Numerical modelling of continental collision with an application to the Zagros fold-and-thrust belt

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To my beloved wife, Fedra, and my parents, Ghasem and Zahra,
who have been always there through the hard times.
Persian calligraphy by Alireza Kadkhodaei.

Poem by Omar Khayyam:

As the union of the elements animates you but a moment / Go! Enjoy life in spite of destiny’s tyranny

Live wisely since the essence of your body / is a little of earth and draught
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1. Introduction
1.1 Introduction to the thesis

1.1.1 Convergent plate boundaries

The relatively rigid outer layer of the Earth, called the lithosphere, is divided into fifteen tectonic plates (Fig. 1.1). These plates may be oceanic in character, or continental, or a combination of both. Oceanic lithosphere consists of a thin (about 8 km thick) mafic crust (basalt), and a lithospheric mantle. The thickness of oceanic lithosphere increases with age, but its average thickness is about 100 km. The oldest oceanic lithosphere is about 200 million years old (Müller et al., 2008), which indicates relatively fast construction and destruction of oceanic lithosphere. Continental lithosphere is typically older, with cratonic kernels of a few billion years old, suggesting that these managed to stay at the surface of the Earth over a long period of time. Continental lithosphere is thicker than oceanic lithosphere, with an average thickness of 120-150 km. It consists of a felsic crust (about 40 km thick, e.g. Turcotte and Schubert, 2002; Christensen and Mooney, 1995), of mainly granite and andesite, on top of the lithospheric mantle. These rigid plates lie on a warmer layer, which can flow over long time-scales (few million years) and loose heat by both conduction and convection. This layer is known as the asthenosphere and reaches...
to the transition zone at ca. 410 km depth (olivine-spinel phase change) (Turcotte and Schubert, 2002). A phase change at ca. 660 km, caused by the transformation of spinel to perovskite and magnesiowustite, separates the upper mantle from the lower mantle (Turcotte and Schubert, 2002). Tectonic plates move relative to each other on top of the asthenosphere with some centimetres per year. According to the relative movement of the plates, the interactions along their boundaries can be categorised in three different groups. 1) At divergent boundaries, plates move away from each other, while hot mantle material ascends and fills the gap between. The hot material cools gradually at the surface and accretes oceanic lithosphere to the plates. Examples are the Mid-Atlantic Ridge and the Southeast Indian Ridge. 2) At convergent boundaries, plates move towards each other. This motion is usually accommodated by one plate sliding under the other into the mantle. This process is called subduction. 3) At transform boundaries, plates slide along each other. The San Andreas is a well known example of a transform boundary, which accommodates lateral sliding between the Pacific and North American plates.

This thesis focuses on dynamic processes at convergent plate boundaries in the last stages of their evolution. Convergent boundaries are the most geologically active places on Earth. They are the sites of earthquakes, volcanism, and mountain building. In addition, they are associated with strategic elements such as gold, silver, uranium, and diamond. These minerals are believed to form through hydrothermal fluids that accompany rising magma (Barley et al., 1989; Landtwing et al., 2005). Taken together with the occurrence of destructive earthquakes, which result in massive damages and life loss, an understanding of processes at convergent boundaries is therefore also of social relevance.

Continental plates are positively buoyant and have a tendency to stay at the surface of the Earth (McKenzie, 1969), whereas oceanic plates older than 30-80 million years are negatively buoyant and gravitationally unstable at the surface (Cloos, 1993). Therefore, three possible scenarios can occur at convergent boundaries, depending on the character of the plates involved. [1] When two oceanic plates meet each other, the older and cooler plate has more negative buoyancy and will usually subduct under the younger plate. The Mariana trench is an example of this type of interaction, where the Pacific plate subducts under the Mariana plate. Oceanic crust contains water because of hydrous minerals such as amphibolite and clay (Mevel et al., 1978; Rüpke et al., 2004). Subduction of oceanic crust, can therefore, transport water to lithospheric depths. Loss of water from the subducting oceanic slab due to dehydration reactions results in lowering of the melting temperature of the overlying asthenosphere and facilitates partial melting of mantle material which rises to the surface and forms an arc-shaped volcanic chain on the overriding plate known
as an island arc (Fig. 1.2a) (Ringwood, 1974; Syracuse and Abers, 2006). The
distribution of shallow to deep earthquakes along the cold subducted slab (Wadati-
Benioff zone) outlines the slab from its adjacent mantle (Benioff, 1954; Barazangi
and Isacks, 1976). Seismic tomography offers another method to obtain information
on the shapes of subducted plates (Hilst et al., 1991; Grand, 2002; Zhao, 2004). Such
images have shown slabs that plunge straight into the mantle, down to lower mantle
depths, or feel the resistance of the endothermic 660 km phase change and bend for-
ward or backwards in the upper mantle. [2] If an oceanic plate converges towards

![Figure 1.2: Cartoon of different convergent plate boundaries, classified by the character of the plates involved: a) Ocean-ocean subduction, b) Continent-ocean subduction, c) Continent-continent collision.](image)

a continental plate, the oceanic plate subducts underneath the light continent. This
type of plate boundary is also accompanied by volcanic activity on the overriding
continent and shallow to deep seismicity. But it also accompanied by the formation
of mountains. These boundaries are known as Andean-type margins, named after
the mountains on the west coast of South America, where the oceanic Nazca plate
subducts under the continental South American plate, along the Peru-Chile trench
(Fig. 1.2b). [3] When two continents meet at a convergent boundary, they will col-
lide because their crusts are too buoyant to subduct deeply (Fig. 1.2c) (McKenzie,
1969). Although the convergence velocity of continents is low (few millimetres to
few centimetres per year) the plates have a considerable amount of mass and carry a
large momentum. Their collision results in extensive deformation, accompanied by
orogeny, folding, faulting, minor volcanic activity, and exhumation of (Ultra-) High
Pressure (U)HP rocks from large depths (Dewey and Burke, 1973; Molnar and Tap-
ponnier, 1975; Pfiffner et al., 1990; Andersen et al., 1991; Leech et al., 2005). The
Himalaya, European Alps, Norwegian Caledonides, and Zagros are typical exam-
ples of continental collision systems. In this thesis, we investigate the dynamics of
1.1.2 Why numerical modelling?

Although geological and geophysical observations provide us with valuable information about the composition, structure, and dynamics of the Earth’s interior, they are limited in both time and space. Many processes in the Earth, such as subduction, continental collision, and mantle convection, occur over long periods of time (thousands to millions of years) and are, therefore, beyond the human observation time-scale. In addition, parts of these processes occur at depths, which are too deep to be accessible for direct observation.

Modelling approaches (both analogue and numerical) can help us bridge these gaps in time and space by simulating geological processes in a quantitative way. Analogue modelling in geosciences is used to study the mechanics of natural large-scale systems with laboratory-scale models. In order to build an analogue model, the geometry and all model parameters need to be scaled down to the laboratory scale. Another approach to study the dynamics of geological systems is to solve equations that are thought to govern the processes of interest. Since solutions to these governing equations can often not be achieved analytically, computers can help to solve them numerically.

Modelling methods provide a powerful means to test hypotheses, help geological interpretations, and improve our understanding of the spatial and temporal distribution of physical quantities. Despite these advantages, we should keep in mind that modelling has its own limitations and restrictions. We list some of these difficulties here (A = analogue, N = numerical, and AN = analogue and numerical):

- Finding affordable materials with properly scaled physical parameters, especially for thermal problems. (A)
- Challenges in monitoring the temporal evolution of physical parameters. (A)
- Approximating Earth’s heterogeneity with simplified material distributions. (AN)
- Simplifications in the geometry of the system and the need to choose boundary and initial conditions for models. (AN)
- Approximating a continuum nature with a discrete model. (N)
- Employing simplified physical rules to approximate the natural process (governing equations). (N)
Even with these difficulties, modelling methods play an important role in transferring geosciences from a deductive to a predictive science. Over the few last decades, numerical modelling, among others, has been successfully used to study subduction zones (e.g. Minear and Toksöz, 1970; Peacock, 1990; Gerya et al., 2002; van Keken et al., 2008) and continent-continent collision systems (e.g. Burg and Podladchikov, 2000; Pysklywec et al., 2000; Toussaint et al., 2004; Andrews and Billen, 2009; Billen and Hirth, 2007; Warren et al., 2008; van Hunen and Allen, 2011; Duretz et al., 2012).

1.1.3 Aim of the thesis and outline

In this thesis, we investigate continental collision systems. Geological and geophysical observations show that deformation at continental collision zones can be accommodated in different styles: subduction of the continent, thickening of continental crust and lithosphere, lithospheric-scale folding, Rayleigh-Taylor (RT) instabilities, and slab break-off (Houseman and Molnar, 1997; Cloetingh et al., 1999; Burg and Podladchikov, 2000; Toussaint et al., 2004). Previous studies pointed out that convergence velocity, temperature at the base of the continental crust, and rheology of the lower continental crust can explain these different styles of collision (Pysklywec et al., 2000; Toussaint et al., 2004; Andrews and Billen, 2009; Burov and Yamato, 2008; van Hunen and Allen, 2011; Duretz et al., 2012). But also other factors and processes could be thought to control the style of deformation, such as, the rheology of the lithosphere and mantle, sedimentary stratigraphy, dip angle of the subducted oceanic slab, sedimentation and erosion, slab buoyancy, melting, phase changes, and the interaction of the system with adjacent plates. The aim of this thesis is to study continental collision systems and quantitatively investigate the effects of the driving conditions (magnitude and type), temperature, sedimentary stratification with embedded weak layers, and lithosphere rheology.

Chapter two reviews two examples of continent-continent collision systems: The Norwegian Caledonides and the Zagros fold-and-thrust belt. The Norwegian Caledonides are one of the best natural laboratories on Earth to study ultrahigh-pressure (UHP) rocks. Exposure of a vast area of UHP rocks in the Western Gneiss Region has led to speculations as to how this massive amount of rock could have been exhumed (Andersen et al., 1991; Wain, 1997; Roermund and Dury, 1998; Hacker et al., 2010).

The existence of Hormuz salt between the thick sedimentary cover and the crys-
talline basement in the SE Zagros is deduced from extruded salt diapirs at the surface, as well as from sedimentological information, and well data. Salt can act as a weak viscous layer, which decouples the sedimentary sequences from the basement, and controls the style of deformation in these regions (Bahroudi and Koyi, 2003; Yamato et al., 2011; Frehner et al., 2012; Mouthing et al., 2012). These two natural examples provide examples of parameters that may control the style of deformation in continental collision systems.

In chapter three, we investigate the effects of convergence rate, lithospheric temperature, rheology, and density differences between oceanic lithosphere and the sub-lithospheric mantle on the style of continental collision using numerical models. This chapter presents two type of models. In the first set-up, continents are driven by a combination of slab pull and a kinematic boundary condition applied at the side boundary of the lithosphere. This so-called 'kinematically driven collision' model represents a continent which is partly driven by surrounding plates. In the second set-up, called 'dynamic collision' model, the subducting continent is isolated from the surrounding plates and after an initiation phase, slab pull is the only driving force in the system. In the discussion, a simplified force balance is employed to explain the styles of deformation of collision systems. This chapter is published as Ghazian and Buiter (2013).

In chapter four, we use upper mantle-scale and upper crustal-scale models to investigate the effect of internal salt layers on crustal deformation in the SE Zagros. We use the upper mantle-scale models of continent-continent collision to better understand the large-scale tectonic picture of the region and to provide constraints for the higher resolution, upper crustal-scale models. We investigate the effects of basal and intervening salt layers, their strength and spatial distribution, as well as top basement dip angle. This chapter is submitted to Tectonophysics.

1.2 Numerical Method

1.2.1 Governing equations

Many processes in the Earth, such as subduction, mountain building, and mantle convection, occur over long periods of time (thousands to millions of years) and at slow rates (millimetres to a few centimetres per year). For these type of processes we can consider acceleration terms as negligible. In addition, a common approximation is to assume that deformation is associated with only small density changes and that we can, therefore, consider the materials as incompressible. The equation of mass
conservation then becomes:
\[
\frac{\partial v_i}{\partial x_i} = 0
\]  
(1.1)

\(v_i\) and \(x_i\) are the velocity and Eulerian coordinate vectors, respectively. The equation of motion is:
\[
\frac{\partial \sigma_{ij}}{\partial x_j} - \rho g_i = 0
\]  
(1.2)

\(\rho\) is density, \(g_i\) is gravitational acceleration \((g_x = 0, g_y = -9.8 \text{ m/s}^{-2})\), and
\[
\sigma_{ij} = \sigma'_{ij} - P \delta_{ij}
\]  
(1.3)

\(\sigma_{ij}\) is the total stress tensor, which consists of the deviatoric stress \(\sigma'_{ij}\) and the mean stress \(P\). \(\delta_{ij}\) is the Kronecker delta \((\delta_{ij} = 1\) for \(i = j\) and \(\delta_{ij} = 0\) for \(i \neq j\)). The mean stress (dynamic pressure) \(P\) is:
\[
P = \frac{1}{3} \sigma_{kk}
\]  
(1.4)

Pressure is positive for compression throughout this thesis. For viscous fluids, stress is related to strain-rate by:
\[
\sigma'_{ij} = 2\eta \dot{\epsilon}_{ij} + \frac{2}{3} \eta \dot{\epsilon}_{kk} \delta_{ij}
\]  
(1.5)

\(\dot{\epsilon}_{ij}\) is the strain-rate tensor, \(\dot{\epsilon}_{ij} = \frac{1}{2} (\frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i})\), and \(\eta\) is the viscosity of the material. The deviatoric strain-rate \(\dot{\epsilon}'_{ij}\) is related to the bulk strain-rate \(\dot{\epsilon}_{ij}\) through \(\dot{\epsilon}_{ij} = \dot{\epsilon}'_{ij} - \frac{1}{3} \dot{\epsilon}_{kk} \delta_{ij}\). For incompressible materials \(\dot{\epsilon}_{kk} = 0\) and, therefore, \(\dot{\epsilon}_{ij} = \dot{\epsilon}'_{ij}\) and \(\sigma'_{ij} = 2\eta \dot{\epsilon}_{ij}\). Substitution of (1.3) into (1.2) gives the equation of conservation of momentum as:
\[
\frac{\partial \sigma'_{ij}}{\partial x_j} - \frac{\partial P}{\partial x_j} - \rho g_i = 0
\]  
(1.6)

This equation can be rewritten in terms of strain-rate rather than deviatoric stress as:
\[
\frac{\partial (2\eta \dot{\epsilon}_{ij})}{\partial x_j} - \frac{\partial P}{\partial x_j} - \rho g_i = 0
\]  
(1.7)

The system of equations for mechanical problems consist of equations (1.1) and (1.7), which can be solved using the iterative penalty method (Cuvelier, 1986; Pelletier et al., 1989; Zienkiewicz and Taylor, 2000). In this method, the continuity equation (1.1) is perturbed as following:
\[
P^{\text{new}} = P^{\text{old}} - \kappa \frac{\partial u_i}{\partial x_i}
\]  
(1.8)
\( \kappa \) is the compressibility factor (generally taken 8 orders of magnitude larger than the largest viscosity in the model) (Zienkiewicz and Taylor, 2000). The perturbed continuity equation (1.8) and the momentum equation (1.7) are solved in an iterative way to minimise the norms of two successive velocity and pressure values to the desired tolerance.

