle lithosphere and eclogitized lower crust a gravitational Rayleigh-Taylor-type instability develops rapidly. In particular, the limb of subducting material has failed and pro-side mantle lithosphere and lower crust descend into the S.L.M. at rates reaching $\sim 32.0$ cm a$^{-1}$.

Because the lower portion of the subducting pro-side material has dripped into the underlying S.L.M. after $\Delta x = 317$ km, the remaining pro-side material that continues to subduct resides higher in the box. More specifically, there is no dense subducting limb of pro-side material capable of pulling down adjacent pro-side mantle lithosphere and lower crust. Consequently, between $\Delta x = 317$ and 1010 km the subducting pro-side mantle lithosphere and lower crust do not retreat. Rather, the pro-plate and retro-plate collide in a form of ablatve subduction (Tao and O’Connell, 1992).

Due to the weight of the subducting material, after $\Delta x = 1010$ km (Fig. 3.6b) ductile necking occurs in the retro-side mantle lithosphere in a matter described by (Sacks and Secor, 1990). The necking is most apparent in the $I'_{xy}$ plot where high strain rate zones penetrate the retro-side and pro-side mantle lithosphere adjacent to the ablatve subduction zone.

After 1268 km of imposed convergence (Fig. 3.6c) the ablatively subducting retro-side and pro-side material has dripped into the underlying S.L.M., leaving thickened upper crustal material in contact with upwelling S.L.M. Between $\Delta x = 1268$ and 1800 km the upwelling S.L.M. intrudes the retro-plate along the lower/upper crustal interface, initiating delamination of retro-side mantle lithosphere and lower crust from the overlying...
crust. At the final stage of the model’s evolution (Fig. 3.6d), the delamination hinge has migrated 500 km retro-ward of its initial position, while upper crustal material continues to accommodate convergence by pure-shear deformation. Surface extensional zones are just “inland” of each delamination hinge.

Figure 3.6: Evolution of the model RUN4 at cumulative convergence of a) 317 km, b) 1010 km, c) 1268 km and d) 1800 km. RUN4 is identical to the reference experiment except that the yield stress of the mantle lithosphere is reduced to 120 MPa. For each primary frame the Lagrangian mesh (initially composed of rectangular cells) is superimposed on the material regions to show internal deformation. Inset frames show filled contours of shear strain rate (\(\dot{\gamma}_{xy}\)) to 300 km depth and the locations of surface extension (1) (\(\dot{I}_{xx} > 0\)), compression (-1) (\(\dot{I}_{xx} < 0\)) or neither (0) (\(\dot{I}_{xx} = 0\)) to illustrate the rate of deformation of the different components of the solution space.

Though modelling lithospheric deformation processes and mantle flow in three-dimensions (3D) is beyond the scope of this contribution, we emphasize that the S.L.M. upwelling in the reference experiment, RUN2, RUN3 and RUN4 is possibly the upper limit of a
S.L.M. upwelling observed in 3D space. Geodynamic models of oceanic subduction and slab rollback in 3D cartesian geometry show that slab rollback may be accommodated by quasi-toroidal return flow around the lateral slab edges with an upwelling component next to the lateral slab edges (e.g., Schellart, 2004, 2008, 2010; Stegman et al., 2006). Because the S.L.M. upwelling in the experiments presented in this study occurs around the slab tip, it is possible that the dominant S.L.M. upwelling component in 3D would flow around the edges of the delaminating continental lithosphere.

Recent experiments (not shown) suggest that the kinematic velocity boundaries conditions applied to the sides of the box to balance the injected lithosphere create suction and enhance the subducting slab’s tendency to retreat from the collision zone. One way to circumvent this issue would be to balance the injected lithosphere by removing material at the base of the box. Given that the width of the box is $\sim 3 \times$ greater than the combined lengths of the sides, the outward flow applied to the bottom would be $\sim 3 \times$ smaller than that at the sides. This would reduce the influence of the kinematic velocity boundary conditions on the evolution of the model.

### 3.3.2 Model surface topography

For the reference experiment, the surface topography at $\Delta x = 317$ km is dominated by uplift of the crust with an amplitude of 2.5 km (Fig. 3.7a). This uplift is a result of pure-shear thickening of the crust and local isostatic adjustment. Adjacent topography lows at $x = 1400$ km and 1900 km are caused by downward deflection of the lithosphere in the regions adjacent to the shear zone.

By $\Delta x = 951$ km the uplift reaches an amplitude of $\sim 5$ km (Fig. 3.7a). The subducting pro-side material is causing the upper crust to thicken by pure-shear and also by entraining upper crustal material in the shear zone. Uplift occurs retro-ward of the $\sim 5$ km peak as hot S.L.M. flows into the lithosphere breach vacated by the retreating pro-side mantle lithosphere and lower crust. Uplift occurs in the retro-plate, at maximum amplitudes
of ~2.5 km, owing to pure-shear thickening of the upper crust. The negative topography adjacent to the region of uplift is caused by downward deflection of the lithosphere adjacent to subducting pro-side material and retro-side material prior to delamination.

By Δx = 1268 km the surface topography is characterized by broad plateau-like uplift with near-vertical margins (Fig. 3.7a). At this stage in the model’s evolution, the peak in elevation at x = ~1775 km is associated with upper crustal thickening above the subducting pro-side mantle lithosphere and lower crust. The peak in elevation at x = 1400 km is associated with crustal thickening adjacent to the delamination hinge. The slight depression in elevation between the two peaks is a result of thinner crust directly above the S.L.M. upwelling.

By Δx = 1800 km surface topography is still characterized by broad plateau-like uplift with pro- and retro-margin surface slopes of ~0.043 and ~0.028, respectively (Fig. 7a). Most of the uplift is associated with the replacement of the mantle lithosphere and lower crust by underlying buoyant S.L.M. The exception to this at x = 1000 and 1700 km where the crust is anomalously thick due to compression-driven pure-shear deformation. The enhanced local subsidence at x = ~600-800 km is due to active delamination-driven downward deflection of the lithosphere. Through the delamination and retreat processes the main topography anomaly is spatially transient as it migrates to the retro-side of the box in a manner investigated by Gogus and Pysklywec (2008a). More specifically, model plateau growth occurs by retro-ward propagation of the retro-margin ~600 km while the pro-margin migrates ~200 km pro-ward.

In RUN1, uplift initially develops above the subducting pro-side mantle lithosphere and lower crust, with a maximum amplitude of ~5 km (Fig. 7b). The positive topography then migrates towards the retro-side of the box in a manner similar to that observed in the reference experiment. The main difference being that at Δx = 1800 km a topographic peak ~1 km higher than the adjacent area of broad uplift has developed above the shear zone in response to ~100 km thick crust above the shear zone. Another notable difference
between the surface topography evolution of RUN1 and the reference experiment is the remarkably flat topography at elevations of $\sim 5$ km, retro-ward of the topographic peak above the shear zone. This flat topography is the result of upper crustal deformation via pure-shear above a relatively stationary retro-side mantle lithosphere and lower crust. Because of the lithosphere deformation similarities between RUN2 and RUN1, RUN2's topographic evolution (not shown) follows a similar progression to that of RUN1.

In general the surface topography in RUN3 (Fig. 7c) is characterized by amplitudes $\sim 1.5$-2.0 km less than those observed in the reference experiment and RUN1. The other notable difference is that the margins of the region of broad uplift in RUN3 are characterized by shallower slopes. These two differences are the direct result of modeling the lower crust in RUN3 after a wet quartzite rheology. Because the development of plateaux require that the middle and lower crust beneath the margins be strong to support the associated large horizontal pressure gradients, the diffuse crustal deformation caused by the weak lower crust in RUN3 leads to (relatively) lower amplitude topography and gentle plateau margins.

The surface topography of RUN4 (Fig. 7d) develops similarly to the reference experiment with a few exceptions. In RUN4, surface uplift at $\Delta x = 951$ km is more symmetrical and $\sim 1.3$ km less than that of the reference experiment. These two differences are a result of the ablative subduction behavior and the thick v-shaped crust above the subduction zone that characterize RUN4 at $\Delta x = 951$ km. Though surface uplift at $x = 1300$-1400 km is $\sim 4$ km, ablative subduction is causing surface uplift attenuation. The reduced mantle lithosphere yield stress in RUN4 results in more localized mantle lithosphere removal. This produces more localized crustal deformation at the plateau margins, leading to topography $\sim 1$ km higher at the margins than in the center of the model plateau.
Figure 3.7: Plots of surface topography for A) the reference experiment, B) RUN1, C) RUN3 and D) RUN4.

3.4 Interpreting the results in the context of the Himalayan-Tibetan orogen

The Himalayan-Tibetan orogen is the quintessential mature orogen. Though collision between the Indian and Eurasian plates has progressed at a rate of $\sim 5 \text{ cm} \text{ a}^{-1}$ for the past $\sim 50 \text{ Ma}$, it has been proposed that at least 70% (i.e., $\sim 1800 \text{ km}$) of the $\sim 2500 \text{ km}$ of convergence has been accommodated by in-plane thickening, while the remaining <30% has been lost by lateral extrusion into southeast Asia, erosion, or absorption by the mantle (e.g., England and Houseman, 1986; Willett and Beaumont, 1994; Harrison et al., 1992). Because of such extensive in-plane thickening, the dynamic behavior of the continental mantle lithosphere and the interaction between the lithosphere and S.L.M. are ideal for study by plane-strain thermo-mechanical geodynamic experiments.

Geophysical imaging of the deep lithosphere in the region has been carried out by a number of seismic surveys and presents a complex, sometimes contradictory, picture of continental collision. Seismically fast and slow upper mantle beneath the southern and northern parts of the Tibetan Plateau, respectively, have been interpreted as evidence of...
Indian mantle lithosphere and lower crust underthrusting the Tibetan Plateau (e.g., Ni and Barazangi, 1984; Kind et al., 2002; Owens and Zandt, 1997; Li et al., 2008). Whether the Indian lithosphere is subducting horizontally (e.g., Owens and Zandt, 1997) or at a moderate angle (e.g., Jin et al., 1996b; Kosarev et al., 1999; Replumaz et al., 2010; Nabelek and the HI-CLIMB Team, 2009) is uncertain. Also, recent seismic analyses showing a southward dipping boundary beneath the northern plateau have been interpreted as southward subducting Eurasian lithosphere (Kind et al., 2002; Kosarev et al., 1999). Recently, the presence of warm mantle beneath north-central Tibet has been proposed to be the result of asthenospheric upwelling counterflow (e.g., Tilmann et al., 2003). As Tilmann et al. (2003) point out, this counterflow is possibly a response to downwelling Indian mantle lithosphere, manifested as a subvertical high-velocity zone ~100-400 km deep, south of the Bangong-Nujiang Suture. That said, the detection of high-velocity mantle material beneath most, if not all, of the Tibetan plateau suggests that high-velocity Indian mantle lithosphere may underthrust most, if not all, of the plateau (e.g., Prietley et al., 2006). The recent detection of a low-velocity anomaly in the crust and upper mantle beneath southeastern Tibet above a high-velocity anomaly may be explained by lithosphere delamination (Ren and Shen, 2008). In this case, the high-velocity anomaly is interpreted to be delaminating Eurasian mantle lithosphere. The numerical experiments presented in the previous section may suggest a complementary interpretation to the deep seismic data. Specifically, upon collision the Eurasian mantle lithosphere and lower crust began to subduct beneath the Indian plate. After ~475 km of convergence, the subducting Eurasian material began to retreat/delaminate to the north, exposing the thickened upper crust to hot upwelling S.L.M. The upwelling S.L.M. progressively intruded the Indian lithosphere along its upper/lower crustal interface, which resulted in delamination of the Indian mantle lithosphere and lower crust from the overlying upper crust. Thus, the present topography of the Himalayan-Tibetan orogen may be the result of crustal uplift due to subduction of Eurasian lithosphere, delamination of Indian lower crust and mantle
lithosphere and local isostatic compensation of the lithosphere breach.