Material properties, such as viscosity and density, can change with temperature. In the Boussinesq approximation, the assumption of incompressibility is slightly relaxed:

\[
\rho = \rho_0 (1 - \alpha(T - T_0)) \tag{1.9}
\]

\( \rho_0 \) is the density at the initial temperature \( T_0 \) and \( \alpha \) is the thermal expansivity. For such cases, the temperature evolution needs to be solved in addition to the mechanical system. The equation for conservation of energy is:

\[
\rho c_p \left( \frac{\partial T}{\partial t} + v_i \frac{\partial T}{\partial x_i} \right) = k \frac{\partial^2 T}{\partial x_i^2} + Q \tag{1.10}
\]

c\( _p \), \( T \), \( k \), and \( Q \) are specific heat at constant pressure, temperature, thermal conductivity, and heat production per volume, respectively.

### 1.2.2 Rheology

The behaviour of materials in response to geological forces can be categorised as viscous, brittle, and elastic. Brittle and elastic behaviour are in general more dominant at low temperatures and pressures (crustal depths), whereas viscous behaviour becomes more dominant at high temperatures (lithosphere and mantle depths).

#### 1.2.2.1 Viscous behaviour

Solid-state creep is the major deformation mechanism when the temperature of a material is a significant fraction of its melt temperature (Ranalli, 1995; Turcotte and Schubert, 2002). Solid-state creep can be categorised into two main types of creep in terms of stress status. At relatively low stresses \((\sigma \approx 10-100 \text{ MPa})\), flow occurs by migration of vacancies between grain boundaries. This type of deformation, in which the stress is linearly related to strain-rate (Newtonian), is known as diffusion creep. Dislocation creep is a dominant creep mechanism when the stress level is higher and is described by a power law relation between stress and strain-rate. Dislocations (defects and impurities in the crystal structure) move through the crystal
under this condition. Since a flow law expresses physical properties of the material, which are independent from the coordinate system, it is appropriate to write a flow law in terms of tensor invariants. Both diffusion and dislocation creep can generally be described by a non-linear effective viscosity:

\[ \eta_{II} = B^{\frac{1}{2}} A^{-\frac{n}{2}} \dot{\epsilon}_{II}^{\frac{1}{2}} A^{-\frac{1}{n}} \dot{\epsilon} \left( \frac{1}{n} - 1 \right) II d^n p^n f^n OH^{-r} e^{(E/PV)} \]  

(1.11)

The effective viscosity is \( \eta_{II} = \sigma_{II}' \dot{\epsilon}_{II}' \), where \( \sigma_{II}' \) is the effective shear stress and \( \dot{\epsilon}_{II}' \) is the effective strain-rate.

A is a constant, \( n \) the stress exponent (\( n = 1 \) for diffusion creep and \( n > 1 \) for dislocation creep), \( d \) grain size, \( p \) grain size exponent (\( p > 0 \) for diffusion creep and \( p = 0 \) for dislocation creep), \( f_{H_2O} \) water fugacity, \( r \) water fugacity exponent, \( E \) activation energy, \( V \) activation volume, \( R \) the gas constant, and \( T \) temperature in degrees Kelvin. It is more practical to consider how viscosity changes with increasing water content rather than water fugacity (Hirth and Kohlstedt, 2003):

\[ \eta_{II} = B^{\frac{1}{2}} A^{\frac{1}{2}} \left( \frac{1}{2} \sigma_{II}' \dot{\epsilon}_{II}' \right)^{\frac{1}{2}} C_{OH}^{-\frac{1}{2}} e^{(E/PV)} \]  

(1.12)

\( C_{OH} \) is the water concentration, which can be calculated from the water fugacity for olivine at lithospheric and mantle conditions (Kohlstedt et al., 1996):

\[ C_{OH} = A_{H_2O} e^{-(E_{H_2O} + PV_{H_2O}) f_{H_2O}} \]  

(1.13)

\( A_{H_2O} = 26 \times 10^{-6} \text{Si}^{-1} \text{MPa}^{-1} \), \( E_{H_2O} = 40 \text{ kJ mol}^{-1} \), \( V_{H_2O} = 10^{-5} \text{m}^3 \text{mol}^{-1} \), and \( f_{H_2O} \) is water fugacity in MPa (Zhao et al., 2004). Rocks can deform simultaneously by both dislocation and diffusion creep. This can be expressed in the form of a composite rheology (van den Berg et al., 1993):

\[ \frac{1}{\eta_{com}} = \frac{1}{\eta_{dis}} + \frac{1}{\eta_{diff}} \]  

(1.14)

\( \eta_{com} \), \( \eta_{dis} \), and \( \eta_{diff} \) are effective composite viscosity, effective dislocation creep viscosity, and effective diffusion creep viscosity, respectively. Figure 1.3 shows examples of olivine viscosity versus depth for a 60 Ma ocean, using flow law parameters from Karato and Wu (1993); Karato and Jung (2003); Hirth and Kohlstedt (2003); Kawazoe et al. (2009). This figure illustrates the range of mantle viscosities that can be present in numerical subduction models and shows that order of magni-
Figure 1.3: (a) Oceanic geotherm for a 60 Myr old ocean calculated from the plate cooling model (Turcotte and Schubert, 2002). (b) Olivine viscosity versus depth for a constant strain-rate of $10^{-15} \text{s}^{-1}$ and lithostatic pressure ($\rho=3250 \text{ kg m}^{-3}$). The grain size is 7 mm for diffusion creep and the water content is $200 \times 10^{-6} \text{Si}^{-1}$ for wet rheologies.

1.2.2.2 Brittle behaviour

When the deviatoric stress of a material reaches the yield stress ($\sigma_Y$), the material fails and experiences irreversible deformation (plastic flow). Brittle failure in nature leads to the formation of discrete faults or finite-width shear zones. In continuum numerical models, as used in this thesis, brittle deformation forms shear zones. The yield strength can be expressed by a Drucker-Prager yield criterion as:

$$\sigma_Y = P \sin(\phi_e) + C \cos(\phi_e)$$  \hspace{1cm} (1.15)

$P$ is dynamic pressure (mean stress), $\phi_e$ effective angle of internal friction, and $C$ cohesion (residual strength at $P = 0$). The effective angle of internal friction can take into account the effect of pore fluid pressure, which decreases the strength of
brittle rocks in wet conditions as (Beaumont et al., 1996):

\[
\sin(\phi_e) = (1 - \lambda) \sin(\phi)
\]  
(1.16)

\(\lambda\) is pore fluid pressure and is equal to the ratio between fluid pressure \((\rho_w g_y y)\) and lithostatic pressure \((\rho g y)\), where \(\rho_w\) is the density of the fluid, \(g\) gravitational acceleration, \(y\) depth, \(\rho\) the average density of the overlying rock, and \(\phi\) the angle of internal friction for dry rocks. Equation (1.15) shows that an increase in confining pressure will increase the strength of the material. After yielding, stresses can remain constant with increasing strain for an ideal plastic material, or either increase (strain hardening) or decrease (strain softening) with increasing strain for a non-ideal plastic material (Fig. 1.4a). A linear reduction of the effective angle of internal friction (Fig. 1.4b) and/or cohesion (Fig. 1.4c) can introduce strain weakening in numerical models (Beaumont et al., 1996) and help to localise deformation. Strain softening in nature can be linked to processes such as cataclastic deformation, formation of fault gouge, and the effect of elevated fluid pressures (Sibson, 1990; Sleep and Blanpied, 1992; Beaumont et al., 1996; Bos and Spiers, 2002).

1.2.2.3 Elastic behaviour

In contrast with irreversible viscous and brittle deformation, elastic materials deform instantaneously upon application of a load and recover the total acquired strain when the load is removed. Elastic strain is linearly proportional to elastic stress. For isotropic bodies, which have no preferred orientation for elastic behaviour, this relation can be expressed in terms of deviatoric stress \(\sigma'_{ij}\) and strain \(\epsilon'_{ij}\):

\[
\sigma'_{ij} = 2\mu \epsilon'_{ij}
\]  
(1.17)
\( \mu \) is the rigidity or shear modulus and has the dimension of stress. The order of magnitude of the rigidity of rocks is the same as their compressibility (Ranalli, 1995). The isotropic assumption is usually valid for the lithosphere and mantle of the Earth, but strongly foliated rocks can show anisotropic behavior (Turcotte and Schubert, 2002). Bending of oceanic lithosphere at subduction trenches and under the Hawaiian Islands, and of continental lithosphere at sedimentary basins can be described with elastic behaviour of the Earth’s lithosphere. Rocks show elastic behaviour at low temperatures (lithospheric temperature) and pressures (10-100 MPa) if the applied stress is not too large to pass the yield stress. The upper part of the lithosphere over short time scales (less than a few hundreds of thousand years) and the deeper part of the Earth over very short time scales (few seconds to few hours) can behave elastically. The importance of elasticity in large scale geodynamical processes is still debated, since the state of stress in the lithosphere is not fully understood yet, but it has been shown that elasticity can be ignored in most cases (Kaus and Becker, 2007). This can also be seen from the Maxwell relaxation time and Deborah number for processes at lithosphere to upper mantle scales. The Maxwell relaxation time, \( \tau_M \), can be expressed as (Peltier, 1985):

\[
\tau_M = \frac{\eta}{G}
\]

\( \eta \) is the effective viscosity and has values of \( 10^{17} \) to \( 10^{24} \) Pa and \( G \) is the elastic shear module, which is relatively well-constrained (\( 10^{10} \) to \( 10^{11} \) Pa) for the lithosphere (e.g. Dziewonski and Anderson, 1981). Hence, typical Maxwell relaxation times for lithospheric to upper mantle scales processes are less than 1 Myr. Elasticity may be important for processes that take place on timescales significantly shorter than \( \tau_M \).

The Deborah number, \( D_e \), can be expressed as (Reiner, 1964):

\[
D_e = \frac{\tau_M}{\tau_0}
\]

\( \tau_0 \) is the timescale of observation. Materials deform purely viscous for \( D_e = 0 \) and purely elastic for \( D_e = \infty \). For typical lithosphere to upper mantle scales models, \( (\tau_0 = 10\text{--}100 \text{ Myrs}) \), the Deborah number is significantly less than 1. Therefore, the effect of elasticity is not included in the simulations of this thesis.

1.2.3 Finite Element Method (FEM) formulation

We now have a system of equations given by the equations for conservation of momentum (1.7) under the condition of incompressibility (1.1), and the conservation of energy (1.10), for brittle (1.15) and viscous (1.12) rheologies. For most geodynamic
models, the set of equations become too complicated to be solved analytically and numerical techniques are required. In the field of subduction dynamics, finite element methods (FEM) and finite differences methods (FDM) are commonly used. Both methods divide the model domain into a finite number of nodes. Solutions to the equations are obtained at these nodes with interpolations between them for domain-covering solutions. In this thesis, a FEM is used.

In general, FEM are able to simulate large deformations, handle strong variations in material properties, and accurate follow material interfaces. FEM can also satisfy the free surface condition (Kaus et al., 2010; Quinquis et al., 2011), which leads to less computationally expensive solutions. Unstructured and arbitrary domain division is allowed in FEM, which provides a better coverage of irregular geometries. Despite the advantages, like all other methods FEM has its own limitations and drawbacks. FEM is can be complex to implement and can require large computational resources. Overall, FEM is one of the most widely used methods to simulate geological processes.

Finite elements are defined by their number of nodes, 3-node or 6-node triangular elements, 4-node or 8-node rectangular elements in two dimensional domains, and by the order of their interpolation functions. As solutions to the partial differential equations are obtained at the nodes only, the solution between nodes is approximated with piece-wise functions, known as shape functions:

$$\tilde{u} = \sum_{i=1}^{n} u_i N_i$$

$\tilde{u}$ is the approximated solution, $N_i$ shape functions, $u_i$ constant coefficients, and $n$ total number of nodes in an element. The accuracy of a FEM solution depends on the shape functions. Common choices are linear in velocity and constant pressure ($Q_1 P_0$) or quadratic in velocity and linear in pressure which is discontinuous between elements ($Q_2 P_{-1}$) (Zienkiewicz and Taylor, 2000). In this thesis, we use $Q_1 P_0$ elements as they are computationally fast. In general, increasing the number of elements even in complex domains with large variations in material properties can result in an approximation of the real solution to an acceptable precision.

In this thesis, we use SULEC (developed by Susanne Buit and Susan Ellis), which uses the Arbitrary Lagrangian-Eulerian (ALE) finite element method. It solves mechanical equations on an Eulerian grid which can be vertically stretched or shrunk to accommodate free surface displacements (Fullsack, 1995). The true free surface includes a stabilisation term to suppress numerical overshoot across interfaces with large density contrasts (Kaus et al., 2010; Quinquis et al., 2011). Material properties are stored on markers which are advected at the end of each time step. Thermal
equations are solved using a semi-implicit scheme.