Our proposed model of retreat and delamination is also in agreement with a number of geophysical surface observations for the Himalayan-Tibetan orogen. A comparison of reference model surface topography at $\Delta x = 1800$ km and present-day surface topography across the western portion of the Himalayan-Tibetan orogen (between $79.3^\circ$E, $25^\circ$N and $86.5^\circ$E, $40^\circ$N) illustrates a similar plateau uplift (Fig. 3.8a). More specifically, the long-wavelength plateau uplift of the western Tibetan Plateau is consistent with retreat of subducting Eurasian mantle lithosphere and lower crust beginning $\sim 37$ Ma followed by delamination of Indian mantle lithosphere and lower crust beginning $\sim 23.5$ Ma. The short-wavelength topographic features in the observed profile are the result of geomorphologic processes not included in the numerical experiments. Though the uplift history of the western Tibetan Plateau is poorly constrained it has been suggested that portions of it reached their current elevation $\sim 20-40$ Ma (Chung et al., 1998; DeCelles et al., 2007; Rowley and Currie, 2006; Wang et al., 2008). The reference experiment shows that the pro- and retro-margins of the model plateau reached elevations of $\sim 4-4.5$ km after $\sim 1268$ km of collision (i.e., corresponding to $\sim 35$ Ma), a time scale in line with the estimates of DeCelles et al. (2007) and Wang et al. (2008). Despite this first-order similarity, it is important to note that uplift of the model plateau continues until the model’s end whereas uplift of the central Tibetan plateau ceased after attaining its current elevation of $\sim 4.5-5$ km $\sim 26-35$ Ma (e.g., DeCelles et al., 2007). A topographic feature present in all the numerical experiments that doesn’t appear in the observed section is the downward deflection of the surface topography in the areas adjacent to the plateau-like uplift. We propose that the flexural strength of the lithosphere (a feature not included in the experiments) may smooth these deflections in areas adjacent to the Himalayan-Tibetan orogen.

Figure 3.8b shows a comparison of reference model surface heat flow and a compilation of measured surface heat flow values available from the central Tibetan Plateau (Jimenez-Munt et al., 2008). Though our model results correspond to the western portion of the
Himalayan-Tibetan orogen (i.e., between 79.3°E, 25°N and 86.5°E, 40°N) and show variations of \(\sim 50 \text{ mW m}^{-2}\) in the northern Himalaya and northern Lhasa, the general trend of model surface heat flow agrees well with measured values in India, the Tethyan Himalaya, Qiangtang and the highest measured values in Qaidam basin. We emphasize that the especially high heat flow values (\(\sim 120-160 \text{ mW m}^{-2}\)) measured in the northern Himalaya are believed to be the result of upper crustal aqueous fluids or widespread crustal melting (e.g., Francheteau et al., 1984; Wei et al., 2001), features not included in the numerical experiments. Perhaps most significantly, our experiments present an upper bound on the amount of surface heat flow due to thickened continental margin upper crust rich in heat producing elements and delamination of mantle lithosphere/lower crust.

![Figure 3.8](image)

**Figure 3.8:** A) Profile of modeled (reference model; \(\Delta x = 1800 \text{ km}\)) and observed topography (between 79.3°E, 25°N and 86.5°E, 40°N). B) Modeled and observed surface heat flow. Surface heat flow measurements were compiled by Jimenez-Munt et al. (2008).

Figure 3.9 shows a comparison of the xx-component of surface strain rate (\(\dot{\varepsilon}_{xx}^\prime\)) in the
reference model and primary structural features across the western Tibetan Plateau. Most interesting is a zone of surface extension in the numerical experiment that corresponds with the observed South Tibetan Detachment System (STDS), a zone of east-west striking, north-dipping normal faults in the Higher Himalayas (Burchfiel et al., 1992). The STDS is particularly enigmatic because it represents extension parallel to and contemporaneous with shortening within the Himalayan-Tibetan orogen to the south. Mechanisms that attempt to explain the STDS range from: 1) gravity-driven topographic collapse of the southern edge of the Tibetan Plateau (Burchfiel and Royden, 1985; Royden and Burchfiel, 1987); 2) exhumation of the greater Himalayan sequence driven by channel flow and ductile extrusion dynamically linked through the effects of surface denudation (Beaumont et al., 2001); 3) Eocene to present southward propagation of deformation from the Indus-Yalu suture zone in a conventional thrust belt (Robinson et al., 2006). In the latter mechanism normal faulting may be the result of orogenic wedge geometry reconfiguration in order to restore the critical taper. Recently, Gogus and Pysklywec (2008b) have shown that mantle lithosphere delamination can cause distinct zones of crustal contraction and extension at the plateau margins and within the plateau, respectively. Similarly, we propose that the anomalous syn-convergent extension of the STDS is the result of delamination of the Indian mantle lithosphere/lower crust. In particular, upwards and diverging S.L.M. flow within the delamination gap is causing significant extension of the overlying crust even though it is within an overall shortening regime. However, we find that the active bands of extension tend to be more localized near the delamination hinge zones (i.e., perhaps akin to a mantle wedge region). Though we suggest that the STDS is the result of delamination of the Indian mantle lithosphere, our model does not reproduce a number of features characteristic of the STDS: 1) the creation of gneiss domes in the southern Tibetan plateau; 2) the juxtaposition of contrasting protoliths across the STDS; and 3) the style and timing of metamorphism in the Greater Himalayan and Lesser Himalayan. These features may be present in future models of continental collision with higher numerical
3.5 Conclusions

We performed thermo-mechanical geodynamic experiments of mature continental collision ($\Delta x = 1800$ km) in an effort to better understand the progression of deep continental lithosphere deformation during periods of collision. The numerical experiments suggest that the early stages of collision ($\Delta x < 475$ km) are accommodated by subduction of lower crust and mantle lithosphere along a discrete shear zone beneath the overriding plate. This period of collision thickens the upper crust and causes uplift. Following this initial stage of subduction, the subducting lower crust and mantle lithosphere retreat from the collision zone by $\sim 100$ km because of their high cumulative density compared to the underlying S.L.M. This rather short-lived period of retreat permits the S.L.M. to upwell, come into contact with the thickened upper crust and intrude the overriding plate along
the upper/lower crustal interface. As the S.L.M. progressively intrudes the overriding plate, the lower crust and mantle lithosphere delaminate from the overlying crust.

Due to local isostatic compensation, subduction- and delamination-driven crustal processes uplift occurs across a zone \( \sim 1000 \) km wide, resulting in topography between \( \sim 4.5-5.5 \) km. The numerical experiments emphasize the importance of dense mantle lithosphere, lower crustal strength, phase change-related density changes in the lower crust and a relatively high mantle lithosphere yield stress. A mantle lithosphere denser than the underlying S.L.M. allows retreat of subducting mantle lithosphere at sufficiently high rates so to allow S.L.M. upwelling. In effect, this retreat sets up a type of “double delamination” as both sides of the mantle lithosphere peel away from the upper crust and fall back from the suture zone. Bands of syn-convergent extension develop near the delamination hinge zones owing to upwards return mantle flow. This process is mitigated if the lower crustal strength is too high (e.g., if the lower crust is modelled after a relatively stiff dry Maryland diabase). The presence of an eclogitizable mafic lower crust both enhances the retreat of subducting mantle lithosphere and provides strength to support the large horizontal pressure gradients produced by plateaux. The high mantle lithosphere yield stress prohibits excessive localization of crustal deformation at the plateau margins, allowing for a relatively flat-topped plateau.

The results are markedly different in the case of the wet quartzite lower crust (RUN3). Here, topography and upper crustal strain are less localized and correspond less firmly to the mantle lithosphere dynamics. Clearly, decoupling between lithospheric horizons has a significant impact on the surface expression of collisional tectonics.

The numerical experiments satisfy a number of surface observables of the \( \sim 50 \) Ma Himalayan-Tibetan orogen: 1) the evolution of model mantle lithosphere deformation is in agreement with the current general mantle lithosphere architecture as interpreted by seismic analyses; 2) at first order, the long-wavelength plateau uplift of the western Tibetan Plateau is consistent with the reference model at \( \Delta x = 1800 \) km; 3) the general
trend of model surface heat flow agrees well with measured values in India, the Tethyan
Himalaya, Qiangtang and the highest measured values in Qaidam basin; and 4) a zone of
surface extension in the model plateau that spatially corresponds to the anomalous STDS.

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Chapter 4

Influence of sediment deposition on deep lithospheric tectonics
Abstract

Previous geodynamic models of continental collision show that the behavior of the lithosphere can be strongly influenced by the presence of surface erosion. That said, absent from these investigations are the effects of sediment deposition. We quantitatively investigate this process using thermal-mechanical numerical experiments of the coupled processes of tectonic deformation and crustal mass flux. The models demonstrate that the inclusion of the effects of sediment deposition can change the style of deformation of the crust and consequently, the evolution of the underlying deforming mantle lithosphere. In the absence of sediment deposition, the early stages of collision are accommodated by subduction of lower crust and mantle lithosphere along a discrete shear zone beneath the overriding plate. Following this initial stage of subduction, the subducting lower crust and mantle lithosphere retreat from the collision zone, permitting the sub-lithospheric mantle to upwell and come into contact with the thickened upper crust. When sediment deposition is imposed subduction-like consumption of the subducting plate remains stable. The presence of sediment deposition introduces a negative vertical component of velocity ($V_y$-component) in the overriding plate in the area adjacent to the collisional zone. The negative $V_y$-component leads to a greater degree of coupling between the colliding continental plates and decoupling of the overriding upper crust and lower crust/mantle lithosphere. The results demonstrate the first quantitative insights into the feedback between surface deposition and deep continental lithosphere deformation.

4.1 Introduction

Geological observations/interpretations as well as crustal- and lithosphere-scale geodynamic models show the behavior of the lithosphere to be strongly influenced by climate-controlled surface processes. The first crustal-scale numerical treatments of climate-tectonic interactions showed that surface denudation, occurring over distances and at rates
comparable to tectonic mass flux, can significantly modify the evolution of a continental orogen (e.g., Beaumont et al., 1992; Willett, 1999a). Later crustal-scale experiments demonstrated that during continental collision the coupled processes of surface erosion and tectonic deformation could produce a modified nature of folding and higher topography than in the absence of erosion (Simpson, 2006); highly localized crustal deformation in response to concentrated erosion (Stolar et al., 2007); topographic asymmetry with shallower slopes facing the subducting plate and an asymmetric pattern of exhumation (Willett, 1999a); and localized crustal shortening as well as the development of an intracontinental mountain range (Avouac and Burov, 1996). The geodynamic investigation of the influence of climate on the tectonic evolution of collisional orogens is not restricted to numerical treatments. Scaled analogue experiments clearly show that the location, mode and rate of surface erosion influence the style, pattern and rate of rock deformation (Whipple, 2009). Extending this theme to the deep lithosphere (i.e., sub-crustal lithosphere), lithosphere-scale models of continental collision showed a remarkably deep reach of surface erosional processes. For example, in the presence of surface erosion a stable subduction-like plate consumption of continental lithosphere may be maintained, whereas in the absence of erosion, continental subduction is inhibited by the accumulation of buoyant crust and can lead to retreat of the initially subducting plate (Pysklywec, 2006). In summary, these studies (and others) have developed an important new understanding of tectonic solid Earth systems with respect to the surface forcings of the active hydrosphere and atmosphere.

The concepts of climate-tectonic interactions derived from this research have been used to explain various enigmatic features of, specifically, collisional zones. The asymmetric elevation profile of the Southern Alps of New Zealand, where the retro-wedge is characterized by wetter and steeper slopes, may be explained by the dominance of tectonic advection over fluvial incision in controlling topography (Willett, 1999a). Pysklywec (2006) suggests that very high surface erosion rates on South Island, New Zealand may explain the presence of
subduction of sub-crustal lithosphere in response to the convergent component of oblique collision. This is in sharp contrast to the deformation beneath the Transverse Ranges of California where subduction-like behavior is not occurring (Houseman et al., 2000), possibly due to relatively subdued surface processes. In the Olympic Mountains of Washington State, high exhumation (~14 km) in the center of the orogen, with decreasing exhumation towards to the periphery of the orogen may be caused by enhanced pro-wedge erosion (Willett, 1999a). As a last example, in the southern portion of the Himalayan-Tibetan orogen the South Tibetan Detachment System, a series of north-dipping, east-west striking normal faults, may be the result of exhumation of the greater Himalayan sequence driven by channel flow and ductile extrusion dynamically linked through the effects of surface denudation (Beaumont et al., 2001).

Most numerical models of continental collision include the combined effects of surface erosion and sediment deposition (e.g., Burov and Toussaint, 2007; Yamato et al., 2008). Burov and Toussaint (2007) for instance showed that the presence of moderate erosion/sedimentation may reduce the resistance of the major thrust and subduction channel to subduction, whereas very strong or weak erosion/sedimentation may enhance plate coupling and promote pure-shear thickening or buckling. Despite the presence of erosion/sedimentation in most numerical models of continental collision, how, specifically, sediment deposition may modify the evolution of the continental lithosphere during plate collision is not well understood. As the conjugate to surface erosion, it may be expected that deposition of the removed mass may similarly influence the lithospheric tectonics. Here, we quantitatively investigate this process using thermal-mechanical numerical experiments of the coupled processes of tectonic deformation and crustal mass flux. A free surface, prescribed erosional laws (e.g., empirically derived relief-dependent erosion) and sediment deposition dependent on the amount of material eroded make up the top boundary of the model domain and allow topography to develop self consistently with the underlying geodynamics. We conduct a series of experiments with varying convergence
rates, initial MOHO temperatures and degrees of sediment deposition to show how the inclusion of the effects of sediment deposition produce a modified style of continental plate behavior during collision, as one example of tectonic activity where climate-tectonic interactions are most often invoked. The results give some of the first insights into how deep tectonics is influenced by climate-controlled erosion-deposition processes.

4.2 Methodology

We solve the coupled conservation equations of mass, momentum and internal energy that govern the behavior of plane-strain (i.e., 2D), visco-plastic incompressible media by using the arbitrary Lagrangian-Eulerian finite element method (Fullsack, 1995). Idealized continental collision is modelled by introducing new lithosphere at the right (pro-) margin of the box at a velocity of 1.5 cm a\(^{-1}\), while lithosphere at the left (retro-) margin of the box is held fixed (Fig. 4.2a). To ensure mass balance, an outward flux of 0.19 cm a\(^{-1}\) is distributed evenly along the sides of the sub-lithospheric mantle (S.L.M.) A small weak seed (10 km \(\times\) 10 km) is inserted into the upper mantle lithosphere to localize the initial deformation. An empirical law for relief-dependent erosion rate, \(E\), in tectonically active regions (Montgomery and Brandon, 2002) is used and applied to the entire width of the model domain: \(E = E_o + \frac{kR}{[1-(R/R_c)^2]}\), where \(k = 2.5 \times 10^{-4}\) mm a\(^{-1}\) is a rate constant, \(R_c = 1500\) m is limiting relief, \(E_o = 0.01\) mm a\(^{-1}\) is the background erosion rate due to weathering and \(R\) is mean local relief. We model deposition by: 1) summing the material eroded on the model orogen’s respective slope faces; 2) dividing the total amount of material eroded on the respective slope faces by the number of Eulerian nodes that lie between the midpoint of the adjacent basin and the Eulerian node at sea-level and adjacent to the model orogen; and 3) adding the result to each Eulerian node that lies between the midpoint of the adjacent basin and the Eulerian node at sea-level and adjacent to the model orogen (Fig. 4.1).

The initial configuration of the experiments consists of 28.8 km of crust (19.2 km of
Figure 4.1: Schematic diagram describing how sediment deposition is implemented in the numerical model. Black arrows demonstrate the direction, up or down, in which the Eulerian mesh moves to accommodate sedimentation or erosion, respectively. Green region describes schematically how eroded material is transferred from the model orogen to its adjacent basins.

upper crust ($\rho_\circ = 2800$ kg m$^{-3}$) overlying 9.6 km of lower crust ($\rho_\circ = 3000$ kg m$^{-3}$)) overlying 91.2 km of mantle lithosphere ($\rho_\circ = 3360$ kg m$^{-3}$), overlying 480 km of sublithospheric mantle ($\rho_\circ = 3310$ kg m$^{-3}$). The relatively thin crustal layer (28.8 km) was chosen to match that of relatively young continental crust (e.g., South Island, New Zealand (Reyners and Cowan, 1993)).

An inherent assumption of the experiments presented in this contribution is that the mantle lithosphere is more dense than the underlying mantle. Recent studies of mantle xenoliths/xenocrysts have been interpreted to imply that Phanerozoic lithosphere is actually less dense than underlying mantle at the same temperature, and the greater density comes from the lower temperatures (e.g., Poudjom Djomani et al., 2001). However, the lower lithosphere may become denser than the underlying mantle by melt intrusion and freezing (Elkins-Tanton and Hager, 2000). Jull and Kelemen (2001) showed that the densities of lower crustal compositions resulting from arc magmatism (e.g., olivine pyroxenite) may exceed that of the S.L.M. by 50-250 kg/m$^3$. Given that olivine pyroxenites are also found as cratonic and non-cratonic mantle xenoliths it is not unreasonable to assume that, under certain conditions, continental mantle lithosphere may be denser than the S.L.M.

The power-law viscous parameters for the upper crust are based on the experimental
results for wet quartzite (Gleason and Tullis, 1995), while those for the lower crust are based on the experimental results of dry Maryland diabase (Mackwell et al., 1998). The impact of the sedimentary fill on the behavior of collisional systems depends not only on the deposition law but also on the assumed thermal and mechanical properties of the sedimentary infill. Because the wet quartzite upper crust is the only lithology eroded in the numerical experiments, the power-law viscous parameters for sedimentary infill are also based on wet quartzite (Gleason and Tullis, 1995). In the models we scale down the effective viscosity of the dry Maryland diabase lower crust by a factor of 0.1 to reproduce weaker (i.e., more hydrous) conditions. The power-law viscous parameters for the mantle lithosphere and the S.L.M. are based on experimental results for wet olivine (Hirth and Kohlstedt, 1996). In the numerical experiments the effective viscosity of the mantle lithosphere is scaled up by a factor of 100 (Hirth and Kohlstedt, 1996) to model the viscous response of more refractory/stronger olivine.

A plastic yield stress of 300 MPa is used for the mantle lithosphere. As in Chapter 3, we employ a defined mantle lithosphere yield stress of 300 MPa because, as opposed to employing a yield stress dependent on pressure, $\phi_{eff}$ and cohesion, it ensures that the subducting portion of the mantle lithosphere remains coherent. This coherence is required for the subducting portion to delaminate from the overlying crust in a manner defined by Bird (1979). The frictional plastic rheology of the mantle lithosphere is permitted to soften by implementing a linear decrease in plastic yield stress with accumulated strain, $(I'_2)^{1/2}$. Through this strain softening mechanism, the plastic yield stress of the mantle lithosphere is reduced from 300 MPa to 50 MPa through the strain range $0.5 < (I'_2)^{1/2} < 1.5$. We introduce this as an arbitrary mechanism of strain softening in the mantle lithosphere (e.g., Jin et al., 1996a). Incorporating the ability of the model to strain soften and to localize strain may be an important agent for initializing and sustaining deformation during the collision of rigid lithospheric plates (e.g., Pysklywec et al., 2000). In all the numerical experiments plastic strain-softening allows the initial lithosphere-scale shear zone and its
orientation to develop self-consistently.

All numerical experiments include a reference density change in the lower crust \((\rho_0 = 3000 \text{ to } 3400 \text{ kg m}^{-3})\) when the pressure-temperature conditions reach eclogite facies conditions \((P \geq 1.2 \text{ GPa and } T \geq 500 \text{ °C})\) (Hacker, 1996; Gray and Pysklywec, 2012a). The density of the eclogitized lower crust \((3400 \text{ kg m}^{-3})\) was chosen to match that of eclogitized rocks that are originally gabbroic (e.g., Bousquet et al., 1997). The surface and bottom temperatures are held constant throughout the experiments at 20°C and 1520°C, respectively. We set the rate of crustal radioactive heat production \((R.H.P.)\) at 2.0 \(\mu\text{W m}^{-3}\) (Beaumont et al., 2004). Because the density of heat producing elements is much higher in felsic rocks than in mafic rocks, we concentrate all the heat-producing elements in the upper crust.

In the suite of experiments with variable MOHO temperatures, the MOHO temperatures were modified by adjusting the heat flow into the base of the crust. Consequently, because the heat flow into the base of the mantle lithosphere was fixed at 20 mW m\(^{-2}\), the thickness of the thermal lithosphere (depth to 1350 °C) was adjusted accordingly, whereas the thickness of the compositional mantle lithosphere was kept constant at 91.2 km.

### 4.3 Results

The results of RUN1 (Fig. 4.2b-e) show how idealized continental collision may proceed when continental lithosphere is subject to varying surface erosion. After 285 km of imposed convergence \((\Delta x = 285 \text{ km})\), corresponding to 19 m.y., retro-plate upper crust and lower crust/mantle lithosphere separate at the collision zone. The upper crustal material is thrust up and the lower crust/mantle lithosphere are thrust down along retrovergent shear zones. Although a portion of the subducting retro-plate has detached and descended into the S.L.M. (Fig. 4.2b), at this stage in the model’s evolution the localized deformation is generally stationary (i.e., the subducting retro-plate material shows no propensity for retreat). Owing to the cumulative density difference between the subducting retro-side
material (i.e., lower crust and mantle lithosphere) and the overriding plate \( (\delta \rho = 53 \text{ kg m}^{-3}) \) and to the decoupling between the upper and lower crust in the area adjacent to the shear zone, by \( \Delta x = 577 \text{ km} \) the subducting retro-side material has started to retreat from the shear zone. The initiation of retreat has allowed a portion of upper crustal material to enter the subduction zone. By \( \Delta x = 600 \text{ km} \), subducting retro-plate lower crust and mantle lithosphere continue to retreat from the subduction zone. By this stage in the model’s evolution, the retreat of subducting retro-plate material has sufficiently progressed to allow upwelling S.L.M. to come in contact with eclogitized retro-plate lower crust that has accumulated in the subduction zone. At the end of the model’s evolution \( (\Delta x = 855 \text{ km}) \), the subducting retro-plate material has retreated from the subduction zone by \(~250 \text{ km} \), allowing the hot S.L.M. to upwell and come in contact with the thickened upper crust. It is important to note that eclogitization of the lower crust occurs without kinetic hindrance in the reaction. Consequently, retreat rates observed in these experiments may be viewed as upper bounds.

In experiment RUN2 where sediment deposition is imposed, subduction-like consumption of the retro-plate remains stable throughout the model’s evolution (Fig. 4.2f-i). The differences in the fundamental behavior of the two experiments may be understood by considering how sediment deposition alters the kinematics of the pro-plate in the area adjacent to the collision zone. Initially experiments RUN1 and RUN2 are similar, but by \( \Delta x = 577 \text{ km} \), important differences develop. In RUN2, by \( \Delta x = 577 \text{ km} \) the incoming pro-plate has a \( y \)-component (vertical) of velocity \( (V_y) \) of \(~0.1 \text{ cm a}^{-1} \) (velocity profile taken at the black vertical line on the pro-plate lithosphere). The negative \( V_y \)-component is due to sediment deposition in the basins adjacent to the model orogen and the resultant eclogitization of the pro-plate lower crust. In essence, sediment deposition is pushing down the incoming pro-plate, leading to eclogitization of the lower crust and a greater degree of coupling with the retro-plate (Figs. 1g and 2f). It is important to note that the formation of lower crustal high density eclogite is fundamental to this process. That said, due to the
negative $V_y$-component, the pro-plate has descended to deeper levels in the computational box.