1.3 Summary and outlook

In this thesis, we investigated styles of deformation in continent-continent collision following a phase of ocean-continent subduction using 2D thermo-mechanical models. We reviewed two continent-continent collision systems, the Norwegian Caledonides and the Zagros fold-and-thrust belt, which showed examples of parameters that may affect the style of deformation. We illustrated that a continental collision system can deform in different styles: stable continental subduction, lithospheric thickening, folding, Rayleigh-Taylor (RT) instabilities, and/or slab break-off. These styles are achieved through variations in driving velocity, crustal and lithospheric temperature, continental rheology, and the interaction with adjacent plates. The initial density contrast between the oceanic lithosphere and the asthenosphere has little effect on the style of deformation. Fast and cold systems ($T_{\text{lithosphere}} \approx 1200 \, ^\circ C$, $V > 4 \, \text{cm yr}^{-1}$) are more likely to experience lithospheric folding, while slow and warm systems ($T_{\text{lithosphere}} > 1350 \, ^\circ C$, $V < 2 \, \text{cm yr}^{-1}$) tend to experience RT-type of dripping. Stable subduction of continental crust and lithosphere occurs over a relatively large range of values for driving velocity (between 1 cm yr$^{-1}$ and 8 cm yr$^{-1}$) and base of lithospheric temperature (between 1200 °C and 1400 °C), and becomes the dominant style when the system has moderate velocity and temperature. We found that continents with a strong upper crust may experience subduction of the entire crust accompanied by lithospheric folding, whereas the upper crust accretes at the trench when the upper crust has a moderate to weak strength. Our results suggest that the entire continental crust can be scraped-off in the case of a weak lower crust. Weakening of the lithospheric mantle also promotes RT-type of dripping in a collision system. We found that a continent-continent collision system may experience slab break-off when the system has little interaction with the adjacent plates and slab pull of the subducted oceanic lithosphere is the main driving mechanism. Slab break-off reduces slab pull and may cause reverse motion of the subducting slab (eduction) and opening of the subduction channel, which allows exhumation of subducted buoyant materials. Eduction has been suggested to have facilitated exhumation in the Norwegian Caledonides (Andersen et al., 1991). We used a simple force balance of slab pull, slab push, slab bending, viscous resistance and buoyancy to explain the different collision styles caused by variations in velocity, temperature, rheology, density differences, and the interaction with adjacent plates.

We built on these results to investigate the role of salt in the Zagros fold-and-thrust
belt. Upper-mantle-scale models of oceanic subduction followed by continental collision, as far as possible constrained by available geological and geophysical data, gave the large-scale tectonic picture of the region of interest. We found that the presence of Hormuz salt plays a crucial role in decoupling overlying sediments from the basement, and localising deformation in the sediments by foreland-verging shear bands. Although this model predicts a topography and top basement dip that agrees reasonably well with the present-day observations, it is not able to reproduce the fold-dominated deformation in the simply folded Zagros. Since the results of the large-scale model show that deformation is localised at the upper crustal part of the incoming continent (Arabian plate), we then used the kinematic boundary conditions, thermal structure, and top basement dip of the upper-mantle-scale model as initial conditions of a series of upper-crustal-scale models. These were aimed at investigating the effects of basal and intervening weak layers, salt strength, basal dip, and lateral distribution of salt on the deformation style of the simply folded Zagros. We found that in addition to the Hormuz salt at the base of the sedimentary strata, at least one intervening weak layer is required to replicate the fold-dominated deformation in the SE Zagros. For our model set-up, a upper-crustal-scale model with an effective viscosity of $5 \times 10^{18} - 10^{19}$ Pa s for weak layers embedded in sediments with $\varphi = 5 ^{\circ}$ and $C = 20$ MPa, overlying a basement dipping with $+1 ^{\circ}$ toward the trench produces the best fit to folding at regular intervals and present-day topography in the simply folded Zagros. Models with a partial distribution of intervening weak layers can produce a deep thrust fault at the edge of distributed salt. The presence of such faults has formerly been attributed to reactivation of pre-existing normal faults in the basement.

The work described in this thesis contributes to the field of numerical modelling of continent-continent collision systems and aims to contribute to a better understanding of observed styles of deformation. Following on the results of this thesis, we would suggest several avenues for future studies:

1) **Melt generation.** This changes the temperature and viscosity of the system and therefore convection patterns and localisation of deformation.

2) **Water.** Slab dehydration together with mantle hydration changes the viscosity structure of the subduction wedge. In addition, released water can reduce the melting temperature and thus facilitate melting.

3) **Adaptive remeshing.** This allows to achieve high resolution in areas with more localised deformation while keeping low resolution in areas with low strain-rates. This reduces computational time by reducing the total number of elements in the simulation. With this method, one model can bridge between high-resolution and low-resolution regions, so that separate models, as in our fourth chapter, would no
longer be required.

4) Oblique collision. The onset of collision varies along strike in an oblique collision system, which could affect style, timing, and partitioning of deformation in the compressional setting.

References


2. A review of continent-continent collision
2.1 Introduction

Continent-continent collision occurs when an intervening ocean has been entirely subducted and two positively buoyant continents meet each other. The deformation of the continents in such a system depends on driving mechanisms and the mechanical and thermal properties of the oceanic as well as continental lithosphere. The continents could experience processes such as crustal and lithospheric thickening, formation of thrust nappes, magmatism in the overriding plate, metamorphism, exhumation, and post-collisional extension, which add complexity to the system (e.g. Dewey and Burke, 1973; Molnar and Tapponnier, 1975; Pfiffner et al., 1990; Andersen et al., 1991; Blanc et al., 2003; Leech et al., 2005). During the last decades, scientists from different disciplines have tried to address and better understand these complexities in, among others, the Himalayas, European Alps, Norwegian Caledonides, and Zagros (e.g. Stöcklin, 1974; Dewey and Burke, 1973; Jackson and Mckenzie, 1984; Smith, 1984; Andersen et al., 1990; Pfiffner et al., 1990; McKenzie and O’Nions, 1991; Torsvik, 1998; Bahroudi and Koyi, 2003; Alavi, 2004; Vernant et al., 2004; Torsvik and Cocks, 2005; Moutherau et al., 2006; Paul et al., 2006; Hacker et al., 2010; Krogh et al., 2011; Simmons et al., 2011; Nilforoushan et al., 2013).

In the following sections, we review the Norwegian Caledonides and the Zagros fold-and-thrust belt. These two continent-continent collision systems differ in terms of their age as well as their deformational complexities. The Norwegian Caledonides is an ancient collision system (430-400 Ma). It is deeply eroded, and has undergone post-orogenic extension, which allows observation of exhumed deep parts of the orogen. The presence of a ca. 5000 km$^2$ ultrahigh-pressure (UHP) terrane, which is surrounded by ca. 30,000 km$^2$ of high-pressure rocks, makes the Norwegian Caledonides an excellent place to study UHP rocks and their exhumation in continental collision systems. The Zagros is a relatively young collision system and is known as a large hydrocarbon province. In addition, it is one of the largest salt diapir provinces in the world, which makes this fold-and-thrust belt an intriguing natural laboratory to study the effects of salt on the style of deformation in continental collision systems. We briefly address the Norwegian Caledonides again in chapter 3, whereas the Zagros fold-and-thrust belt and the role of salt is the subject of chapter 4.
2.2 The Scandinavian Caledonides

The Scandinavian and East-Greenland Caledonides formed as a result of the continent-continent collision between Baltica and Laurentia following the closure of the Iapetus Ocean in the Middle Silurian (e.g. Roberts, 2003; Torsvik and Cocks, 2005). The closure is recorded by the termination of subduction-related magmatism at ca. 430 Ma (Corfu et al., 2006), as well as by the youngest preserved marine sediments of Middle Silurian (Wenlock) age below the obduction melange of the Solund-Stavford Ophiolite complex in the hinterland of the orogen (Furnes et al., 1990; Andersen et al., 1990). After the Iapetus Ocean had closed, ophiolites and island-arc complexes (commonly referred to as the Upper Allochthon), were obducted and emplaced onto Baltica. At the same time, subduction of the distal parts of the Caledonian margin of Baltica continued (see Fig. 2.1, from Hacker et al., 2003). The initial subduction of continental crust is recorded by the oldest eclogites (430-425 Ma) related to the continental collision. These originated in Baltica and are presently found in the Lindás nappe of the Bergen arcs in western Norway (e.g. Austrheim, 1987; Glodny et al., 2008). After the initial continental collision between Laurentia and Baltica, the orogeny continued to evolve for more than 30 Myr, into the Lower Devonian. This is recorded both by the change in the sedimentary facies from shallow-marine to continental molasse-type sediments in the foreland in the uppermost Silurian as a response to crustal thickening and uplift (Hacker and Gans, 2005) as well as by widespread eclogite facies metamorphism in the Western Gneiss Region (WGR). The eclogite facies metamorphism in the WGR commenced at approximately 420 Ma and lasted until ca. 397 Ma (e.g. Kylander-Clark et al., 2009; Krog et al., 2011). The 30 million year duration of the collision thickened the originally wide hyperextended Caledonian continental margin of Baltica (Andersen et al., 2012).

The distal Caledonian margin of Baltica had experienced a complex and still
poorly understood pre-collision history. This included accretion of terranes of several generations, which now are identified as pre-collision (older than 430 Ma) deformed and metamorphic rocks occurring in a number of localities along the length of the orogen, from Jæren in the south to Finnmark in the north (see Fig. 2.2, and Birkeland (1981); Smith et al. (2008)). Some of these were entirely outboard, whereas rocks in the distal margin of Baltica were also affected by deformation and metamorphism in the lower and middle Ordovician (Corfu et al., 2014). Several papers have, however, identified important events at ca. 470-480 Ma and at ca. 440-450 Ma (e.g. Andersen et al., 1998; Sturt and Ramsay, 1999; Roberts, 2003; Brueckner and van Roermund, 2007; Root and Corfu, 2011), but a consensus of how to interpret this complex history has not yet been established, and will not be discussed further here.

The 30 million years long duration of the continental collision eventually resulted in a very large and thick nappe-stack that can be traced continuously along strike for nearly 2000 km, from the North Sea in the South to the Barents Sea in the North (Fig. 2.3). This is of similar size as the Himalayan collision (Jackson et al., 2004; Labrousse et al., 2010).
2.2.1 Tectonostratigraphy

The Caledonian collision produced a nappe-stack in the Scandinavian Caledonides, which consist of four major allochthon complexes, named the Lower, Middle, Upper and Uppermost Allochthons, respectively (for a full review, see Roberts and Gee, 1985) (Fig. 2.3). The classical interpretation of this tectonostratigraphy is that the Lower and Middle Allochthons as well as the Seven nappes of the Upper Allochthon consist of rocks that originally formed part of the outermost (distal) Caledonian margin of Baltica, and hence are interpreted as ‘inboard’ with respect to Baltica (Stephens and Gee, 1985). The Upper Allochthon consists of rock units which are ‘outboard’ terranes formed in the Iapetus Ocean and Ægir Sea (Torsvik and Cocks, 2005). These rocks are mostly ophiolites, island-arc and back-arc complexes of lower Ordovician to Middle Silurian (495-430 Ma) age (Roberts, 2003). The Uppermost Allochthon consists of rocks, which are believed to have originated in Laurentia. The Uppermost Allochthon is dominated by magmatic rocks including the batholithic intrusions of the Helgeland and Hitra-Smøla regions (Roberts et al., 2006). It also contains high-grade metamorphic rocks, including the ca. 470 Ma old eclogites of the Tromsø Nappe (Corfu et al., 2006, 2014).

The rocks within the WGR constitute one of the best preserved and exhumed HP to UHP provinces of the world. It therefore represents an excellent natural laboratories
to study deep crustal processes during collision and burial as well as exhumation of such complexes (e.g. Andersen et al., 1991; Austrheim and Boundy, 1994; Krabben-dam and Dewey, 1998; John et al., 2009; Hacker et al., 2010; Warren, 2013).

2.2.2 Plate Tectonics and Palaeogeography

At about 1 Ga, Laurentia, Baltica, and several intervening continents had amalgamated to form a supercontinent known as Rodinia (Torsvik, 2003). Laurentia constituted the core of this super-continent (Fig. 2.4). Disruption of Rodinia commenced at ca. 750 Ma and was followed by an extensive phase of volcanism at about 630 Ma. A new supercontinent, Gondwana, formed at ca. 600-550 Ma (Fig. 2.5). Gondwana stretched from the South Pole (South America) to the Equator (e.g. Australia), and was a result of Pan-African orogenic events, which mark one of the most spectacular mountain-belt building episodes in Earth history (Torsvik and Cocks, 2005). In the lower Cambrian, Baltica became a continental entity, separated from Gondwana by the Tornquist Sea, from Siberia by the Ægir Sea, and from Laurentia by the Iapetus Ocean (Fig. 2.5, from Torsvik and Cocks, 2005). Baltica was proposed to be geographically inverted at this time (Hartz and Torsvik, 2002). The plate-tectonic motion of Baltica during the Ordovician was characterized by counter-clockwise rotation of up to 3° Myr\(^{-1}\) and by latitude velocities of up to 8 cm yr\(^{-1}\). Palaeomagnetic studies reveal that Baltica was positioned between 30° and 60° South in the
upper Ordovician (Fig. 2.6 from Torsvik and Cocks, 2005). A distinct fauna characteristic of Baltica also developed at this stage (Torsvik and Cocks, 2005). The Caledonian margin of Baltica faced Laurentia for the first time in the upper Ordovician. Rifting of Avalonia from Gondwana occurred contemporaneous with subduction of the Tornquist Sea crust beneath Avalonia. These two events led to rapid northward movement of Avalonia and eventually a "soft-docking" with Baltica along the Tornquist suture (Torsvik and Rehnström, 2003; Torsvik and Cocks, 2005). The closure of the Tornquist Sea is marked by a unified Baltic-Avalonian fauna, by the termination of subduction related volcanism in the British Isles, and by low-grade metamorphism identified from basement cores in the North Sea (Torsvik, 1998). Closure of the Iapetus Ocean happened quickly by rapid northward drift of Baltica-Avalonia at a maximum latitude velocity of 16 cm yr$^{-1}$, comparable only to that
of India immediately prior to the collision with Asia. By 430 Ma, Baltica-Avalonia collided with Laurentia and formed Laurussia. The Caledonian orogeny was a result of the rapid plate-collision and deep subduction of Baltica underneath Laurentia (Andersen et al., 1991). During and after the collision, the Caledonides went into a stage of late- to post-orogenic extension, which eventually resulted in exhumation of the high and ultra-high pressure rocks (e.g. Andersen et al., 1991; Krabbendam and Dewey, 1998; Fossen, 2010).

In the upper Carboniferous, Gondwana merged with Laurussia and all intervening terranes to form the supercontinent Pangea (Van der Voo and Torsvik, 2001; Torsvik and Cocks, 2005; Murphy and Nance, 2008; Domeier et al., 2012).

### 2.2.3 Collision, deep burial and metamorphism of the Western Gneiss Region

As described above, continental collision of Baltica and Laurentia formed the Caledonides presently exposed in Scandinavia and East-Greenland (Fig. 2.7). High- and Ultrahigh pressure metamorphic rocks related to this collision are exposed in both Scandinavia and East Greenland. The discussion here will primarily concentrate on the best studied UHP rocks of the Western Gneiss Region (WGR) of southwest Norway.

The WGR consists of the Western Gneiss Complex basement (WGC) overlain by highly attenuated rock units interpreted to represent Caledonian allochthons (e.g. Lutro and Tveten, 1998; Krogh et al., 2011) (Fig. 2.3). The WGC is dominated by Middle Proterozoic orthogneiss (ca. 1700 to 950 Ma) of intermediate to grani-
toid composition, but there are also numerous anorthositic to mafic and ultramafic enclaves. Both the rocks of the WGC and the overlying allochthons experienced (U)HP metamorphism, indicating that the WGR was deeply buried and partly subducted to mantle depths during the collision (Hacker et al., 2010). The domains containing UHP eclogites, as well as majorite and diamond-bearing peridotites that record pressures up to ca. 6 GPa (Lappin and Smith, 1978), are found in the northwestern parts of the WGR (see Fig. 2.7, from Hacker et al., 2010). Pressures generally diminish southeastward from the UHP domains (Hwang et al., 2013). The first eclogites crystallised at temperatures and pressures of ca. 600°C and 1.8-2.0 GPa (Hacker et al., 2010). Southeast of the eclogite-in isograd (Fig. 2.7), Caledonian regional metamorphism attained amphibolite-facies conditions, but the main structure of the gneisses in this area is pre-Caledonian expect near the contact with the overriding nappes (Fauconnier et al., 2014). A "lid" of Proterozoic crust is preserved at the highest structural levels of the WGR, below the Caledonian floor thrust. This is also the case further to the north in the Caledonides, where large basement complexes in the Lofoten archipelago and the Western Troms Basement complex are characterised by mostly intact Proterozoic to Archean basement interpreted to represent the continuation of the Baltic Shield below the Caledonian nappes (Bergh et al., 2010). The most common metamorphic assemblages in the WGR, as indicated already by Bryhni (1966), is amphibolite-facies formed at temperatures of 600-700°C and pressures of 1.2-0.5 GPa. These conditions are characteristic of Barrovian to Buchan metamorphism (Hacker et al., 2010). Where eclogites are present, it is obvious that the amphibolitization post-dates the eclogites. A general north-westward increase in metamorphic temperature is indicated by calculated temperatures from Pressure-Temperature-estimates, the progressive Caledonian resetting of U/Pb ages of sphene, the distribution of sillimanite, and evidence of partial melting (Tucker and Krogh, 1988; Labrousse et al., 2011; Hacker et al., 2010) (Fig. 2.7). This progressive regional metamorphism is apparently valid for both the (U)HP and later events. Of particular importance is the widespread occurrence of partial melts which are abundantly common in the northwest (Labrousse et al., 2011; Ganzhom et al., 2014).

There is, however, a major problem with a straightforward interpretation of metamorphism in the WGR: in many areas several metamorphic assemblages may coexist within small areas, at outcrop-scale or even at the scale of individual samples (e.g. Straume and Austheim, 1999; Krabbendam et al., 2000; Vrijmoed et al., 2009). This demonstrates that metamorphic equilibration during any given event was incomplete, particularly in domains where the rocks are little deformed. In most areas this variation in metamorphic grade may be assigned to preservation of
different metamorphic events of widely different ages (i.e. Precambrian vs. Caledonian) such as at Flatraket and Kråkeneset (Krabbenberg et al., 2000; John et al., 2009), and within mantle peridotites (Scambeluri et al. 2009). In parts, a Proterozoic amphibolite- to granulite-facies metamorphism is preserved which is associated with extensive plutonism (Cotkin, 1997; Root et al., 2005). In the southern part of the WGR this Proterozoic metamorphism (950 Ma, e.g. Rohr et al., 2004) peaked at ca. 900°C and 1.0 GPa. In other areas, however, different pressure assemblages of the same age appear to be preserved in the same outcrop (Vrijmoed et al., 2009). Such observations are highly enigmatic and are not explained by the commonly accepted models for metamorphic petrology where pressure is always correlated with burial depth. These observations are also of major importance to models that attempt to quantify burial- and exhumation depths in the WGR (and elsewhere).

There are several challenging questions concerning the formation of the WGR (U)HP rocks and their exhumation: a) What is the maximum metamorphic pressure that the UHP rocks experienced? b) What would be the deepest depth that the rocks reached? c) Do we have an estimate of the duration of metamorphism? d) How were the UHP rocks exhumed? These are discussed in turn below.

2.2.3.1 Maximum metamorphism pressure of UHP rocks in the WGR

Griffin et al. (1978) originally suggested that approximately 2 GPa and 750-800°C was the maximum metamorphic conditions recorded by the basement gneisses in the WGR. This declaration was challenged by Lappin and Smith (1978), who took low Al-orthopyroxene coexisting with garnet (down to 0.15 or 0.20 wt% Al₂O₃) within opx-eclogite or garnet websterite as the first mineral-chemical evidence to suggest that the eclogite-facies had experienced ultra-high pressure conditions. The confirmation of the presence of regional metamorphic coesite in the Selje district by Smith (1984), clearly demonstrated that the maximum pressures experienced by the continental crust rocks in the NW part of the WGR was much higher than previously anticipated. This is now generally accepted and coesite, or evidence for the former presence of this SiO₂ polymorph, is relatively common in the WGR. Smith’s discovery had profound impact, because it indicated that a minimum pressure of 2.7 to 2.8 GPa had affected large areas in the WGR. Some years later, two other major discoveries indicated that even higher metamorphic pressures had affected these rocks: microdiamonds were found on the island of Fjørtoft and later elsewhere (Dobrzheinetskaya et al., 1995; Vrijmoed et al., 2008) and evolved majoritic garnet was discovered on the island of Otrøy (Van Roermund et al, 2001). At present mantle
peridotites included in the basement units of Otrøy and Nordøyane, paragneiss at Fjørtoft, and a peridotite at Svartberget near Bud demonstrate that locally pressures of up to 6 GPa have affected these rocks (Van Roermund, 2009; Vrijmoed et al., 2008, 2009). The most extreme pressures recorded by rocks in the WGR are therefore now more than twice the pressures required for coesite-grade eclogite facies.

### 2.2.3.2 Maximum subduction depth of UHP rocks in the WGR

The pressures recorded in the WGR, of up to 6 GPa, would indicate that these rocks reached depths of almost 200 km, if a lithostatic pressure field is assumed. Also, exolved majoritic garnet microstructures and diamond have been found in several garnet peridotites within the northernmost UHP domain of the WGR (see review in Van Roermund, 2009). Because majoritic garnet is thought to be common at depths of more than 180 km (Haggerty and Sautter, 2010), the former existence of majoritic garnet has been taken to indicate that the northernmost UHP province in the WGR has been buried and exhumed to such depths or deeper (Van Roermund, 2009). At present, we have no mechanism that can explain exhumation of the UHP rocks from these depths to the conditions of 2.3-3 GPa (max 100 km depth). However, data from the diamond bearing garnet veins in the Svartberget peridotite challenge the commonly accepted models of burial (see details in Vrijmoed et al., 2009). Vrijmoed et al. (2009) suggested that local melt-included tectonic overpressure (by 1-2 GPa) could have contributed to the high pressures. Such alternative models should clearly be explored and quantified.

### 2.2.3.3 UHP metamorphism duration in the WGR

The age of the (U)HP metamorphism in the WGR has been a topic of discussion for many years, because the protoliths to the (U)HP rocks are Proterozoic and a whole range of radiometric whole-rock- and mineral ages have therefore been recorded. The first indication of a Caledonian origin came from early U/Pb and Sm/Nd studies by Gebauer and Grünfelder (1976) and Griffin and Brueckner (1985), who obtained relatively young Caledonian ages, although with considerable uncertainty. We now have a large number of dated eclogite and peridotite localities and the main result of these is that (U)HP metamorphism in the WGR commenced at approximately 420 Ma and lasted until 400 Ma (see recent summary by Krogh et al., 2011). The interpretation of the age of (U)HP metamorphism is complicated, since commonly dated zircon crystals providing high-precision ages are not necessarily part
of the eclogite facies mineralogy, and it is not known if these ages date the peak-pressure. Therefore, other methods dating the eclogite-facies mineral assemblages with higher precision than could be achieved in the early studies have been used (Vrijmoed et al., 2008; Kylander-Clark et al., 2009). These methods (with their uncertainties) may potentially be used to date the duration of the (U)HP metamorphism. Kylander-Clark et al. (2009) used the age difference between Lu-Hf and Sm-Nd in the WGR and reported nearly 20 Ma for the duration of garnet growth, from 415 Ma to 391 Ma. The combination of this result and previous works led to the suggestion that the UHP metamorphism of the WGR occurred over a 20 Ma period (Kylander-Clark et al., 2009). This agrees relatively well with the distribution of U/Pb ages of zircons (Krogh et al., 2011). Calculated P-T paths for the WGR indicate that the metamorphosed rocks remained at more than 750°C from the time of UHP metamorphism to the time they had reached mid-crustal depths of 15-20 km (Terry et al, 2000; Root et al., 2005). If the crustal materials spend 15-20 Ma at UHP conditions, what could be their size? By assuming a thermal diffusivity of ca. 10^{-6} \text{ m}^2\text{s}^{-1}, the characteristic thermal diffusion distance would be 20-25 km and slabs of only a few kilometres thick could not have survived without melting (Root et al., 2005). Kylander-Clark et al. (2009) used numerical thermal models to show that large UHP terranes, that experienced a long period of peak or near-peak metamorphism, can be modelled by slow subduction (2-4 mm yr^{-1}) of a relatively warm and thick slab (tens of kilometres), while smaller, thinner UHP units, that reached high temperature and retained them through the bulk of exhumation, experienced rapid exhumation.

2.2.3.4 Formation of UHP rocks and their exhumation in the WGR

The Laurentia-Baltica collision at 430-400 Ma resulted in emplacement of the allochthons onto Baltica and eventually subduction of Baltica underneath Laurentia (e.g. Andersen and Jamtveit, 1990; Jolivet et al., 2005). During this period, the WGR and part of the overlying allochthons started to subduct and the leading edge of Baltica reached HP conditions depths already at 430 Ma (Gladny et al., 2008). The steep lineations in the Volda, Åndalsnes and Romsdalen domains imply vertical thickening of crust during the subduction phase (Hacker et al., 2010) (Fig. 2.8a). Convergence between Baltica and Laurentia continued until ca. 400 Ma and caused subduction of the WGR and allochthons to UHP conditions, contemporaneous with thrusting in the foreland (Fossen and Dunlap, 1998). Eclogites with a mapped field gradient from 1.8 to 2.8 GPa formed across the WGR (Fig. 2.7). Contemporaneously, the HP eclogites of the Lindås nappe were exhumed in the hinterland, e.g.
Figure 2.8: Schematic, composite profile illustrating formation of UHP rocks and their exhumation in the WGR. From Hacker et al. (2010).
in the Bergen area (Jolivet et al., 2005; Glodny et al., 2008; Hacker et al., 2010) (Fig. 2.8b). After 400 Ma, extension and transtension became the dominant large-scale tectonic process in Baltica and Laurentia (e.g. Krabbendam and Dewey, 1998; Fossen, 2010). The westward decrease in U/Pb sphene and $^{40}\text{Ar}/^{39}\text{Ar}$ muscovite ages indicates that the WGR was progressively exhumed to higher crustal levels from the southeast to the northwest, as the original deeper levels were exhumed and cooled at progressively younger ages in the northwest (Spencer et al., 2013). Amphibolite-facies structures throughout the WGR indicate that this period was accompanied by coaxial E-W extension, vertical thinning, and minor N-S shortening (Hacker et al., 2010) (Fig. 2.8c). By 395 Ma, the entire southern half and eastern third of the WGR had been exhumed from pressures of ca. 2 GPa to upper crustal depths and had cooled below $400^\circ$C, whereas high-temperature deformation and metamorphism continued in the northwestern part. Decompression from 3 GPa to 0.5 GPa in the western part of the WGR caused conversion of the gneisses to lower pressure assemblages characterised by formation of symplectites and widespread decompression melting (Hacker et al., 2010; Labrousse et al., 2011) (Fig. 2.8d).