This has lead to higher crustal temperatures and consequently a zone of high shear strain rate at the pro-plate upper crustal/lower crustal interface (Fig. 4.3g). Effectively, this zone of high shear strain rate decouples the pro-plate upper crust from the underlying lower crust/mantle lithosphere, leading to a more diffuse style of upper-crustal deformation (Fig. 4.3g). By $\Delta x = 600$ km, the $V_y$-component of the pro-plate persists, continuously coupling the retro- and pro-plates and disallowing the minor component of retro-plate retreat that is present in RUN1 at the same stage of the model’s evolution. By $\Delta x = 855$ km, the continued sedimentation has permitted the retro- and pro-plates to remain coupled (i.e., the collision zone is characterized by compressional stresses (Faccenda et al., 2009)). By continuing to create a negative $V_y$-component in the pro-plate, sedimentation has also resulted in eclogitization of the pro-plate’s lower crust along a horizontal distance of $\sim 250$ km. It is interesting to note how the average $V_y$-component in the pro-plate in the area adjacent to the collisional zone is dependent on the deposition % (e.g., 20% deposition means that 20% of the material eroded on the orogen’s slopes is deposited in the adjacent basin).

Figure 4.4 shows that for increasing deposition %’s the average $V_y$-component of the pro-plate in the area adjacent to the collisional zone increases in negativity. This occurs because with increasing deposition % more mass is deposited in the basins adjacent to the model orogen, exerting a greater downward force on the pro-plate.

Experiments RUN1 and RUN2 illustrate how sedimentation can control deep lithosphere processes; namely active sedimentation prevents continental lithosphere retreat during continental collision. We ran a suite of experiments that test the sensitivity of this model to varying convergence rates and deposition %’s as well as varying initial MOHO temperatures and deposition %’s. Figure 4.5a illustrates how the horizontal distance along which the pro-plate’s lower crust is eclogitized increases with decreasing conver-
A) Upper crust (19.2 km)
Lower crust (9.6 km)
Mantle lithosphere:
dry olivine
Weak zone
Sub-lithospheric mantle:
wet olivine

V_o = 0.19 cm/yr
z = 0 km

B) ΔX=285 km
C) ΔX=577 km
D) ΔX=600 km
E) ΔX=855 km

F) ΔX=285 km
G) ΔX=577 km
H) ΔX=600 km
I) ΔX=855 km

Vy (cm/yr)
0 0.2-0.2
1.4-1.6
0.2 0.2
0 0.4-0.4
0 0.2-0.2
0 0.4-0.4
0 1.6-1.6
0 0.2-0.2
**Figure 4.2:** a) Illustration of physical properties and initial configuration of numerical experiments modeling idealized continental collision (see text for further explanation). For all numerical experiments, $\rho_0 = 2800 \text{ kg m}^{-3}$, $\rho_0 = 3000 \text{ kg m}^{-3}$, $\rho_0 = 3360 \text{ kg m}^{-3}$ and $\rho_0 = 3310 \text{ kg m}^{-3}$ are used for upper crust, lower crust, mantle lithosphere and sub-lithospheric mantle, respectively. Frames b), c), d) and e) show the evolution of RUN1 at $\Delta x = 285 \text{ km}$, $\Delta x = 577 \text{ km}$, $\Delta x = 600 \text{ km}$ and $\Delta x = 855 \text{ km}$, respectively. Frames f), g), h) and i) correspond to $\Delta x = 285 \text{ km}$, $\Delta x = 577 \text{ km}$, $\Delta x = 600 \text{ km}$ and $\Delta x = 855 \text{ km}$, respectively, but for RUN2. For each primary frame the Lagrangian mesh (initially composed of rectangular cells) and the velocity vectors are superimposed on the material regions to show internal deformation and the instantaneous material displacement rate of change, respectively. Inset frames show change of the y-component of velocity ($V_y$) of the overriding plate with depth for each model to illustrate the effect of sediment deposition on $V_y$ of the overriding plate.

gence rates and increasing deposition %. Because the numerical experiments undergo the same amount of shortening ($\Delta x = 855 \text{ km}$), the scaled duration of the experiments increases with decreasing convergence rates. Consequently, once the pro-plate’s lower crust undergoes eclogitization it will undergo a greater degree of eclogitization at lower convergence rates because the numerical experiments with lower convergence rates require more time to achieve total shortening of $\Delta x = 855 \text{ km}$. The degree of eclogitization of the pro-plate’s lower crust increases with increasing deposition % because the negativity of the $V_y$-component in the pro-plate increases with increasing deposition %. Figure 4.5b illustrates how the amount of retro-plate deep lithosphere retreat increases with decreasing convergence rates and decreasing deposition %. The amount of deep lithosphere retreat increases with decreasing convergence rates because the retro-plate is subjected to the pull of the subducting/retreating retro-plate material for greater amounts of time and because the incoming plate is too slow to fill the gap as the retro-plate retreats; retro-plate retreat increases with decreasing deposition % because the decreasing negativity of the pro-plate’s $V_y$-component that accompanies a decrease in deposition % leads to increasing retreat. Perhaps most interesting is how deposition % trumps convergence rate in controlling the degree of pro-plate lower crust eclogitization and retro-plate deep lithosphere retreat. We emphasize that this is the result of the $V_y$-component applied to the incoming pro-plate
Figure 4.3: Shear strain rate ($\dot{I}_{xy}$) in numerical experiments RUN1 and RUN2. Frames a), b), c) and d) correspond to $\Delta x = 285$ km, $\Delta x = 577$ km, $\Delta x = 600$ km and $\Delta x = 855$ km, respectively, in RUN1. Frames e), f), g) and h) correspond to $\Delta x = 285$ km, $\Delta x = 577$ km, $\Delta x = 600$ km and $\Delta x = 855$ km, respectively, in RUN2.
Figure 4.4: Illustration of the relationship between the percentage of deposition and the average Vy-component in the pro-plate in the area adjacent to the collisional zone at $\Delta x = 665$ km for the suite of experiments where the convergence velocity is set to $1.5 \text{ cm a}^{-1}$. as a result of sediment deposition.

Figure 4.5: Illustration of the relationship between a) the percentage of deposition and the horizontal distance of pro-plate lower crust eclogitization for various convergence velocities at $\Delta x = 855$ km; b) the percentage of deposition and the total retreat of retro-plate deep lithosphere for various convergence velocities at $\Delta x = 855$ km. Retro-plate deep lithosphere retreat is measured as the horizontal distance along which sub-lithospheric mantle is in contact with upper crust, retro-ward of the shear zone.

Figure 4.6a illustrates how the horizontal distance along which the pro-plate’s lower crust is eclogitized increases with increasing MOHO temperatures and increasing deposition %. Increasing the initial MOHO temperature increases the eclogite stability area (pressure $\geq 1.2$ GPa and temperature $\geq 500 \degree \text{C}$) in the computational domain. Similar to the suite of experiments where the convergence velocity was varied, increasing deposition %’s increase the horizontal distance along which the pro-plate lower crust is eclogitized because the negativity of the $V_y$-component in the pro-plate increases with increasing de-
Figure 4.6b illustrates how, for initial MOHO temperatures between 475°C and 625°C the amount of retro-plate deep lithosphere retreat increases with increasing initial MOHO temperatures. Increasing the initial MOHO temperature decreases the viscosity of the upper crust, increasing the degree of decoupling between the retro-plate upper crust and lower crust/mantle lithosphere. This increases the propensity of the retro-plate deep lithosphere to retreat from the collisional zone. With an initial MOHO temperature of 700°C the high temperatures in the model place more of the upper mantle lithosphere into the viscous regime and result in a thinner layer of plastic mantle lithosphere. The initially high MOHO temperature results in rapid Rayleigh-Taylor downwellings of the underthrusting deep retro-plate material (e.g., Pysklywec et al., 2002). Consequently, retreat of subducting deep retro-plate material occurs at a slower rate and results in less retreat than the models with initially cooler MOHO conditions. With an initial MOHO temperature of 400°C the deep upper crust in the area adjacent to the collision zone is in the plastic regime, producing an initially coupled upper crust and lower crust/mantle lithosphere. As the model progresses in time the comparatively low deep lithosphere temperatures result in gravitational instability greater than that which can be supported by coupling of the upper crust and lower crust/mantle lithosphere. The result is deep-lithosphere retreat independent of deposition %’s.

4.4 Discussion and conclusions

The numerical geodynamic experiments demonstrate that the effects of sediment deposition may alter the style of deep lithosphere deformation during continental collision. Sediment deposition in the basins adjacent to the model orogen significantly influences crustal mass flux within an actively deforming orogen. This consequently alters crustal evolution (particularly in the overriding plate) at the lower crust/mantle lithosphere interface and the style of mantle lithosphere deformation a the plate boundary. The presence of sediment deposition introduces a negative $V_y$-component in the overriding plate in the
Figure 4.6: Illustration of the relationship between a) the percentage of deposition and the horizontal distance of pro-plate lower crust eclogitization for various initial MOHO temperatures at \( \Delta x = 855 \) km; b) the percentage of deposition and the total retreat of retro-plate deep lithosphere for various initial MOHO temperatures at \( \Delta x = 855 \) km.

area adjacent to the collisional zone. The negative \( V_y \)-component leads to: a) a greater degree of coupling between the colliding continental plates, mitigating retreat of deep retro-plate lithosphere; and b) decoupling of the pro-plate upper crust and lower crust, producing a more diffuse style of upper-crustal deformation. In the absence of sediment deposition, the negative \( V_y \)-component in the overriding plate vanishes. As a result, 1) the resulting lack of coupling between the colliding plates combined with the weight of the subducting retro-plate material promote retreat of the subducting plate material from the collision zone; and 2) the strong coupling between the crust and mantle lithosphere of the overriding plate results in more localized crustal deformation. It is worth mentioning that in most “real-world” cases deposition histories and the amount of sedimentation are better constrained than denudation rates and the amount of material eroded. Consequently, present models of continental collision account for the effects of sedimentation better than for erosion and the risk of underestimating the impact of sedimentation on collision is limited.

In these experiments we have focused on continental orogenesis as the tectonic regime where climate-tectonic interactions are studied. The two-dimensional nature of the mod-
els enables the surface-tectonic dynamics to be treated at a high numerical resolution. However, we recognize that out-of-plane sediment transport and tectonic motion will alter the results. Retreat/non-retreat of the mantle lithosphere is the deep tectonic observable that we have focused on for illustrative clarity of the interpretations. Other tectonic processes (topography, crustal structure, thermal evolution) are similarly strongly modified by varying surface deposition. The results demonstrate the initial qualifications of first-order significance of climate-controlled deposition and tectonic interactions.

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Chapter 5

Influence of viscosity pressure-dependence on deep lithospheric tectonics during continental collision
Abstract

Previous geodynamic models of continental collision show that the behavior of the continental lithosphere is strongly influenced by its rheology. We build on previous work by quantitatively investigating with numerical experiments the influence of the pressure-dependence of viscosity on the process of tectonic deformation during collision. The models demonstrate how the inclusion of viscosity pressure-dependence can quite substantially alter the style of continental mantle lithosphere deformation. At low activation volumes, the subduction style of mantle lithosphere deformation is dominant only for high convergence rates and low-to-moderate initial Moho temperatures. Increasing the activation volume of mantle material allows the subduction style of deformation to occur at all convergence rates studied in the experiments, at the expense of the subduction-drip and ablative-drip styles of deformation. At low activation volumes, high convergence rates, and high initial Moho temperatures the distributed pure-shear style of deformation occurs. With these same conditions, increasing the activation volume of mantle material produces an ablative subduction style of mantle lithosphere deformation. At low activation volumes, low convergence rate, and moderate to high initial Moho temperatures the mantle lithosphere prefers a convective removal style of deformation; increasing the activation volume here yields an ablative-drip and distributed pure-shear styles of deformation. The results demonstrate that inclusion of the pressure-dependence of viscosity—quite often neglected in lithosphere-scale geodynamic models—can be significant in modulating deformation of the deforming lithosphere.

5.1 Introduction

While the first-order behavior of oceanic plates at convergence is well-resolved, the deformation of convergent continental plates is still not well understood. The lack of seismic zones to delineate the plate at depth (e.g., as with a Wadati-Benioff zone) and the poten-
tial presence of a highly deformed thick crust can help obscure the nature of continental collision zones. Forward numerical models have been used to attempt to gain insight into some of the possible tectonic processes that occur at depth during continental collision, investigating some of the fundamental styles of lithospheric deformation and/or thermal-mechanical parameters influencing these styles.

Geodynamic models of continental collision suggest that the continental mantle lithosphere may deform by, for example: 1) lithospheric “dripping”; 2) delamination; or 3) continental subduction. In the lithospheric “dripping” (i.e., Rayleigh-Taylor gravitational instability) model (e.g., Houseman et al., 1981) the thickened part of the dense lower mantle lithosphere becomes gravitationally unstable and descends into the underlying sub-lithospheric mantle (S.L.M.) as a viscous Rayleigh-Taylor-gravitational instability. Numerical and analogue geodynamic experiments have shown that Rayleigh-Taylor-type instabilities of the lithosphere are controlled by the viscous rheology of the lithosphere (e.g., Buck and Toksöz, 1983; Lenardic and Kaula, 1995; Houseman and Molnar, 1997; Molnar et al., 1998; Houseman et al., 2000; Molnar and Houseman, 2004), horizontal shortening of the lithosphere (e.g., Conrad and Molnar, 1997; Molnar et al., 1998; Houseman et al., 2000), the presence of crustal material (e.g., Neil and Houseman, 1999; Houseman et al., 2000; Pysklywec and Cruden, 2004; Molnar and Houseman, 2004), crustal radioactive heat production (Pysklywec and Beaumont, 2004), the presence of a weak zone and its physical parameters (e.g., Billen and Houseman, 2004) and the top surface boundary conditions (e.g., Harig et al., 2008). Gogus and Pysklywec (2008a) showed that dripping continental mantle lithosphere produces: 1) symmetric topographic uplift above the downwelling; and 2) symmetric crustal shortening above the downwelling and crustal extension away from it.

Bird (1979) first proposed that an elongated conduit connecting the continental crust and the S.L.M. would permit the dense continental mantle lithosphere to peel away (i.e., delaminate) from the overlying crust, directly exposing the crust to upwelling S.L.M. Using
two-dimensional (2-D) numerical experiments of thermal convection with a viscous-plastic rheology, Morency and Doin (2004) showed that delamination initiates with localized thinning of the mantle where the Moho temperature is the highest (∼800°C). This localized thinning places S.L.M. in contact with continental lower crust, permitting portions of the mantle lithosphere to sink into the S.L.M. by the process of delamination or by progressive thermal-mechanical erosion of the mantle lithosphere. Gogus and Pysklywec (2008a) showed that delaminating continental mantle lithosphere produces a broad region of plateau uplift above the asthenospheric window and a localized zone of subsidence at the delamination hinge. Gogus and Pysklywec (2008b) suggest that the pattern of plateau uplift, syn-convergent extension, heating and lithospheric thinning observed at eastern Anatolia can be explained by mantle lithosphere delamination. Similarly, Gray and Pysklywec (2012a) suggest that a combined process of mantle lithosphere retreat and delamination produces a number of surface observables that satisfy those of the Himalayan-Tibetan orogen (e.g., mantle lithosphere architecture as defined by seismic analyses, topography of the western Tibetan Plateau, surface heat flow and syn-convergent extension in the southern portion of the orogen.) The experiments presented by Ueda et al. (2012) show that the process of continental lithosphere delamination is controlled by the collision rate, crustal/mantle rheology, age of the collision zone and partial melting in the mantle.

Deformation of the continental plates in the continental subduction model occurs by underthrusting/subduction of a portion of one continental plate beneath the other along a discrete shear zone. Some of the first attempts at modelling crustal deformation during continental collision showed that a mantle lithosphere subduction basal boundary condition produces asymmetrical crustal deformation (e.g., Willett et al., 1993; Willett and Beaumont, 1994; Beaumont et al., 1996) and topography (e.g., Willett, 1999a). Some studies show that different styles of mantle lithosphere deformation may occur during collision, including subduction/underthrusting, ablative subduction, slab-break off and Rayleigh-Taylor-type gravitational instability; and that these different styles of deforma-
tion are dependent on the rheology of the mantle lithosphere, convergence velocity and the continental lithosphere geotherm (Pysklywec et al., 2000; Pysklywec, 2001; Pysklywec et al., 2002). This includes the suggestion that the upper portions of the mantle lithosphere characterized as “plate-like” exhibit subduction/underthrusting while the viscous lower portion simultaneously undergoes distributed Rayleigh-Taylor instability (Pysklywec et al., 2002).

These modes of deformation (Rayleigh-Taylor-type instability, delamination, continental subduction) have variably been applied to continental collision regions around the globe; again, often having interpretations shaped by numerical geodynamic modelling. For example, the South Island continental collision zone, comprising part of the Australia-Pacific plate boundary in New Zealand, is a prototypical small-scale continent-continent collision. In the last \( \sim 6.4 \) Ma \( \sim 90 \) km of convergence has been accommodated across the center of the Alpine fault, whereas the southern and northern ends of the island have accommodated \( \sim 70 \) km and \( \sim 110 \) km, respectively (Walcott, 1998). It has been proposed that the South Island collision is accommodated through continental subduction (Beaumont et al., 1996), Rayleigh-Taylor-type instability (Molnar et al., 1999), or a combination of these (Pysklywec et al., 2002). Though deformation of the tectonic plates most often takes place at plate boundaries, continental plate interiors are also susceptible to non-rigid behavior and intra-plate shortening. For example, the Phanerozoic tectonic history of Canada’s High-Arctic represents a complex and enigmatic history of collision, rifting and intra-plate deformation. Beginning in latest Devonian to earliest Carboniferous times, development of the Palaeozoic Franklinian basin was terminated by the Ellesmerian Orogeny. Development of the Sverdrup Basin followed the Ellesmerian Orogeny. Subsequently, \( \sim 200 \) km of shortening occurred during the Eocene intra-plate Eurekan Orogeny that affected parts of the Franklinian and Sverdrup Basins (reviewed in Trettin, 1989).

In this study we build on previous work to explore the styles of deep continental lithosphere deformation during the initial development of a continent-continent collision.
or intra-plate shortening, and specifically the role of pressure-dependence of viscosity on modifying collisional models. It was recently shown that rheological results for olivine determined at pressures below \( \sim 0.5 \) GPa cannot be extrapolated to pressures greater than several GPa because at these pressures the pressure effect is large (e.g., Karato, 2010; Kawazoe et al., 2009). Consequently, these ‘low-pressure results’ cannot be used to explore upper mantle deformation at depths greater than \( \sim 20 \) km. That said, technical difficulties have historically hindered the capacity of experimental studies to accurately obtain quantitative data on the rheology of olivine at high pressures (reviewed in Karato, 2010). For example, Karato (2010) describes how the value of olivine’s activation volume in these experimental studies ranged from 0 to \( 27 \times 10^{-6} \) m\(^3\) mol\(^{-1}\). This uncertainty in the activation volume of olivine corresponds to \( \sim 10 \) orders of magnitude difference in viscosity given constant stress. That said, owing to the development of new types of deformation apparatus (e.g., the rotational Drickameter) there now exists a better understanding of the deformation of olivine at high pressures from experimental results (e.g., Kawazoe et al., 2009).

In light of these findings it is essential to investigate and understand the contribution from the pressure-dependence of viscosity on creeping flow. Here, we present a series of geodynamic experiments that test the sensitivity of continental mantle lithosphere deformation to varying activation volumes of mantle material. We explore this sensitivity in the presence of a range of modelled lithospheric convergence rates and initial Moho temperatures so that the experiments yield insight into a range of tectonic environments.

5.2 Methodology

We use the SOPALE geodynamic code (Fullsack, 1995) to solve the coupled conservation equations of mass, momentum and internal energy that govern the behavior of plane-strain (i.e., 2-D), viscous-plastic incompressible media. SOPALE is based on an arbitrary Lagrangian-Eulerian finite element technique. Here, the calculations are performed on an
Eulerian finite element grid (vertical x horizontal resolution = 157x401), whereas the La-
grangian grid (vertical x horizontal resolution = 469x1201) is advected with the computed
velocity field and in turn is used to update the evolving material and thermal property dis-
tributions on the Eulerian grid. The Eulerian grid is restricted to small vertical dilations
corresponding to the development of topography on the free top surface; the Lagrangian
elements are freely deformable. Plastic deformation is modeled with a pressure-dependent
Drucker-Prager criterion. A thermally activated power-law creep relation is used to solve
for the effective viscosity of materials undergoing viscous flow.

We study the influence of viscosity pressure-dependence by varying the activation
volume term ($V^*$) of mantle material in the power-law creep equation; e.g.:

$$
\eta^v_e = (3^{-(1+n)/2n}2^{(1-n)/n})A^{(-1/n)}(I_2'/(1-n))^{2n}e^{\exp(\frac{Q + PV}{nRT})},
$$

(5.1)

where $\eta^v_e$ is the effective viscosity, $I_2'$ is the second invariant of the deviatoric strain rate
tensor, T is the temperature and the variables A, n, Q, P, and V are the material parameter,
power exponent, activation energy, pressure, and activation volume, respectively (e.g.,
Gerya, 2009).

Continental collision is imposed as a kinematic boundary condition by introducing new
lithosphere into the right edge of the box at horizontal velocities between 1.0 and 5.0 cm
a$^{-1}$, depending on the experiment, while lithosphere at the left edge of the box is held
fixed (Figure 5.1). A small outward flux is distributed evenly along the sides of the S.L.M
to balance the injection of lithosphere into the solution space. The models do not include
a previous phase of ocean lithosphere subduction. Given the temperature-dependence of
viscosity, we recognize that a previous phase of ocean lithosphere subduction can modify
the strength of the continental lithosphere and its propensity for deformation. Our ex-
periments are designed to model the collision of continental blocks that were already in
contact prior to the onset of collision—for example, at the South Island collision zone or in
the case of an intra-plate orogen. The mechanical boundary conditions are defined by a
### Table 5.1: Physical parameters for reference experiment

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Upper crust</th>
<th>Lower crust</th>
<th>Mantle lithosphere</th>
<th>Sub-lithospheric mantle</th>
<th>Weak zone</th>
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<tbody>
<tr>
<td>Ref. density (kg m(^{-3}))</td>
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<td>2900</td>
<td>3250</td>
<td>3250</td>
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<td>Ref. density T (K)</td>
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<td>293</td>
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<td>1609</td>
<td>293</td>
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<td>(\phi_1) (deg.)</td>
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<td>15</td>
<td>15</td>
<td>15</td>
<td>5</td>
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<tr>
<td>(\phi_2) (deg.)</td>
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<td>2</td>
<td>2</td>
<td>1</td>
</tr>
<tr>
<td>((I_2^1)^{1/2})</td>
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<td>0.5</td>
<td>0.5</td>
<td>0.5</td>
<td>0.5</td>
</tr>
<tr>
<td>((I_2^2)^{1/2})</td>
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<td>1.5</td>
<td>1.5</td>
<td>1.5</td>
<td>1.0</td>
</tr>
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<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Flow law</td>
<td>Wet</td>
<td>Diorite</td>
<td>Dry olivine</td>
<td>Dry olivine</td>
<td>-</td>
</tr>
<tr>
<td>Reference</td>
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<td>2</td>
<td>3</td>
<td>3</td>
<td>-</td>
</tr>
<tr>
<td>(f)</td>
<td>1</td>
<td>10</td>
<td>1</td>
<td>1</td>
<td>-</td>
</tr>
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<td>(A) (Pa(^{-n})s(^{-1}))</td>
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<td>(5.17 \times 10^{-18})</td>
<td>(4.85 \times 10^{-17})</td>
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<td>3.5</td>
<td>3.5</td>
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<tr>
<td>(Q) (kJ mol(^{-1}))</td>
<td>223</td>
<td>219</td>
<td>535</td>
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<tr>
<td>(V) (m(^3) mol(^{-1}))</td>
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<td>0</td>
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<td>750</td>
</tr>
<tr>
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<td>2.25</td>
<td>2.25</td>
<td>2.25</td>
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<tr>
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<td>(3.0 \times 10^{-5})</td>
<td>(3.0 \times 10^{-5})</td>
<td>(3.0 \times 10^{-5})</td>
</tr>
<tr>
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<td>0.75 (\times 10^{-6})</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>