2.3 The Zagros fold-and-thrust belt

The Zagros fold-and-thrust belt formed as a result of continent-continent collision between Arabia and Central Iran following the closure of the Neo-Tethys ocean in the upper Cretaceous. The fold-and-thrust belt extends over a vast area with a width of 200-300 km (NE-SW) and a length of about 1800 km (NW-SE). The Zagros are surrounded by the East Anatolian fault in Turkey in the north-west (Scott, 1981; Jackson and Mckenzie, 1984), the central Iranian micro-plate in the north and north-east (Scott, 1981), the Oman fault, which separates the Zagros from the Makran accretionary prism to the east and south-east (Falcon, 1967; Haynes and Mcquillan, 1974; Jackson and Mckenzie, 1984), and the Persian gulf foreland in the south and south-west (Jackson and Mckenzie, 1984) (Fig. 2.9). The Main Zagros Thrust Fault (MZTF) is considered to be the suture of the collision between Arabia and Central Iran (Dewey et al., 1973; Sengör, 1984; Berberian, 1995). This NW-SE trending fault was the only active fault in the Zagros until the middle-Miocene (Falcon, 1967; Berberian, 1995). Recent studies suggested that the MZTF may extend to Moho depths (Agard et al., 2005; Paul et al., 2006). The exact timing for the onset of continental collision in the Zagros is still debated (Dewey et al., 1973; Falcon, 1974; Scott, 1981; Dercourt et al., 1986; McQuarrie et al., 2003; Alavi, 2004; Agard et al., 2005). This is partly because oblique convergence between the Arabian plate and Eurasia resulted in complex deformation within the suture zone (Mohajjel and
Fergusson (2000; Sarkarinejad and Azizi, 2008; Sarkarinejad et al., 2008; Hatzfeld and Molnar, 2010). Agard et al. (2005) infer that collision started at about 25-23 Ma from dating blueschist facies rocks found south of the Sanandaj-Sirjan zone (Fig. 2.9). In contrast, dating of the end of arc-related magmatism in Central Iran supports initiation of collision at 34 Ma (Ballato et al., 2011; Mouthereau et al., 2012). Since the onset of collision, ca. 500-800 km of convergence has occurred and GPS measurements show that convergence is still ongoing today (Vernant et al., 2004; Hessami et al., 2006; Hatzfeld and Molnar, 2010; Hatzfeld et al., 2010). Deformation in the Zagros, which mainly occurred in the form of thrust faulting and folding, started a few million years after the onset of collision at the MZTF and then migrated southwestwards through time (Berberian, 1995; Hatzfeld et al., 2010). The style of deformation differs significantly from the NW Zagros to the SE Zagros, which are separated by the north-south trending Kazerun fault. This is apparent in dramatic changes in the sedimentary stratification (Bahroudi and Koyi, 2003; Alavi, 2004; Hatzfeld et al., 2010). In the SE Zagros, the sedimentary column contains several weak layers consisting of halite, gypsum, anhydrite, dolomite, and shale (Alavi, 2004). The presence of these weak layers is believed to be the main cause of fold-dominated deformation in the SE Zagros (Colman-Sadd, 1978; Davies and Engelder, 1985; Talbot and Alavi, 1996; Bahroudi and Koyi, 2003; Alavi, 2004;
Yamato et al., 2011). The deformation is mainly thrust faulting-dominated in the NW Zagros, where these weak intervening layers are absent.

2.3.1 Plate Tectonics and Paleogeography

From upper Precambrian to upper Paleozoic, parts of present-day central Iran, Pakistan, central Afghanistan, Arabia, and Turkey formed a passive margin along the north rim of Gondwana (Stöcklin, 1968; Berberian and King, 1981; Scotese and Mckerrow, 1990; Torsvik and Cocks, 2013). In the Permian, the Cimmerian terranes rifted away from this margin, opening the Neo-Tethys Ocean in their wake (Fig. 2.10) (Sengör, 1988; Stampfli, 2000; Muttoni et al., 2009). Following this, the north-eastern margin of the Arabian plate remained a passive margin during the Mesozoic (Berberian and King, 1981; Stoneley, 1981; Berberian, 1995). In the upper Cretaceous, Arabia and Africa converged toward Eurasia (Dercourt et al., 1986). This convergence resulted in subduction of the Neo-Tethys ocean and parts of the Arabian passive margin beneath the central Iranian micro-plate (Dewey et al., 1973; Berberian and King, 1981; Dercourt et al., 1986). Subduction was accompanied by two tectonic events: a) widespread Mesozoic and syn- to post-Eocene calc-alkaline magmatism occurred in the Central Iranian micro-plate along the Sanandaj-Sirjan zone (SSZ) and the Urumieh-Dokhtar magmatic arc (UDMA) (Berberian and King,
1981; Agard et al., 2006; Moutheau et al., 2012), and b) syn-convergence ex-
humation of blueschist facies rocks from 35-50 km to depths lower than 15-20 km 
ocurred before 80 Ma (Agard et al., 2006; Omrani et al., 2008; Sarkarinejad et 
al., 2008; Agard et al., 2011). These rocks now form kilometre-scale bodies in the 
Central Iranian micro-plate and southeast of the SSZ (Sabzehei et al., 1994). These 
blueschist facies rocks experienced maximum P-T conditions of ca. 1.0-1.1 GPa 
and 520-530°C at about 95-85 Ma, followed by a pressure decrease to 0.5-0.6 GPa 
in less than 10 Myr (Agard et al., 2006; Omrani et al., 2008; Sarkarinejad et al., 
2008; Agard et al., 2011). Flat-slab subduction followed by slab roll-back has been 
suggested as a mechanism to explain the exhumation of HP rocks in the Zagros (e.g. 
Rosenbaum et al., 2002; Agard et al., 2006; Moghadam et al., 2010; Agard et al., 
2011). The NE convergence between Arabia and Central Iran continued until the 
lower Miocene when the direction of convergence changed to N-S due to the onset 
of continental rifting in the Red Sea and opening of the Gulf of Aden (McQuarrie et 
al., 2003; Moutheau et al., 2012).
Convergence between Arabia and Central Iran slowed down after the Neo-Tethys 
Ocean was entirely subducted and the distal Arabian margin under-thrusted the 
Central Iranian micro-plate at ca. 25 Ma (McQuarrie et al., 2003; Moutheau et al., 
2012). East of the collision system, subduction of Oman oceanic lithosphere 
underneath Eurasian resulted in accretion of submarine sediments on top of the 
oceanic crust and growth of the Makran accretionary prism seawards. The present-
day Arabian-Eurasian convergence is partly consumed by continent-continent colli-
sion in the Zagros and partly by the Makran subduction system. These two systems 
are separated from each other by the Oman fault line. Tomography data suggest 
steepening or break-off of the subducted slab under the NW Zagros (Fig. 2.12), 
which is likely concurrent with slab break-off in Anatolia at about 12 Ma (Regard 
et al., 2005; Authemayou et al., 2006; Faccenna et al., 2006; Hafkenscheid et al., 
2006; Agard et al., 2011; Verges et al., 2011). However, more recent tomographic 
inversions imply a continuous underthrusting of the Arabian plate with a low dip 
gle beneath the Central Iranian block in the SE Zagros (Chang et al., 2010; Sim-
show that slab break-off is likely to have initiated nearly 16 Myr after the onset of 
the Arabia-Central Iran collision under the NW Zagros and then propagated par-
allel to the trench southeastwards with an average rate of 3 cm yr$^{-1}$ (Fig. 2.11). 
This could explain the observed slab break-off under the NW Zagros and ongoing 
derthrusting in the SE Zagros.
2.3.2 Folding- and thrust-dominated deformation

The convergence rate between Arabia and Eurasia has probably not changed much since ca. 22 Ma (ArRajehi et al., 2010), after the decrease from 3 cm yr\(^{-1}\) to 2 cm yr\(^{-1}\), caused by the onset of crustal thickening in the Zagros (Mouthereau, 2011). This implies 440 km of convergence, which has been suggested to be distributed as 135 km of shortening in the Zagros, 180 km of shortening throughout Central Iran, 50 km across the Alborz, and 75 km taken up by subduction of the Caspian Sea (Agard et al., 2005; Mouthereau et al., 2007; Mouthereau, 2011). Shortening in the Zagros is inferred to have started at the MZTF and propagated South-Westwards away from the suture zone (Hessami et al., 2006; Hatzfeld et al., 2010). Using balanced cross-sections, Mouthereau et al. (2007) proposed that only 5% of the shortening is consumed in the Zagros fold-and-thrust belt and that the rest is accommodated by underthrusting across the MZTF and underplating of Arabian crust below the Sanandaj-Sirjan Zone. The evolution of sedimentary successions in the Zagros, which has been investigated by balanced and restored cross-sections, isopach maps, and tectonic subsidence curves, indicates that the basal and intermediate decollements play a major role in controlling and partitioning deformation in the simply
Figure 2.12: P-wave tomographic transects indicating slab break-off under the NW Zagros. High seismic velocities (blue colour) are interpreted as cold areas (slab) and slow velocities (red colour) represent warm areas. From Verges et al. (2011).
folded Zagros (e.g. Blanc et al., 2003; Sepehr and Cosgrove, 2004; Sherkati and Letouzey, 2004). The deformation is accommodated in thrusting of the sedimentary sequences in the NW Zagros and folding of the sedimentary cover in the SE Zagros (Stöcklin, 1974; Stoneley, 1981; Davies and Engelder, 1985; Bahroudi and Koyi, 2003; Moutheureau et al., 2012).

The sedimentary cover of the Zagros consist of a 12-14 km thick sediment package of lower Cambrian to Pliocene strata, which overlie the crystalline basement (Falcon, 1969; Stöcklin, 1968; Colman-Sadd, 1978; Alavi, 2004). The thickness and stratification of this sedimentary cover exhibits significant spatial variations (Alavi, 2004). The Hormuz salt is believed to be the oldest and the most important sedimentary unit in the Zagros. It was deposited on basement from the upper Proterozoic to lower Cambrian (Falcon, 1969; Colman-Sadd, 1978). The rest of the sedimentary cover consist of 6-10 km thick typical continental passive margin sedimentary sequences (with limestone, marl, sandstone, conglomerate, gypsum, anhydrite, dolomite, and shale) deposited either on the Hormuz salt unit or directly on the basement rocks (Alavi, 2004; McQuarrie, 2004; Ramsey et al., 2008). The spatial distribution of the Hormuz salt can be approximated from the spatial distribution of extruded salt diapirs. Regions with no extruded salt are assumed to indicate that the sedimentary cover lies directly on the basement, whereas regions with extruded salt diapirs would indicate that salt layers are present between the basement and the overlying sedimentary layers.

Bahroudi and Koyi (2003) suggested that the Zagros fold-and-thrust belt can be divided into five different domains based on the lithological facies sequences and the structural styles (Fig. 2.13): a) Lorestan is the north-western domain with many NE-dipping thrust faults and fault-related-folds. The absence of extruded or buried salt structures indicates that the Hormuz salt layer is not present in this domain. b) Izeh in the north is characterised by the exposure of salt structures along the fault zones, accompanied by southward-verging asymmetric, long fault-related-folds. c) Kazerun is separated from the adjacent Izeh domain by the right-lateral Kazerun fault and from the Fars province to the southwest by the left-lateral Nezamabad fault. Characteristic of this domain is the existence of many extruded salt structures and further propagation of the deformation front into the foreland in comparison with the other domains. d) Fars is characterised by the absence of salt structures, which implies that either the Hormuz salt was never deposited or is very thin on the Fars platform. e) Laristan is the south-eastern domain and is separated from the Fars platform by the right-lateral Razak fault and from the Makran by the Oman fault. Salt structures are abundant in this domain.
2.3.3 Examples of modelling of the Zagros fold-and-thrust belt

Analogue modelling by Bahroudi and Koyi (2003), using a simplified distribution of Hormuz salt (Fig. 2.14), suggests that domains with a viscous decollement experience folding and faulting with gentle and long tapers, whereas domains with a frictional decollement experience more intensive deformation, accommodated in thrust faults, and form shorter and steeper tapers. Because both domains in the experiments experienced the same amount of shortening, the domain with the frictional decollement accommodated larger displacements along the faults (Fig. 2.14). Later analogue modelling by Nilforoushan and Koyi (2007) of shortening above a frictional decollement adjacent to a viscous decollement, confirms these results.

The numerical modelling results of Yamato et al. (2011) show that a single viscous decollement at the base of the sedimentary succession (representing the Hormuz salt) is not enough to obtain fold-dominated deformation, because it results in thrust-dominated deformation (Fig. 2.15). In addition to a weak basal layer, intervening weak layers are required to reproduce the regular 14±3 km wavelength folding that is observed in the SE Zagros (Yamato et al., 2011). Their results suggest that a viscosity of $10^{18}$ Pa s for the weak layers, and an average low friction angle of 5° for the sediments gives the best fit to the observations (Yamato et al., 2011).

Thermal-mechanical modelling by Nilforoushan et al. (2013) indicates that, in addition to the salt distribution, the geothermal gradient is a key factor controlling the crustal-scale deformation of the Zagros. They find that a cold rheology model (Moho temperature 400°C) simulates the present-day structure of the Zagros best. Their results also predict that most of the shortening in the Zagros has occurred in the sedimentary cover rather than in the basement, which supports the integrated
2.4 A brief discussion of deformation in continental collision

Continent-continent collision systems can be found in various places around the world, e.g., the Himalayas, European Alps, Norwegian Caledonides, and Zagros.
Figure 2.15: a) Topography of the Fars province and distribution of fold wavelengths indicating that the dominant folding wavelength is 14.4 ± 3 km. b) Influence of multiple weak layers and elasticity on folding. VEP = visco-elastic-plastic, $\lambda$ = dominant wavelength, $H$ = thickness of sedimentary cover, $q$ = instability growth, $\dot{\varepsilon}$ = strain rate. Figure from Yamato et al. (2011).
Although the systems obviously differ in some aspects, such as convergence velocity, thermal structure, and lithologies and thickness of the continents, they share many characteristics as well. Hence, investigation of collision system may help to understand others. For example, investigation of a young collision system might aid understanding of the deformational history in an old system. In this chapter, we reviewed the Norwegian Caledonides and the Zagros fold-and-thrust belt as two examples of continent-continent collision systems. Here, we briefly point out similarities and differences of these systems.