1) Gleason and Tullis (1995); 2) Ranalli (1997); 3) Hirth and Kohlstedt (1996); Kawazoe et al. (2009)

free top surface and “free-slip” (i.e., zero tangential stress) at the other three sides. These models do not include surface processes (i.e., surface erosion and sedimentation). Though surface processes clearly play a role in the style of continental lithosphere deformation during continental and intra-plate collision (e.g., Beaumont et al., 1992; Willett, 1999a; Batt and Braun, 1997; Koons, 1990; Simpson, 2006; Stolar et al., 2007; Willett, 1999a; Avouac and Burov, 1996; Burov and Toussaint, 2007; Yamato et al., 2008; Pysklywec, 2006; Gray and Pysklywec, 2012b), the intention of this contribution is to investigate the effects of the pressure-dependence of viscosity on the different styles of continental lithosphere deformation during intra-plate collision without having to distinguish its importance relative to surface processes.
Figure 5.1: a) Initial configuration of the numerical experiments. In the models, a 120 km thick continental lithosphere- made up of a 24 km thick upper-crust ($\rho_o = 2700$ kg $m^{-3}$), a 12 km thick lower-crust ($\rho_o = 2900$ kg $m^{-3}$) and an 84 km thick mantle lithosphere ($\rho_o = 3250$ kg $m^{-3}$)- overlies 480 km of S.L.M. ($\rho_o = 3250$ kg $m^{-3}$). Power-law viscous parameters for the upper crust are based on the experimental results of wet quartzite (Gleason and Tullis, 1995). Those for the lower crust are based on the experimental results of quartz diorite (Ranalli, 1997) (Table 5.1). Effective viscosity of the quartz diorite lower crust are scaled up by a factor of 10 to reproduce stronger, more mafic conditions. Power-law viscous parameters for the mantle lithosphere and the S.L.M. are based on the experimental results for dry olivine (Hirth and Kohlstedt, 1996; Kawazoe et al., 2009). The frictional plastic rheology of the upper crust, lower crust, mantle lithosphere and S.L.M. is permitted to soften by implementing a linear decrease in $\phi_{eff}$ with accumulated strain, $(I'_2)^{1/2}$. Through these strain softening mechanisms, the frictional strength of the upper crust, lower crust, mantle lithosphere and S.L.M. is reduced from $\phi_{eff} = 15^\circ$ to $2^\circ$ through the strain range $0.5 < (I'_2)^{1/2} < 1.5$. The initial geotherm for the experiments is laterally uniform and is defined by a surface temperature of 20$^\circ$C, an increase to 475$^\circ$C, 550$^\circ$C, 625$^\circ$C or 700$^\circ$C, depending on the experiment, at the Moho; an increase to 1350$^\circ$C at the base of the mantle lithosphere and an increase to 1520$^\circ$C at the bottom of the model. The upper crust and lower crust have radioactive heat production rates of 2.0 $\mu$W $m^{-3}$ and 0.75 $\mu$W $m^{-3}$ (e.g., Beaumont et al., 2004), respectively. A weak zone ($\phi_{eff} = 5^\circ$ to $1^\circ$ through the strain range $0.5 < (I'_2)^{1/2} < 1.0$) of dimension 10 km by 10 km, is inserted into the top 10 km of the mantle lithosphere to localize the initial deformation in the center of the model.
5.3 Results

5.3.1 Experiments with a Moho temperature of 475°C

Figure 5.2 shows the results of three experiments (EXP1-3), after 250 km of imposed convergence ($\Delta x = 250$ km). Experiments EXP1-3 are characterized by a convergence velocity of $1 \text{ cm a}^{-1}$ and an initial Moho temperature of $475^\circ \text{C}$. They differ only in the magnitude of the activation volume for mantle material. Fig. 5.2a ($V = 0 \times 10^{-6} \text{ m}^3 \text{ mol}^{-1}$) illustrates the results of experiment EXP1. After $\Delta x = 250$ km, the mantle lithosphere is characterized by a mixed style of deformation and, overall, exhibits an ablative-drip style of mantle lithosphere deformation (Pysklywec, 2001). Initially, imposed convergence is accommodated by subduction of strong retro-side mantle lithosphere and lower crust beneath pro-side mantle lithosphere along a discrete shear zone. Between $\Delta x = 150$ and 162.5 km (not shown) a Rayleigh-Taylor-type gravitational instability develops and the lower portion of the subducting retro-side mantle lithosphere drips quickly ($\sim 5 \text{ cm a}^{-1}$) into the underlying S.L.M. By $\Delta x = 250$ km the upper mantle lithosphere does not accommodate convergence by underthrusting/subduction of retro-side material beneath pro-side mantle lithosphere. Rather, the retro- and pro-side mantle lithospheres are colliding in a form of ablative subduction (Tao and O’Connell, 1992). Fig. 5.2b (EXP2; $V = 5 \times 10^{-6} \text{ m}^3 \text{ mol}^{-1}$) illustrates the drastic effect that a relatively small increase in activation volume of mantle material can have on the behavior of an otherwise identical system. In EXP2 the increased activation volume leads to increased mantle viscosities and a considerably more stable mantle lithosphere. As in EXP1, the initial stages of convergence are accommodated by the development of a single shear zone and the subduction of pro-side mantle lithosphere and lower crust beneath the overriding plate. Owing to the increased viscosity of the mantle the growth of the Rayleigh-Taylor-type instability is slower in EXP2 than in EXP1. The most striking difference between EXP2 and EXP1 is the continued propensity of pro-side mantle lithosphere and lower crust to
subduct beneath the overriding plate. By $\Delta x = 250$ km the subducting plate has not been damaged because of the relatively low strain rates in the mantle associated with an increased activation volume.

Fig. 5.2c (EXP3; $V = 15 \times 10^{-6}$ m$^3$ mol$^{-1}$) further illustrates this effect. The initial stages of convergence in EXP3 are characterized by subduction of pro-side mantle lithosphere and lower crust beneath retro-side material along a discrete shear zone. By $\Delta x = 250$ km, EXP3 shows that the viscosity of material at the base of the mantle lithosphere is $\sim 5.0 \times 10^{21}$ Pa·s. This hinders the tendency of the material at the base of the subducting mantle lithosphere to drip into the underlying S.L.M. That said, there is some evidence for the onset of Rayleigh-Taylor instability of the lower mantle lithosphere: a small amount of subducting mantle lithosphere is entrained at the base of the thickened mantle lithosphere root.

5.3.2 Experiments with a Moho temperature of 550°C

Figure 5.3 shows the results of experiments EXP4 (Fig. 5.3a) and EXP5 (Fig. 5.3b) at $\Delta x = 250$ km. They are characterized by a convergence velocity of 1 cm a$^{-1}$ and an initial Moho temperature of 550°C. Following the philosophy of the previous suites of experiments, EXP4 and EXP5 differ only in the magnitude of the activation volume for mantle material. Fig. 5.3a ($V = 10 \times 10^{-6}$ m$^3$ mol$^{-1}$) illustrates that the mantle lithosphere in EXP4 is characterized by the ablative-drip style of deformation. An initial Moho temperature of 550°C results in warm temperatures at the top of mantle lithosphere, placing this region almost entirely within the viscous regime. Since the upper mantle is a strong viscous fluid strain softening does not occur and the mantle lithosphere does not accommodate convergence by the development of a shear zone where coherent underthrusting of one plate beneath the other occurs. As in EXP1, the retro- and pro-side mantle lithospheres are ablatively subducting (Tao and O’Connell, 1992). An increase in activation volume ($V = 15 \times 10^{-6}$ m$^3$ mol$^{-1}$) causes the ablative-drip style of deformation to transition into
Figure 5.2: Evolution of thermal-mechanical numerical experiments a) EXP1 (V = 0 × 10^{-6} m^3 mol^{-1}), b) EXP2 (V = 5 × 10^{-6} m^3 mol^{-1}) and c) EXP3 (V = 15 × 10^{-6} m^3 mol^{-1}) after cumulative convergence of 250 km (Δx = 250 km). Experiments EXP1-3 are characterized by a convergence velocity of 1 cm a^{-1} and an initial Moho temperature of 475°C. The main frames of each experiment show material regions as differently colored regions (as labeled in Figure 5.1). The Lagrangian mesh is superimposed on the material fields to show internal deformation. The inset frames to the right of the main frames show filled contours for log of effective viscosity, log(η_v), for the region within the dashed box. The inset frames to left of the main frames show material in the plastic regime (black) and viscous regime (white) for the region within the dashed box.
the subduction style of deformation (EXP5; Fig. 5.3b). It is important to note that the subduction-drip deformation is absent from this transition. The mantle’s increased activation volume allows for the upper-most portion of the mantle lithosphere to be within the frictional-plastic regime. Hence, the convergence is accommodated by underthrusting/subduction of the pro-side mantle lithosphere beneath that of the retro-plate. The activation volume increase also produces effective viscosities at the base of the mantle lithosphere of $\sim 10^{22}$ Pa·s. Except for the region near the base of the thickened mantle lithosphere, the growth of the Rayleigh-Taylor instability is too slow for the development of mantle lithosphere drips.

![Figure 5.3](image)

**Figure 5.3:** Evolution of thermal-mechanical numerical experiments a) EXP4 ($V = 10 \times 10^{-6}$ m³ mol⁻¹) and b) EXP5 ($V = 15 \times 10^{-6}$ m³ mol⁻¹) at $\Delta x = 250$ km. Experiments EXP4 and EXP5 are characterized by a convergence velocity of 1 cm a⁻¹ and an initial Moho temperature of 550°C.

### 5.3.3 Experiments with a Moho temperature of 625°C

Figure 5.4 shows the results of experiments EXP6-8 at $\Delta x = 250$ km. EXP6-8 are characterized by a convergence velocity of 1 cm a⁻¹ and an initial Moho temperature of 625°C. They differ only in the magnitude of the activation volume for mantle material. Experiment EXP6 (Fig. 5.4a; $V = 5 \times 10^{-6}$ m³ mol⁻¹ for mantle material) exhibits the
convective removal style of deformation first discussed in Houseman et al. (1981). At $\Delta x = 250 \text{ km}$ the mantle lithosphere in this experiment is characterized by a Rayleigh-Taylor-type gravitational instability. The high geothermal gradient in the crust warms the mantle lithosphere, lowering its effective viscosity. Consequently, upper mantle lithosphere material does not behave in the frictional-plastic regime and the weak zone does not permit the development of strain softening at all. Rather, during the initial stages of the model's evolution the weak zone allows for a small amount of localization of deformation in the viscous regime. This causes a small amount of lithosphere thickening in response to convergence, setting the stage for the development of a Rayleigh-Taylor-type gravitational instability of the thickened mantle lithosphere.

In EXP7 (Fig. 5.4b) $V = 10 \times 10^{-6} \text{ m}^3 \text{ mol}^{-1}$ for mantle material. This leads to higher viscosities in the mantle and, as a result, lower strain-rates. As in EXP6, the upper mantle lithosphere does not behave in the frictional plastic regime. The upper mantle lithosphere is a strong viscous fluid. Because deformation cannot strain soften in the viscous regime, the retro- and pro-side mantle lithospheres are ablatively subducting (Tao and O’Connell, 1992). Though the initial development of EXP7 is different than that in EXP1 (i.e., no initial shear zone in EXP7), we focus on the governing styles of deformation after $\Delta x = 250 \text{ km}$. Consequently, we consider the outcomes of EXP1 and EXP7 the same. Owing to the increased activation volume, the growth of the Rayleigh-Taylor-type instability at the base of the thickened mantle lithosphere root is slower in EXP7 than in EXP6.