From dating of the end of subduction-related magmatism and preserved marine sediments, the Norwegian Caledonides are thought to have entered the collision phase at ca. 430 Ma (Furnes et al., 1990; Andersen et al., 1990; Corfu et al., 2006). In contrast, the Zagros is a relatively young collision system with the onset of collision inferred at ca. 35-25 Ma, based on dating of the end of arc-related magmatism and blueschist facies rocks (Agard et al., 2005; Ballato et al., 2011).

After the initiation of continental collision, convergence only lasted 30 million years in the Norwegian Caledonides and was mostly accommodated with crustal and lithospheric thickening and subduction of the continental edge of Baltica (Hacker and Gans, 2005). In the Zagros on the other hand, convergence is still ongoing (Vernant et al., 2004; Hatzfeld et al., 2010). Here convergence is mainly accommodated by crustal and lithospheric thickening and subduction of the continental margin of the Arabian plate (Dahlen et al., 1984; Dehghani and Makris, 1984; Davies and Engelder, 1985; Paul et al., 2006; Priestley et al., 2012). Crustal thickening occurred in the form of nappe-stacking in the Norwegian Caledonides, while it takes place in the form of thrust faulting- and fold-dominated deformation in the Zagros (Andersen et al., 1991; Colman-Sadd, 1978). The presence of several weak layers in the sedimentary succession is thought to play an important role in controlling the style of deformation in the Zagros (Colman-Sadd, 1978; Davies and Engelder, 1985; Bahroudi and Koyi, 2003; Yamato et al., 2011).

Convergence ceased in the Norwegian Caledonides either because of delamination at the mantle lithosphere from the crust or slab break-off at ca. 400 Ma (Andersen et al., 1991; Duretz et al., 2011). This was followed by 180 km of post-collisional extension, which resulted in rapid uplift (1-2 mm yr\(^{-1}\)) and exhumation of high and ultra-high pressure rocks during 20-30 million years (Andersen and Jamtveit, 1990; Andersen et al., 1991; Fossen, 2010; Hacker et al., 2010). Despite continuing convergence in the Zagros, tomography data indicate slab break-off in the NW Zagros (Regard et al., 2005; Simmons et al., 2011). Numerical modelling results suggest that slab break-off started at 16 Myr after the onset of the collision in the NW Zagros and propagated along strike with time (van Hunen and Allen, 2011).

Many factors and processes clearly play a role in determining the evolution of
collision systems. In the next two chapters, we investigate controls on styles of continental collision with a series of numerical models.

References


3. A numerical investigation of continental collision styles
3.1 Abstract

Continental collision after closure of an ocean can lead to different deformation styles: subduction of continental crust and lithosphere, lithospheric thickening, folding of the unsubducted continents, Rayleigh-Taylor (RT) instabilities, and/or slab break-off. We use 2-D thermo-mechanical models of oceanic subduction followed by continental collision to investigate the sensitivity of these collision styles to driving velocity, crustal and lithospheric temperature, continental rheology, and the initial density difference between the oceanic lithosphere and the asthenosphere. We find that these parameters influence the collision system, but that driving velocity, rheology, and lithospheric (rather than Moho and mantle) temperature can be classified as important controls, whereas reasonable variations in the initial density contrast between oceanic lithosphere and asthenosphere are not necessarily important. Stable continental subduction occurs over a relatively large range of values of driving velocity and lithospheric temperature. Fast and cold systems are more likely to show folding, whereas slow and warm systems can experience RT-type dripping. Our results show that a continent with a strong upper crust can experience subduction of the entire crust and is more likely to fold. Accretion of the upper crust at the trench is feasible when the upper crust has a moderate to weak strength, whereas the entire crust can be scraped-off in the case of a weak lower crust. We also illustrate that weakening of the lithospheric mantle promotes RT-type of dripping in a collision system. We use a dynamic collision model, in which collision is driven by slab pull only, to illustrate that adjacent plates can play an important role in continental collision systems. In dynamic collision models, exhumation of subducted continental material and sediments is triggered by slab retreat and opening of a subduction channel, which allows upward flow of buoyant materials. Exhumation continues after slab break-off by reverse motion of the subducting plate ("eduction") caused by the reduced slab pull. We illustrate how a simple force balance of slab pull, slab push, slab bending, viscous resistance, and buoyancy can explain the different collision styles caused by variations in velocity, temperature, rheology, density differences, and the interaction with adjacent plates.

3.2 Introduction

Continent-continent collision occurs when an intervening ocean has been closed by subduction of the oceanic lithosphere. Continental collision systems in the Himalayas, European Alps, Norwegian Caledonides, and Zagros show that this process can be accompanied by crustal thickening, magmatism, metamorphism, the formation of thrust nappes and high topography, and the exhumation of (Ultra-) High Pressure ((U)HP) rocks from large depths (Dewey and Burke, 1973; Molnar and Tapponnier, 1975; Pfiffner et al., 1990; Andersen et al., 1991; Blanc et al., 2003; Leech et al., 2005). UHP rocks record pressures of up to 2-3.5 GPa and temperatures of up to 600-800°C (Jamtveit et al., 1991; Andersen et al., 1991; Hacker, 2006; Hacker et al., 2010). They are exhumed during subduction or collision through mechanisms that are still debated, but that may likely combine internal drivers, such as, buoyancy differences, and external drivers, such as, reverse motion in the subduction channel, crustal-scale extension, and squeezing of weak material between two stronger blocks (Warren, 2012). Previous studies have pointed out that continental collision can be accommodated in different styles (Houseman and Molnar, 1997; Cloetingh et al., 1999; Burg and Podladchikov, 2000; Toussaint et al., 2004a) (Fig. 3.1): [1] Stable subduction of continental lithosphere: Even though continental lithosphere is positively buoyant, continental material has been interpreted to have subducted to depths of 100-250 km (Andersen et al., 1991; Ye et al., 2000; Ding et al., 2003). Continental lithosphere could, for example, be pulled down entirely by an attached negatively buoyant oceanic slab. Or the overall positive buoyancy of the continental lithosphere could be reduced by offscraping part of the continental crust at the trench or delaminating the mantle lithosphere from the crust (Bird, 1979), which would facilitate subduction of the remaining thinned crust and underlying lithospheric mantle. Continental subduction is favoured for systems in which the continent is strong, driving velocity is high, and the subduction fault (or channel) is weak (Toussaint et al., 2004a; Selzer et al., 2008). [2] Thickening of crust and lithosphere: Most collision systems experience crustal thickening as evidenced by a deep Moho. This thickening is accommodated by faulting, formation of nappes, and/or a pure-shear style of deformation (Tapponnier et al., 1986; Roecker et al., 1987; Andersen et al., 1991). [3] Lithospheric-scale folding: This is a basic response of the continental plate to large-scale shortening which becomes feasible when there is a strong coupling between crust and mantle and/or a strong subduction fault (Burg and Podladchikov, 2000; Toussaint et al., 2004a; Luth et al., 2010). [4] Rayleigh-Taylor (RT) instabilities: Continents with a weak lithospheric mantle, either because of a warm geotherm or because of weak rheological properties, can...
experience RT-type of lithospheric dripping (Houseman and Molnar, 1997; Pysklywec et al., 2000; Toussaint et al., 2004a). [5] Slab break-off: If the transition from subduction to continent-continent collision is accompanied by subduction of positively buoyant continental crust and/or locking of the inter-plate contact, a resistance to slab pull arises that can lead to an extensional regime in the subducted slab, which can cause slab break-off (Davies and von Blanckenburg, 1995; Li and Liao, 2002). Slab break-off has been inferred from gaps in the hypo-central distribution of seismicity within subducted slabs (Barazangi et al., 1973; Pascal et al., 1973; Fuchs et al., 1979) and from low velocity regions in tomographic images (Wortel and Spakman, 2000; Hafkenscheid et al., 2006). The process is associated with syn-collision magmatism, surface uplift, and exhumation of UHP rocks (Davies and von Blanckenburg, 1995; Buitert et al., 2002; Gerya et al., 2004; Andrews and Billen, 2009; Duretz et al., 2011).

Using upper mantle-scale viscous-plastic numerical experiments, Pysklywec et al. (2000) showed that plate driving velocity is a primary controlling factor for the deformation style of a collision system. They found that RT-style of dripping is a dominant mechanism for slow systems, whereas stable subduction becomes more dominant with increasing driving velocity. Toussaint et al. (2004a,b) (see also Burov and Yamato, 2008) also employed numerical models on the scale of the upper mantle.
and showed that the style of collision strongly depends on both driving velocity and continental temperature (parameterized as temperature of the Moho). They found that continental subduction is favoured for strong lithospheres ($T_{\text{Moho}} < 550^\circ C$) and fast initial driving velocity ($> 3 - 5 \text{ cm yr}^{-1}$). Pure-shear thickening and lithospheric folding become dominant when the lithosphere is weak because of a warmer geotherm ($550^\circ C < T_{\text{Moho}} < 650^\circ C$) or a slower driving velocity ($< 5 \text{ cm yr}^{-1}$), which allows the lithosphere to warm up by heat conduction. RT-type of instabilities occur only in hot systems ($T_{\text{Moho}} > 800^\circ C$), in which high temperatures reduce the effective viscosity of the continental lithosphere, but the lithosphere still remains denser than the asthenosphere, resulting in a high instability growth-rate. In addition, the rheology of the lower crust may also have a significant effect on continental subduction and the tendency for folding (Toussaint et al., 2004a). A strong lower crust increases the coupling between upper crust and lithospheric mantle and results in a system that is more likely to experience folding. A strong lower crust also promotes subduction of the upper crust in a situation of stable subduction, whereas the upper crust can detach and accumulate at the surface or at mid-crustal depths for continents with a weak lower crust (Pysklywec and Cruden, 2004; De Franco et al., 2008; Faccenda et al., 2008; Luth et al., 2010). Other factors that could be thought to play a role in controlling the style of continental collision are the rheology of the upper crust and the lithospheric mantle, slab dip (flat slab versus steep slab subduction), the strength of the interface between the converging plates, the role of adjacent plates versus slab pull, surface processes (erosion and sedimentation), buoyancy, melting, and phase changes (von Blanckenburg and Davies, 1995; Ranalli et al., 2000; Pysklywec, 2006; Billen and Hirth, 2007; De Franco et al., 2008; Selzer et al., 2008; Warren et al., 2008; Li et al., 2011; Luth et al., 2010). Previous numerical models of continental collision have often focussed on crustal and lithospheric-scale structures in models with a simplified mantle representation (Beaumont et al., 1996; Stockmal et al., 2007; Selzer et al., 2008) or on the dynamics of the whole system on the scale of at least the upper mantle (Pysklywec et al., 2000; Toussaint et al., 2004a; Burov and Yamato, 2008). Lithosphere-scale models allow a high resolution that can resolve thrust and nappes, but need to simulate mantle behaviour with boundary conditions. Many subduction models on the scale of the upper mantle have in turn more limited resolution in the crust and lithosphere and sometimes employ a free-slip surface boundary, which suppresses the development of vertical topography (e.g. Capitanio et al., 2009; van Hunen and Allen, 2011). In this study, we investigate crustal- and lithospheric-scale deformation in combination with mantle dynamics through numerical models that have a free upper surface and a reasonable resolution ($2 \times 2 \text{ km per element}$) in the lithosphere. We aim to
investigate the sensitivity of collision styles to driving velocity, lithospheric temperature, the strength stratification of the continents, the initial density difference between the oceanic lithospheric mantle and the sub-lithospheric mantle, and forcing by adjacent plates. Our 2-D models are driven by a combination of slab pull and a kinematic lateral boundary condition that simulates the effects of ridge push and movements of surrounding plates. The kinematic condition is maintained during collision, resulting in a kinematically-driven collision style. We study the role of adjacent plates on collision styles with a model in which the kinematic boundary condition is removed once the system is self-sustaining (Beaumont et al., 2010). This simulates a collision system isolated from surrounding plates, which we call a dynamic collision model.

3.3 Numerical approach

3.3.1 Modelling method

We solve the standard equations for conservation of mass, momentum, and energy for incompressible slow creeping flows:

\[ \nabla \cdot \mathbf{u} = 0 \]  (3.1)

\[ \nabla \cdot \sigma' - \nabla P + \rho g = 0 \]  (3.2)

\[ \rho c_p \left( \frac{\partial T}{\partial t} + \mathbf{u} \cdot \nabla T \right) = k \nabla^2 T + Q \]  (3.3)

\( \mathbf{u} \) is the velocity vector, \( \sigma' \) the deviatoric stress tensor, \( P \) dynamic pressure (mean stress), \( \rho \) density, \( g \) gravitational acceleration \( (g_x = 0, g_y = 9.8 \text{ m s}^{-2}) \), \( c_p \) specific heat, \( T \) temperature, \( t \) time, \( k \) thermal conductivity, and \( Q \) heat generation. We ignore shear heating. We use the 2-D version of the numerical code SULEC v3.4, which uses the Arbitrary Lagrangian-Eulerian (ALE) finite element method. Pressure is solved using the iterative penalty method (Cuvelier, 1986; Pelletier et al., 1989; Zienkiewicz and Taylor, 2000), in which the continuity equation (3.1) is perturbed as following:

\[ P^{\text{new}} = P^{\text{old}} - \kappa \nabla \cdot \mathbf{u} \]  (3.4)

\( \kappa \) is the compressibility factor (generally taken 8 orders of magnitude larger than the largest viscosity in the model) (Zienkiewicz and Taylor, 2000). The perturbed continuity equation (3.4) and the momentum equation (3.2) are solved in an iterative
way to minimise the norms of two successive velocity and pressure values to the desired tolerance. Equations 3.2 - 3.4 are solved on an Eulerian grid which can be vertically stretched or shrunk to accommodate free surface displacements (Fullsack, 1995). Our models have a true free surface which includes a stabilisation term to suppress numerical overshoot across interfaces with large density contrasts (Kaus et al., 2010; Quinquis et al., 2011). At the surface, diffusive erosion and sedimentation is applied with a diffusion coefficient of $5 \times 10^{-6}$ m$^2$ s$^{-1}$ (Culling, 1960). We use a quadrilateral element with four velocity nodes (with two degree of freedom each) and constant pressure, which gives continuous velocity and discontinuous pressure between elements. Material properties are stored on markers which are advected at the end of each time step. We use the extended Boussinesq approach, in which density is temperature-dependent:

$$\rho = \rho_0 (1 - \alpha (T - T_0))$$

(3.5)

$\rho_0$ is the density at temperature $T_0$ and $\alpha$ is the coefficient of volume expansion. SULEC uses the direct solver PARDISO (Schenk and Gartner, 2004) to solve the mechanical and thermal equations.

In our models, materials deform by viscous flow or brittle behaviour, as determined by the mechanism that requires least effective stress. Viscous behaviour is described by a power-law relation of viscosity, strain-rate, pressure, and temperature:

$$\eta = s_c \frac{1}{2} A^{1/2} \dot{\varepsilon}^{(1/2)-1} d^{(1/n-1)} n (1/n - 1)^{-1/2} C_{OH}^{r/v} e^{(\frac{r+pV}{R})}$$

(3.6)

$\eta$ is viscosity, $s_c$ a scaling factor, $A$ a constant, $\dot{\varepsilon}_{II}$ effective strain-rate ($\dot{\varepsilon}_{II} = (\frac{1}{2} \dot{\varepsilon}_{ij} \dot{\varepsilon}_{ij})^\frac{1}{2}$), $n$ stress exponent, $d$ grain size, $p$ grain size exponent, $C_{OH}$ water content, $r$ water exponent, $E$ activation energy, $V$ activation volume, and $R$ the gas constant. We use dislocation creep flow laws of wet quartzite (Gleason and Tullis, 1995) for the upper continental crust, wet anorthite (Rybacki et al., 2006) for the lower continental crust, gabbro (Wilks and Carter, 1990) for the oceanic crust, and wet olivine (Hirth and Kohlstedt, 2003) for the oceanic and continental lithospheric mantle. For the sub-lithospheric mantle, we use composite dislocation and diffusion creep of wet olivine (Hirth and Kohlstedt, 2003). In the latter case, it is assumed that both the dislocation (dis) and diffusion (diff) mechanisms of olivine provide a portion of the total deformation. The composite (comp) viscosity is computed from (van den Berg et al., 1993):

$$\frac{1}{\eta_{comp}} = \frac{1}{\eta_{dis}} + \frac{1}{\eta_{diff}}$$

(3.7)
Brittle behaviour is described by the Drucker-Prager relation:

$$\sigma'_{II} = P \sin \phi + C \cos \phi$$  \hspace{1cm} (3.8)

Where $\sigma'_{II}$ is the effective deviatoric stress ($\sigma'_{II} = (\frac{1}{2}(\sigma'_{ij} \sigma'_{ij}))^{\frac{1}{2}}$), $\phi$ angle of internal friction, $C$ cohesion and $P$ dynamic pressure. We use linear softening of the angle of internal friction from $\phi_1$ to $\phi_2$ in an interval of finite strain values (0.5 - 1.0), described by the second invariant of the strain tensor ($\epsilon_{II} = (\frac{1}{2}(\epsilon_{ij} \epsilon_{ij}))^{\frac{1}{2}}$), as a mechanism for localisation of strain in our models (Beaumont et al., 1996). This simulates the effects of mineral transformations, development of foliation, or grain size changes that can occur in natural shear zones (Mandl, 1988; Rice, 1992; Bos and Spiers, 2002). The magnitude of the softening of the angle of internal friction follows Bos and Spiers (2002) who suggest a decrease in friction coefficient by approximately a factor 2, based on a microphysical model for shear deformation of foliated, phyllosilicate-bearing fault rock. An alternative approach to achieve localisation of deformation in numerical models is through the introduction of shear heating (Regenauer-Lieb and Yuen, 2003). This softening mechanism operates on the viscous deformation field, whereas our strain softening operates on the brittle field. The effectiveness of shear heating can be seen from the dimensionless number $L_o$ (Kaus and Podladchikov, 2006; Braeck et al., 2009; Crameri and Kaus, 2010), whereby localisation through shear heating is effective if $L_o \geq 1$:

$$L_o = \frac{\dot{\epsilon}_{bg} L}{1.4 \sqrt{\eta_0 E nRT_0^2 k}}$$  \hspace{1cm} (3.9)

$\dot{\epsilon}_{bg}$ is the back ground strain-rate, $L$ characteristic length of heterogeneity, and $\eta_0$ the viscosity at reference temperature $T_0$. $E$, $n$, $R$, and $k$ are as above in equations 3.3 and 3.6. All parameters of this equation are known for a specific model set-up, except $L$. Crameri and Kaus (2010) pointed out that localisation could be predicted if $L$ is considered as the brittle thickness of the lithosphere. Substitution of corresponding values from Table 3.1, using a characteristic length $L$ of 20 km, and a back ground strain-rate of $10^{-15}$ s$^{-1}$ results in a value less than one. Shear heating may, therefore, play a role in models similar to ours, but is not expected to represent a dominant process for our setup. All rheological and thermal parameters are listed in Table 3.1.
Table 3.1: Rheological and thermal parameters of the models.

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(e) Adjusted to a general state of stress (Ranalli, 1995).
3.3.2 Model set-up

Our model is 2000 km wide by 660 km deep and contains two continents with an intervening ocean (Fig. 3.2a). Tests with model domains up to 3000 km width (for the same depth of 660 km) show the same subduction and collision evolution, confirming that the width of 2000 km is sufficient for our models. The oceanic slab is about to start subduction and to initiate this, the model has a weak seed wide enough (∼4 - 5 elements) to localise strain along the leftmost continent-ocean interface. The dip of the seed and its adjacent continent-ocean interface is 38°. This dip angle leads to a steeply dipping subducted slab in our models, similar to other models showing that initial conditions can impact slab dip angle during later evolution stages (Buiter and Ellis, 2010; Li et al., 2011). The continents have a 20 km thick upper crust, 20 km thick lower crust, and 80 km thick lithospheric mantle overlying the sublithospheric mantle. The intervening ocean is initially 550 km wide and consists of a 8 km thick oceanic crust and a 72 km thick oceanic mantle lithosphere. We use a non-uniform mesh with a resolution which increases vertically upward in order to reach 2 km in the crust and lithosphere. The highest horizontal resolution (2 km) is imposed around the trench where subduction and later collision is expected to happen. The total number of elements in our model is $338 \times 124$ and we use at least nine tracers in each elements (377208 tracers initially, a number which changes slightly because tracer injection or deletion keeps tracer density per element between 9 and 20).

The initial temperature of the continental domains is computed by solving the steady-state conductive heat equation (Chapman, 1986). This results in a Moho temperature of 586 °C and a lithospheric temperature at 120 km depth of 1300 °C for our reference model. The oceanic domain has an initial geotherm determined from the plate cooling model (Turcotte and Schubert, 2002) for a 60 Ma old ocean (Fig. 3.2b). A transition zone provides a gradual change from the initial continental to oceanic geotherms. The temperature is held fixed at 0 °C at the surface during model evolution. The heat flux is 25 mW m$^{-2}$ at the bottom (at 660 km depth) and zero at the sides. The mantle adiabat is 0.3 °C km$^{-1}$. We use a high thermal conductivity value ($k = 83.33$ W m$^{-1}$ K$^{-1}$) for the sub-lithospheric mantle deeper than 120 km to keep the adiabatic mantle geotherm and heat flux into the base of the lithosphere nearly constant (Pysklywec and Beaumont, 2004). Mantle above 120 km depth has a lower conductivity ($k = 2.3$ W m$^{-1}$ K$^{-1}$).
Figure 3.2: a) kinematically-driven collision reference model setup showing two continents with an intervening 60 Ma ocean. M = mantle, SE = sediment, O = oceanic crust, CL = continental lithosphere, OL = oceanic lithosphere, CUC = continental upper crust, CLC = continental lower crust, WS = weak seed. b) Initial thermal profiles. O = ocean, C = continent. c) Dynamic collision model setup. d) Initial effective stress profiles.
We use two set-ups which differ in the displacement freedom of the continent on the subducting plate. In the reference model, the rightmost continent is pushed with an inward velocity (4 cm yr\(^{-1}\)) and the leftmost continent is held fixed. We call this the kinematically-driven collision model (Fig. 3.2a). The velocity is applied along the side boundary to 120 km depth. This inflow is balanced by outflow on both side boundaries below the lithosphere (120 - 660 km). The change from inflow to outflow is applied gradually over four elements in order to avoid building up of high strains. The vertical component of velocity on the side boundaries is free. The model has a free slip boundary condition at the base and a true free surface at the top.

Our second set-up differs from the kinematically-driven collision model in that the continent on the subducting plate now has a free right-hand side boundary (Fig. 3.2c). We call this model the dynamic collision model. The right-hand continent extends from x = 1400 km to 1800 km and is therefore 200 km shorter than in the kinematically-driven collision model. During the first steps of oceanic subduction, the continent grows by conversion of ocean to continental material at x = 1800 km. We simulate a very simple mid-ocean ridge at the top right-hand side of the model by enforcing oceanic crust formation in the domain from x=1800 km to 2000 km, with a 8 km thick oceanic crust which is underlain by mantle with a thermal conductivity of 2.3 W m\(^{-1}\) K\(^{-1}\) to 120 km depth. This material has an initial temperature equal to the mantle at 120 km depth (1300 °C). The right side boundary also has T = 1300 °C from the surface to 120 km depth. In the initial stages of the model, an internal constant velocity (4 cm yr\(^{-1}\)) is applied within the right continent at x = 1600 km in order to start subduction. This velocity is removed when the system gained enough slab pull to be driven self-consistently. This is typically after 7 Myr of convergence.

Our two set-ups are end-members in the manner in which they address the interaction between the internal forces of slab pull and mantle flow and the external constraints supplied by surrounding plates. We realise that natural subduction zones will lie somewhere in between, but these end-members allow us to examine the case in which surrounding plates influence the subducting plate, providing a constant driving velocity (kinematically-driven collision model), and the case in which surrounding plates have negligible influence on the subducting plate and slab pull is dominant (dynamic collision model).
3.4 Evolution of the reference models

3.4.1 Oceanic subduction

We first show the evolution of a model in which a 60 Ma oceanic plate (without attached continent) subducts underneath a continent (model M1, Table 3.2, and Fig. 3.3). This provides a reference against which we can test the effects of continental collision. The oceanic and continental domains in our models initially have a same surface elevation of zero. The models need therefore to be run for a few time steps in order to achieve a regional isostatic surface topography which is determined by the temperature-dependent density and internal strength of the layers. Model M1 achieves a topographic difference between the continental and oceanic domains of about 5 km after ca. twenty thousand years. This agrees with the topography difference calculated from Airy isostasy and with general topographic and bathymetric data across passive margins (Turcotte and Schubert, 2002). The evolution of surface topography during subduction is shown in Fig. 3.4. It shows a deep trench with subdued surface topography on the overriding and subducting plates. Subduction of the oceanic slab starts by forming a low viscosity and high strain band within the pre-existing weak seed that helps the oceanic slab to decouple from the continent. The oceanic plate starts to subduct with a dip of about 40° that gradually increases as the slab pull becomes more dominant (Fig. 3.3a - 3.3d). The oceanic slab subducts in a bend-backwards mode. This is similar to a subduction mode seen in other numerical and analogue studies (Guillaume et al., 2009; Li et al., 2011; Schellart, 2005; Quinquis et al., 2011; van Hunen and Allen, 2011). Placed in the context of previous studies that relate slab strength to slab bending behaviour, the bend-backwards mode indicates that our oceanic slab is relatively (but not overly) stiff (Schellart, 2008).

3.4.2 Continent-continent collision

The evolution of the kinematically-driven continental collision reference model (M2, set-up in Fig. 3.2a) is shown in Figure 3.5. It shows a transition from initiation of subduction, via a phase of oceanic subduction, to collision. Before collision, the model behaves similar to the oceanic subduction model M1 (compare Fig. 3.3a - 3.3c with Fig. 3.5a - 3.5c). At about 13 Ma, the oceanic slab has completely subducted and collision starts. Model M2 shows a combination of three collision styles: [1] Continental subduction: The upper continental crust is scraped off from
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<th>$T_{lithosphere}$ (°C)</th>
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<th>$\Delta \rho_{30}$ (kg m$^{-3}$)</th>
<th>Mantle adiabat ($^\circ$C km$^{-1}$)</th>
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<th>LC$^a$</th>
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<td>11,12,13,15</td>
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$^a$ Initial contrast in density between unsubducted oceanic lithosphere and underlying asthenosphere.

$^b$ Scaling factor ($s_c$ in equation 6) for continental upper crust (UC), lower crust (LC), and lithospheric mantle (LM).
the lower crust and accretes at the trench to produce a high topography of about 5 km (Fig. 3.5f), but the lower crust subducts together with the lithospheric mantle. Subduction of continental material continues because of the continuous kinematic boundary condition at the right-hand side of the model domain. The topography increases, but never reaches values higher than 5 km (Fig. 3.5f). [2] Folding: The
right continent shows the onset of large-wavelength (~ 300 km), low-amplitude (~ 0.1 km) folding as shown in Fig. 3.5d - 3.5f. Folding appears in the model 9 Myr after initiation of continental collision and progressively increases in time, but remains of low amplitude. [3] Thickening: Continental thickening is accommodated by off-scraping of the upper continental crust at the trench and formation of a frontal accretionary wedge (Fig. 3.5e). The accretionary wedge grows during the collision phase. Sediments and crustal material that are not off-scraped subduct to sub-crustal depths. Our kinematically-driven collision models do not show evidence of exhumation of high pressure rocks.