EXP8 (Fig. 5.4c; $V = 15 \times 10^{-6} \text{ m}^3 \text{ mol}^{-1}$) further illustrates the effect of increasing the activation volume of mantle material. By $\Delta x = 250 \text{ km}$, the viscosity of material at the base of the thickened mantle lithosphere root is $\sim 10^{22} \text{ Pa} \cdot \text{s}$. This severely hinders the propensity of thickened mantle lithosphere to drip into the underlying S.L.M. Again, because of the high geothermal gradient in the crust, upper mantle lithosphere material in this suite of experiments does not behave in the frictional-plastic regime. The lack of strain softening in the mantle lithosphere prevents the incoming pro-side mantle lithosphere
from subducting beneath the overriding plate. Rather, a significant mantle lithosphere root develops below the weak zone, as in EXP7, where the retro- and pro-side mantle lithospheres are subducting ablatively.

Figure 5.4: Evolution of thermal-mechanical numerical experiments a) EXP6 (V = 5 × 10^{-6} m^3 mol^{-1}), b) EXP7 (V = 10 × 10^{-6} m^3 mol^{-1}) and c) EXP8 (V = 15 × 10^{-6} m^3 mol^{-1}) at Δx = 250 km. Experiments EXP6-8 are characterized by a convergence velocity of 1 cm a^{-1} and an initial Moho temperature of 625°C.

5.3.4 Experiments with a Moho temperature of 700°C

Figure 5.5 shows the results of experiments EXP9 (Fig. 5.5a) and EXP10 (Fig. 5.5b) at Δx = 250 km. EXP9 and EXP10 are characterized by a convergence velocity of 1 cm a^{-1} and an initial Moho temperature of 700°C. They differ only in the magnitude of the activation volume for mantle material. EXP9 (Fig. 5.5a; V = 5 × 10^{-6} m^3 mol^{-1}) illustrates the “convective removal” style of deformation. This style of deformation was previously discussed for EXP6, so we omit a detailed explanation of the model’s evolution.
The mantle lithosphere’s response to the increased activation volume in EXP10 (Fig. 5.5b; \( V = 10 \times 10^{-6} \text{ m}^3 \text{ mol}^{-1} \)) is higher viscosities and lower strain-rates. The combination of a very high initial Moho temperature of 700°C and a relatively high viscosity mantle does not permit strain localization of any kind to develop in the mantle lithosphere in EXP10. Consequently, imposed convergence is not accommodated in the mantle lithosphere by the development of a shear zone or a Rayleigh-Taylor-type instability, but by distributed pureshear thickening.

![Figure 5.5: Evolution of thermal-mechanical numerical experiments a) EXP9 (\( V = 5 \times 10^{-6} \text{ m}^3 \text{ mol}^{-1} \)) and b) EXP10 (\( V = 10 \times 10^{-6} \text{ m}^3 \text{ mol}^{-1} \)) at \( \Delta x = 250 \text{ km} \). Experiments EXP9 and EXP10 are characterized by a convergence velocity of 1 cm a\(^{-1}\) and an initial Moho temperature of 700°C.](image)

### 5.4 Discussion/Conclusions

Experiments EXP1-10 illustrate how the pressure-dependence of viscosity can control deep lithosphere processes; namely increasing the activation volume term in equation (1) produces more viscous mantle material and lower strain-rates. As a more comprehensive exploration of the parameter space, we ran a large suite of experiments (125 in total) that test the sensitivity of this model to varying activation volumes of mantle material, con-
vergence rates and initial Moho temperatures. Figure 5.6 plots the mode of deformation that the models evolve to (variably coloured dots) for the given rheological and velocity vs. Moho temperature state. The results illustrate that at low activation volumes ($V = 0 \times 10^{-6} \text{ m}^3 \text{ mol}^{-1}$), high convergence rates (5 cm a$^{-1}$) and low to moderate initial Moho temperatures (400-550°C) the subduction style of deformation dominates. Initially, increasing the activation volume ($V = 0-10 \times 10^{-6} \text{ m}^3 \text{ mol}^{-1}$) allows the subduction style of deformation to occur at all convergence rates, provided that the initial Moho temperature is low as well. This occurs at the expense of the subduction-drip style of deformation. With even higher activation volumes ($V = 10-20 \times 10^{-6} \text{ m}^3 \text{ mol}^{-1}$) the subduction style of deformation occurs at low convergence rates and moderate initial Moho temperatures ($\sim 550°C$) at the expense of the ablative-drip style of deformation. Essentially, increasing the activation volume of mantle material in experiments with a low to moderate initial Moho temperature (400-550°C) hinders the propensity of mantle lithosphere to drip into the underlying S.L.M. and allows the upper portion of the mantle lithosphere to behave in the frictional-plastic regime, favoring the development of a plastic shear zone.

At low activation volumes ($V = 0 \times 10^{-6} \text{ m}^3 \text{ mol}^{-1}$), high convergence rates (5 cm a$^{-1}$) and high initial Moho temperatures (625-700°C) the pure-shear style of deformation occurs. Increasing the activation volume of mantle material to $5-10 \times 10^{-6} \text{ m}^3 \text{ mol}^{-1}$ increases the viscosity of the mantle and yields an ablative style of mantle lithosphere subduction at moderate to high initial Moho temperatures ($\sim 625°C$) and low to high convergence rates (2-5 cm a$^{-1}$). This transition occurs because the increased activation volume produces a stronger viscous fluid that is unable to localize deformation in the plastic regime and strain soften. Consequently, the colliding retro- and pro-plates subduct ablatively.

At low activation volumes $V = 0 \times 10^{-6} \text{ m}^3 \text{ mol}^{-1}$, a low convergence rate (1 cm a$^{-1}$) and moderate to high initial Moho temperature (625-700°C) the mantle lithosphere exhibits the convective removal style of deformation first discussed in Houseman et al. (1981).
Figure 5.6: Illustration of the relationship between activation volume of mantle material ($V = 0-20 \times 10^{-6} \text{ m}^3 \text{ mol}^{-1}$), convergence velocity (1-5 cm a$^{-1}$), initial Moho temperature (425-700°C) and the style of mantle lithosphere deformation at $\Delta x = 250$ km for the complete suite of thermal-mechanical experiments carried out for this study.
Increasing the activation volume when the initial Moho temperature is high (700°C) produces the distributed pure-shear style of deformation. A corresponding activation volume increase for an initial Moho temperature of 625°C leads to the ablative-drip style of mantle lithosphere deformation. The combination of a very high initial Moho temperature of 700°C and a relatively stiff mantle favors the distributed pure-shear deformation style. At slightly cooler temperatures, the stiff mantle lithosphere subducts ablatively, whereas the low-viscosity thickened mantle root develops as a Rayleigh-Taylor-type gravitational instability and drips into the underlying S.L.M.

The experiments demonstrate the first sensitivity analysis of the effects of the pressure-dependence of viscosity on continental lithosphere deformation during collision. They show the effects of the pressure-dependence of viscosity may substantially alter the style of deep lithosphere deformation during collision. We recognize that temperature-dependence of viscosity is the more fundamental rheological control on tectonic behavior, easily yielding orders of magnitude viscosity variation through the temperature range in the lithosphere. However, given the capability of modern geodynamic computational codes, we advocate that it is important to include pressure-dependence of viscosity in the rheological formulation of these numerical tools. This is bolstered by the findings that the pressure-dependence of viscosity is significant and should be considered when extrapolating experimental rheological laws—which are derived at low pressures—to the high pressures characteristic of sub-crustal tectonics (Karato, 2010; Kawazoe et al., 2009). New experimental rheological results for olivine at high pressures (e.g., reviewed in Karato, 2010) provide constraints for these high pressure regimes studied by geodynamic modelling.

Convective removal, pure-shear thickening, ablative-drip, ablative-subduction, subduction-drip and subduction are the modes of deformation arising in our continental collisional models and as such are the tectonic styles that we have focused on for illustrative clarity of the interpretations. Other tectonic modes may obviously arise in parameter space different from what we considered. Furthermore, other observables (e.g., topography, crustal
structure, thermal evolution) are also modified by varying the activation volume of mantle material and a natural progression of this work is to explore in greater detail how these other tectonic process are also affected by the pressure-dependence of viscosity. Also, the 2D nature of the models enable the crust-mantle dynamics to be treated at a high numerical resolution. However, we recognize that in a three-dimensional Earth out-of-plane tectonic motion will alter the results.
Chapter 6

Principal conclusions and discussion
6.1 Principal conclusions

The conclusions of this thesis can be divided into four sections, each representing the conclusions of the four “research” chapters.

6.1.1 Geodynamic models of Archean continental collision and the formation of mantle lithosphere keels

Numerical geodynamic experiments identify three dominant styles of continental mantle lithosphere deformation under Neoarchean-like conditions. These are: 1) a pure-shear thickening style; 2) an imbrication style; 3) and an asymmetric “flat-subduction” style accompanied by localized exhumation of the mafic lower crust. The results demonstrate that the degree of coupling between the crust and the mantle lithosphere is a contributing factor in controlling the behavior of the mantle lithosphere during collision.

The pure-shear thickening style of deformation occurs when the temperature of the mantle lithosphere is sufficiently raised by moderate-high R.H.P. in a wet quartzite crust or felsic granulite lower crust. Owing to the temperature-dependence of viscosity, moderate-high R.H.P. in the crust permits a dominantly ductile distribution of deformation in the continental mantle lithosphere. Under this style of deformation plate-like mantle lithosphere extends to depths of $\sim 150$ km.

The imbrication style of deformation occurs when the degree of R.H.P. in the wet quartzite crust or felsic granulite lower crust is low to moderate and when the lower crust is rheologically weak (i.e., either wet quartzite or felsic granulite). Beneath the weak lower crust, the strong upper portion of the buoyant mantle lithosphere subducts beneath adjacent mantle lithosphere. Under this regime of deformation plate-like mantle lithosphere extends to depths of $\sim 150-200$ km.

The flat-subduction style of deformation occurs when the lower crust is rheologically strong (e.g., a mafic granulite). Under these conditions, the lower crust does not de-couple from the underlying mantle lithosphere to the extent that it would if it were more felsic. As
a result, the lithosphere deforms more as a single plate. Because R.H.P. is concentrated in
the dry granite upper crust, mantle lithosphere temperatures are relatively independent of
crustal R.H.P. Therefore, under the flat-subduction regime, plate-like mantle lithosphere
extend to depths greater than \( \sim 200 \) km.

The results demonstrate that an imbrication style of mantle lithosphere deformation
may evolve to the flat-subduction style of deformation after a possible crustal inversion
event. The modeling study may demonstrate that the emplacement of mafic rocks at
the base of the crust, via vertical tectonics, and the resulting coupling of the lower crust
with the mantle lithosphere, in conjunction with a secular increase in mantle lithosphere
density (Poudjom Djomani et al., 2001, e.g.), may have been a reorganizing force that
contributed to the development of modern-style subduction beneath collisional orogens.

6.1.2 Geodynamic models of mature continental collision: Evolution of
an orogen from lithospheric subduction to continental retreat/delamination

Numerical geodynamic experiments demonstrate that deformation of continental litho-
sphere during mature collision (\( \Delta x = 1800 \) km) is dependent on mantle lithosphere den-
sity, lower crustal strength, the presence of mineralogical phase changes in the lower crust-
and their associated density changes- and to mantle lithosphere’s yield stress.

The experiments demonstrate that \( \sim 475 \) km of convergence is accommodated by sub-
duction of pro-plate lower crust and mantle lithosphere beneath the retro-plate. Because
the cumulative density of the subducting mantle lithosphere and eclogitized lower crust is
greater than that of the S.L.M. the subducting material retreats \( \sim 100 \) km from the collision
zone. This allows the underlying S.L.M. to ascend, come into contact with the pro-plate’s
upper crust and be juxtaposed with the retro-plate’s upper crustal/lower crustal inter-
face. Given this juxtaposition, the ascending S.L.M. intrudes the retro-plate along the
upper crustal/lower crustal interface. This intrusion initiates and, in combination with
the cumulative density of the retro-plate’s lower crust and mantle lithosphere, sustains
retro-plate lower crust/mantle lithosphere delamination. The combined process of deep lithosphere retreat and delamination produces a ∼1000 km wide lithosphere window where upwelled S.L.M. supports plateau-like surface topography of ∼4.5-5.5 km.

If the reference densities of the mantle lithosphere and S.L.M. are the same or if the strength of the lower crust is too high, retreat of subducting pro-plate material is not fast enough to allow upwelling of S.L.M. Conversely, the presence of eclogitized mafic lower crust enhances the tendency of subducting mantle lithosphere to retreat from the collision zone. Eclogitizable mafic lower crust will also couple crustal strain and the corresponding topography to mantle lithosphere deformation. If the mantle lithosphere yield stress is low, crustal deformation is excessively localized at the plateau margins. This produces uniform topography in the plateau interior and higher topography at the plateau’s margins.

The combined process of mantle lithosphere retreat and delamination may explain the seismically defined mantle lithosphere configuration, surface heat flow, syn-convergent extension and long-wavelength surface topography of the western portion of the Himalayan-Tibetan orogen.

6.1.3 Influence of sediment deposition on deep lithospheric tectonics

Numerical geodynamic experiments demonstrate that sedimentation in the basins adjacent to a model orogen may significantly modify continental lithosphere behavior during continental plate collision. Specifically, sedimentation initiates downward motion of the overriding plate in the area neighboring the collision zone. This results in a greater degree of coupling between the colliding continental plates and de-coupling between the pro-plate’s lower crust and upper crust. Conversely, in the absence of sedimentation the overriding plate does not undergo the characteristic downward motion in the area neighboring the collision zone. Consequently, the colliding plates are less coupled and the pro-plate’s upper crust and mantle lithosphere are strongly coupled. Owing to the reduced degree of coupling between the colliding plates and the weight of the subducting
A suite of numerical experiments is presented that tests the sensitivity of this model behavior to varying convergence rates and deposition %s. This suite illustrates that the horizontal distance along which the overriding plate’s lower crust is eclogitized increases with increasing deposition %’s and decreasing convergence velocities. With increasing deposition %’s the downward motion of the overriding plate in the area neighboring the collision zone increases. This pushes the overriding plate’s lower crust further into the eclogite stability field. Decreasing the convergence velocity increases the time available for the gravitational instability of the overriding plate material to develop. Consequently, more of the overriding plate’s lower crust is present within the eclogite stability field. Subducting retro-plate material undergoes greater degrees of retreat with decreasing convergence rates and decreasing deposition %’s. As with the overriding plate, decreasing the convergence rate increases the time available for the gravitational instability of the subducting plate material to develop. Furthermore, the incoming plate is too slow to fill the lithospheric window created by the retreating retro-plate. Combined, these two process increase the tendency of subducting material to retreat from the collision zone given decreasing plate convergence. Decreasing the deposition % decreases the downward motion of the overriding plate in the area neighboring the collision zone and decreases plate coupling. This increases retro-plate retreat.

The second suite of numerical experiments tests the sensitivity of the model behavior to varying initial Moho temperatures and deposition %’s. This suite illustrates that increasing the initial MOHO temperature increases the eclogite stability area in the computational domain. This increases the horizontal distance along which the pro-plate’s lower crust is eclogitized. As in the previous suite of experiments, increasing the deposition % increases the downward motion of the overriding plate in the area neighboring the collision zone. This pushes the overriding plate’s lower crust further into the eclogite stability field. In general, increasing the initial Moho temperature increases the propensity of subducting
retro-plate material to retreat from the collision zone. This occurs because increasing
the initial Moho temperature decreases the viscosity of the upper crust and increases the
degree of de-coupling between the retro-plate upper crust and mantle lithosphere.

6.1.4 Influence of viscosity pressure-dependence on deep lithospheric
tectonics during continental collision

A suite of numerical geodynamic experiments demonstrates that the style of deep con-
tinental lithosphere deformation during continental plate collision is dependent on the
activation volume of mantle material (i.e., olivine). Given the range of activation volumes
of mantle material ($0-20 \times 10^{-6} \text{ m}^3 \text{ mol}^{-1}$), convergence velocities ($1-5 \text{ cm a}^{-1}$) and initial
Moho temperatures ($400-700^\circ \text{C}$) studied in this contribution, we identified 7 styles of con-
tinental mantle lithosphere deformation during collision: convective removal, pure-shear
thickening, ablative-drip, ablative-subduction, subduction-drip and subduction. At low
activation volumes, high convergence rates and low-moderate initial Moho temperatures,
subduction of continental mantle lithosphere is stable during plate collision. With increas-
ing activation volumes of mantle material, the subduction style of deformation becomes
dominant at low-moderate initial Moho temperatures and at all convergence velocities
modelled ($1-5 \text{ cm a}^{-1}$) at the expense of the subduction-drip and ablative-drip styles of
deformation.

At low activation volumes, high convergence rates and an initial Moho temperature
of $625^\circ \text{C}$ the pure-shear thickening style of continental mantle lithosphere deformation
is stable. Increasing the activation volume of mantle material results in the continental
mantle lithosphere adopting an ablative style of deformation. This occurs at the expense
of the pure-shear thickening style of deformation at convergence velocities between $2-5 \text{ cm}
\text{ a}^{-1}$ and the convective removal style of deformation at a convergence velocity of $1 \text{ cm a}^{-1}$.

Increasing the activation volume of mantle material when the convergence velocity is
low and the initial Moho temperature is high produces the pure-shear thickening style of
continental mantle lithosphere deformation at the expense of the convective removal style of deformation.

### 6.2 Discussion

The research presented in this thesis poses as many questions as it answers. As discussed in Chapter 2 of this thesis the tectonic processes that operated during the Archean are poorly understood and the processes responsible for the formation of thick Archean S.C.L.M. beneath Archean cratons remain elusive. An aspect of the models proposed to explain the formation of S.C.L.M. that remains enigmatic is the presence of a flat Moho within most Archean lithosphere. For example, results based on seismic studies suggest that the average Moho depth in the western Superior is $41\pm1$ km (Kay et al., 1999) and that in the Slave is $37 - 40$ km (Bank et al., 2000). This remarkable feature may be the result of mafic-ultramafic lower crustal material foundering into the underlying mantle during a thermal event in the period between crust formation and cratonization (Nui and James, 2002). Percival and Pysklywec (2007) investigated such a process. In their thermo-mechanical models, foundering of dense mafic lower crust into the underlying mantle caused inversion of a lithosphere cell and a $>40$ m.y. pulse of $>1000^\circ$C temperatures in the lower crust. The models, though, resulted in a Moho with a distinctive topography following the lithospheric inversion. I suggest that future geodynamic research on the formation of Archean S.C.L.M. should consider the issue of the relatively flat Moho beneath the western Superior and the Slave. Numerical geodynamic experiments could be used to investigate the potential roles that foundering of dense mafic-ultramafic material into the underlying mantle and small scale convection erosion in the lower crust may have played in generating a flat Moho.

A consistent problem during the investigation of the dynamics of mature continental collision was the stability of the free-surface. None of the experiments presented in Chapter 3 permit plastic strain-softening to occur in the upper crustal layer because of the tendency of the free-surface to develop node by node oscillations under this condi-
tion. This instability is indicative of an unstable system and characteristic of linear and non-linear flow. Because of its inability to strain-soften the upper crust of the subducting plate (pro-plate) is not permitted to participate in subduction. Rather, this upper crustal layer acts as a plow and “bulldozes” the upper crust of the over-riding plate (retro-plate) retro-ward. In natural systems (e.g., the Himalayan-Tibetan orogen) a portion of the upper crust of the subducting plate may participate in subduction and the entire upper crust of the subducting plate does not act as a plow (e.g., Nabelek and the HI-CLIMB Team, 2009). A natural progression of the research presented in this thesis is to stabilize the free-surface and investigate mature collision when the upper crust is allowed to strain-soften. However, this is not a trivial task because it necessitates substantial modifications to the finite element discretization of the Stokes equations (Kaus et al., 2009).

The objective of the research presented in Chapter 3 was to understand the most basic physics behind the development of an evolved orogen (i.e., not to reproduce the tectonic evolution of the Himalayan-Tibetan orogen). For example, the initial set-up of the numerical experiments was the simplest model design consistent with first-order features of continental plate collision. In order to better understand the physical processes taking place in evolved orogens I suggest that more sophisticated and “realistic” numerical experiments, that build upon the work presented in Chapter 3, should be conducted in the future. More realistic experiments may include: 1) a previous phase of ocean subduction; 2) continental plates with different thermal histories; 3) continental plates with different mantle lithosphere density structures; and 4) collision with a variable convergence velocity history. The increased sophistication of the experiments will allow us to explore the physics of continental plate collision in more detail without the constraint of simulating particular natural settings.

An underlying assumption in all geodynamic experiments that investigate the effects of surface processes on continental plate collision is that different crustal lithologies have identical erodibilities. Invoking material-dependent erodibility, Schlunegger and Simpson
(2002) suggest that the shift from vertically directed exhumation to lateral growth of the Central Alps in the mid-to-late Miocene was in response to exposure of low-erodibility crystalline rock in the Alpine hinterland. In contrast, Rosenberg and Berger (2009) suggest that lateral growth of the Central Alps was caused by changes in rheology and the configuration of the deep crust. Future experiments that: 1) include material-dependent erodibility; and 2) systematically explore changes in deep crust rheology and configuration during collision will improve our understanding of the process of deep-crust exhumation during collision and its applicability to mid-to-late Miocene lateral growth of the Central Alps.

In all the experiments presented in this thesis thermal diffusivity ($\kappa$) is constant. Whittington et al. (2009) recently showed that $\kappa$ is temperature-dependent and that it decreases from $1.5-2.5 \times 10^{-6}$ m$^2$ s$^{-1}$ to $\sim 0.5 \times 10^{-6}$ m$^2$ s$^{-1}$ over the temperature range 250 - 850 K. Their one-dimensional thermal models of doubly thickened crust show that a temperature-dependent thermal diffusivity produces maximum lower crustal temperatures of $\sim 800$ °C. Assuming a constant strain rate of $10^{-14}$ and using the power-law creep parameters of dry Maryland diabase (Mackwell et al., 1998), $\sim 800$ °C lower crustal material will have a viscosity of $2 \times 10^{20}$ Pa·s. Lower crust with such a low viscosity is unlikely to participate in subduction. Rather, it is more likely that this lower crustal material will be scrapped off and accumulate above the subducting mantle lithosphere. Depending on the amount of convergence, if lower crust does not subduct and undergo eclogitization the subducting mantle lithosphere may not retreat from the collision zone. Therefore, crustal material with a temperature-dependent thermal diffusivity may exert a strong control on mantle lithosphere behavior during collision. Future work should investigate the sensitivity of continental plate collision to this temperature-dependence.

Finally, it is important to highlight that continental collision is modelled in all the experiments presented in this thesis by employing kinematic velocity boundary conditions. The use of these velocity boundary conditions neglects a consideration of the forces
required to drive continental collision. These driving forces must: 1) be greater than
the mechanical resistance of the lithosphere to deformation; and 2) work against gravity
because of the gravitational potential energy stored in the excess topography that is char-
acteristic of active collisional orogens (reviewed in Schellart, 2010). Copley et al. (2010)
show that the magnitude of the force exerted on the Indian plate by the Tibetan plateau
is $\sim 2 \times$ greater than the magnitude of the force exerted on the Indian ocean sea-floor
by ridge-push. Consequently, in the Himalayan-Tibetan orogen ridge-push alone cannot
drive continental collision. However, the total ridge length of the Indo-Australian plate is
much larger than the arc-length of the Himalayan-Tibetan orogen and the net subduction
zone force exerted on the Indian plate is roughly equal in magnitude to that of ridge-push
(Copley et al., 2010). Future work that investigates the relationship between continent
width, subduction zone length, ridge length and compositional variations in continental
and oceanic lithosphere in self-consistent 3D models will improve our understanding of the
forces that drive continental collision.
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