Our results are consistent with previous studies (Pysklywec et al., 2000; Toussaint
et al., 2004a; Burov and Yamato, 2008; Selzer et al., 2008), which found that a fast driving velocity (> 3 – 5 cm yr⁻¹), a weak lower continental crust, and cold to moderate Moho temperatures (< 600 °C) can result in a combination of stable subduction, lithospheric-scale folding, and thickening, but that stable subduction will be the dominant style of deformation for these systems.

### 3.5 Controls on collision styles

#### 3.5.1 The influence of driving velocity and temperature

Previous studies have shown that driving velocity and temperature are two primary factors which can control the style of collision (Pysklywec et al., 2000; Toussaint et al., 2004a; Burov and Yamato, 2008). Toussaint et al. (2004a,b) investigated the effect of continental temperature using models with an initial Moho temperature which was varied between 400 °C and 1000 °C, while the surface temperature was kept at 0 °C. Their models have 1330 °C at the base of the asthenosphere at 250 km depth, and 2200 °C at 660 km depth. This implies a hot mantle with an adiabat of 2.12 °C km⁻¹, which is about one order of magnitude larger than the expected adiabat of the mantle (Turcotte and Schubert, 2002). Because temperatures of lithological layers cannot be varied individually, but are linked through the geothermal gradient, not only crustal temperature is varied, but also the temperature of the lithospheric mantle. It is therefore an open question whether the different styles of the models are caused by changes in crustal temperature, changes in lithospheric temperature, or both.

To investigate the roles of crustal and lithospheric mantle temperature, we first set up cold (M3) and warm (M4) continental crust models using the thermal parameters of Table 3.3 (see also Table 3.2). The initial Moho temperature for our cold and warm models is 480 °C and 680 °C, respectively (our reference model M2 has a Moho temperature of 586 °C). Surface temperature is 0 °C, and temperature at the base of the continental lithosphere is 1300 °C (Fig. 3.6a). Our models also dif-
Figure 3.6: Initial thermal structure of the continents for models M2-M12 that examine the effects of variations in crustal and lithospheric temperature. a) Thermal structure for model M3 with a cold crust ($T_{Moho} = 480 \degree C$), reference model M2 with an intermediate temperature crust ($T_{Moho} = 586 \degree C$), and model M4 with a hot crust ($T_{Moho} = 680 \degree C$). Temperature is fixed at 1300 $\degree C$ at the base of the lithosphere (at 120 km depth). b) Thermal structure for models M2 and M5-12 in which temperature is varied in the entire lithosphere and mantle, from a cold lithosphere ($T_{lc} = 1200 \degree C$, M5-7), through an intermediate lithosphere ($T_{lc} = 1300 \degree C$, M2 and M8-9), to a hot lithosphere ($T_{lc} = 1400 \degree C$, M10-12). c) Thermal structure for models with an intermediate temperature crust and lithosphere (as in M2) and a cold (M15), intermediate (M2), and hot (M16) mantle.

Figure 3.7: Effect of Moho temperature on continental collision in the kinematically-driven collision model (see also Fig. 3.6a) Results after 20 Myr of convergence. a) Cold crust model (M3) with a Moho temperature of 480 $\degree C$. b) Hot crust model (M4) with a Moho temperature of 680 $\degree C$. Both models show essentially the same style of deformation, but the warmer (weaker) crust model builds less topography.

In lithospheric temperature, but the largest temperature difference occurs at the Moho. They have the same sub-lithospheric mantle temperature which follows a $0.3 \degree C \text{km}^{-1}$ mantle adiabat. Models M3 and M4 give the largest variation in Moho temperature that can be achieved within a realistic range of thermal parameters. Our Moho temperature variation is less than in Toussaint et al. (2004a). Figure 3.7 shows that the effect of Moho temperature on the style of continental collision is essentially small over the range of temperatures we investigated. The models show stable continental subduction and the formation of an accretionary wedge at the trench. The warmer (and therefore rheologically weaker) crust model can not sustain the high topography of the colder (stronger) model (Fig. 3.7). This can be understood as the
Figure 3.8: Influence of continental temperature, as characterised by the temperature at the base of the lithosphere (Fig. 3.6b), and driving velocity on the kinematically-driven collision model. Results show material and velocity fields after 24 Myr of convergence, with surface topography at the top of each panel.
response of an accretionary wedge building above a weak base. For low basal dip angles, a reduction in basal strength leads to a reduction in topography (Davies and Engelder, 1985; Huiqi et al., 1992; Buiter, 2002). Our results indicate that different styles of collision are probably not only caused by variations in the temperature of the Moho (within reasonable thermal values) and that other parameters may play a role.

We therefore next investigate the role of crustal and mantle temperature by varying the initial geotherm from a cool geotherm, with \( T_{lc} = 1200 \, ^\circ C \) at the base of the continental lithosphere and \( T_{lo} = 1188 \, ^\circ C \) at the base of the oceanic lithosphere (models M5 - M7), to a warm geotherm with \( T_{lc} = 1400 \, ^\circ C \) and \( T_{lo} = 1388 \, ^\circ C \) (models M10 - M12) (Fig. 3.6b). The largest temperature difference between these models occurs in the lithospheric and sub-lithospheric mantle. For these models we also examine the role of driving velocity by varying the velocity of the incoming plate between 1 cm yr\(^{-1}\) and 8 cm yr\(^{-1}\) (Table 3.2). Figure 3.8 shows the models after 24 Myr of convergence in temperature-velocity space. The models show almost all collision mechanisms: stable subduction, thickening, folding, and lithospheric dripping. Slab break-off does not occur because the combination of the continuous push applied by the driving boundary velocity with slab pull overcomes the upward buoyancy of the subducted continental materials, for our model setup and parameter values. Fig. 3.8 allows us to deduce the following trends:

[1] Stable continental subduction occurs over a wide range of temperatures (\( T_{lc} = 1200 \, ^\circ C \) to 1400 \( ^\circ C \)) and velocities (between 1 cm yr\(^{-1}\) and 8 cm yr\(^{-1}\)).

[2] Lithospheric folding plays a role only in cool systems (\( T_{lc} \approx 1200 \, ^\circ C, T_{moho} = 484 - 685 \, ^\circ C \)) at high driving velocities (above 4 cm yr\(^{-1}\)).

[3] RT-type of dripping of continental lithosphere is seen only in hot systems (\( T_{lc} \geq 1350 \, ^\circ C \)). RT-dripping is more effective at low driving velocities, as the lower velocity implies a lower viscous strength of the continental lithosphere.

[4] Lithospheric thickening occurs by two mechanisms: First, thickening in the trench region associated with off-scraping of the crust and accumulation of crustal material at the trench occurs in all models of continental subduction. Second, thickening of the continents accommodated by pure shear occurs in models with a hot geotherm (\( T_{lc} = 1350 \, ^\circ C \)).

These models illustrate that crust and mantle temperature (rather than Moho temperature alone) and driving velocity play crucial roles in determining continental collision styles. Since both lithospheric and mantle temperature change from a cold
to a hot geotherm in our models (see Fig. 3.6b), it is an open question whether lithospheric temperature or mantle temperature is the primary factor. To address this question, we varied mantle temperature for a fixed lithospheric temperature (Fig. 3.6c). Mantle temperature changes from cold (mantle adiabat = 0.2 °C km$^{-1}$), via intermediate (mantle adiabat = 0.3 °C km$^{-1}$), to hot (mantle adiabat = 0.4 °C km$^{-1}$) for a cold (M6), intermediate (M2), and hot (M11) lithosphere model (resulting in 6 extra models) (Fig. 3.16). We find that sub-lithospheric thermal variations in a realistic range do not effect the style of collision for a cold and intermediate temperature lithosphere on the time-scale of our simulations. An increase in the mantle temperature for a warm lithosphere ($T_{lc} = 1400$ °C) facilitates RT-type dripping, because the warm mantle further warms up the continental lithosphere, inducing additional weakening of the lithosphere from the base (we come back to this in the discussion). Based on our model results, we can therefore conclude that lithospheric temperature (rather than crustal or mantle temperature alone) constitutes an important control on collision style.

3.5.2 The influence of continental rheology

Temperature variations imply only a limited strength variation (within realistic values for thermal parameters) and it is difficult to vary crustal and mantle temperature separately. We therefore here investigate a wider range of variations in continental strength by varying the rheological stratification of the continents in the kinematically-driven collision model (Fig. 3.9). The driving velocity is again 4 cm yr$^{-1}$ and $T_{lc}$ is 1300 °C. The rheological strength of the continental upper crust, lower crust, and lithospheric mantle is varied by multiplying the viscosity with a factor $s_c$ following Eq. 3.6 (Table 3.2). The rheology of each continental layer is changed separately, keeping the other two layers the same as in the reference model. The strength of the oceanic lithosphere is the same for all these models.

The upper crust is scraped off from the lower crust and accumulated at the trench to form an accretionary wedge for models with an upper crust of moderate to weak strength (models M2 and M19). An increase in upper crustal strength (model M20) facilitates subduction of the entire continental crust to lithospheric depths. Continents with a strong upper crust are also more likely to experience folding during collision (model M20). Our results show that in the case of a weak lower crust (model M21), the entire continental crust decouples from the subducting lithospheric mantle and accumulates at the trench. This increases crustal thickening, but stable subduction remains the dominant style of deformation. Weakening of the continental lithospheric mantle (model M23) promotes pure shear-type of thickening and RT-type
Figure 3.9: a) kinematically-driven collision models after 24 Myr convergence showing the effects of variations in strength of the continental crust and lithospheric mantle. Strength variations are obtained by multiplying the viscous flow-laws with a scaling factor (see Eq. 3.6, see also Table 3.2). Variations in strength of the upper crust in the top row, lower crust in the middle row, and lithospheric mantle in the bottom row. Left column weak case, middle column intermediate case, and right column strong case. Surface topography is shown at the top of each panel. b) Initial continental effective stress profiles.
of lithospheric dripping, whereas stable subduction occurs for stronger lithospheres (models M2 and M24). Our variations in continental strength are large (Fig. 3.9b) and could in nature only be achieved by lithological differences or localised weakening mechanisms, as changes in temperature or velocity would not effect individual layers only. The strength of the upper and lower crust controls continental crust subductability, whereas the strength of the lithospheric mantle controls lithospheric thickening, RT-style dripping, and subduction of the continental lithosphere.

### 3.5.3 The influence of adjacent plates

Our models so far all examined kinematically-driven collision (Fig. 3.2a), in which convergence across the inter-plate contact is maintained by a driving velocity which simulates a combination of a push by adjacent plates and ridge push. We investigate the role of adjacent plates with a dynamic collision setup (Fig. 3.2c), in which the kinematic boundary condition is removed once the system is self-sustaining. The first steps of the dynamic collision model are similar to the kinematically-driven model (Figs. 3.5 and 3.10). We initiate subduction and drive the rightmost continent towards the trench by applying a driving velocity of 4 cm yr\(^{-1}\). After 7 Myr, we remove the imposed velocity, leaving the model to develop internally. The production of new continental material at x = 1800 km also ceases at this stage. Since the rightmost continent is decoupled from the side boundary, the continent is now more free to move in response to the competition between slab pull, mantle resistance, and positive buoyancy of subducted continental material.

Subduction of the oceanic slab continues until the whole slab is subducted and the continent starts to subduct (Fig. 3.10b). From this stage on, the convergence velocity between the two continents decreases owing to the increased resistance to subduction caused by the subducting positively buoyant continental material (Fig. 3.11a). The competition between slab pull and positive buoyancy of the subducting continent leads to an extensional regime in the subducted slab, finally resulting in necking and slab break-off similar to, for example, Andrews and Billen (2009) and Duretz et al. (2012) (Fig. 3.10c). Our dynamic collision reference model experiences a deep slab break-off (280 km depth) at about 18 Myr after initiation of collision. Slab break-off is accompanied by rapid surface uplift at a rate of ca. 0.6 km Myr\(^{-1}\), which is in the range of post break-off uplift rates found in previous studies (Andrews and Billen, 2009; Duretz et al., 2011). Maximum surface topography reaches about 4.2 km at ca. 2 million years after break-off, which is similar to break-off related uplift found in previous studies (Buiter et al., 2002; Beaumont et al., 2010; Duretz et al., 2011). Maximum topography is therefore similar in the kinematically-
Figure 3.10: Evolution of the dynamic continental collision model (M27, Fig. 3.2c). Material properties (left) and logarithmic viscosity snapshots (right) after a) 8.7, b) 13.3, c) 21.5, d) 32.0, and 51.7 Myr. The right continent is driven with $v = 4 \text{ cm yr}^{-1}$ until $t = 7 \text{ Myr}$, after which the driving velocity is removed and slab pull is the only driving force in the model. The two isotherms at 586 $^\circ \text{C}$ and 1300 $^\circ \text{C}$ represent the temperature at the Moho and base of the continental lithosphere in the initial state, respectively.

Figure 3.11: a) Relative velocity between the two continents of the dynamic collision model (M27) plotted versus time. b) Surface topography of the dynamic collision model (M27) in the oceanic subduction (13.3 Myr), necking (27.8 Myr), break-off (32.0 Myr), and reverse motion (56.0 Myr) phases.

83
driven (M2) and dynamic (M27) collision models (Figs. 3.5f and 3.11b), but the mechanisms creating surface uplift are different: The kinematically-driven collision model creates a long-term topography through accretionary wedge formation and the maximum topography increases with time (Fig. 3.12). In contrast, topography in the dynamic collision model is supported dynamically and increases after break-off before slowly reducing to ca. 2 km over a long period of time (∼24 Myr) (Fig. 3.12). The topography reduction occurs when the formerly subducting slab experiences about 90 km of reverse motion termed ‘eduction’ (Andersen et al., 1991) (measured horizontally along the surface) (Figs. 3.11b and 3.12). This amount of reverse motion is smaller in magnitude than the eduction of 180 km that Andersen et al. (1991) determined for the Norwegian Caledonides.

Exhumation of subducted continental crust and sediments is triggered before slab break-off by opening of a subduction channel due to slab retreat and upward buoyancy flow of crustal material and sediments in the opened channel. Although exhumation initiates before break-off, most exhumation occurs after break-off when the model experiences reverse motion. Subducted materials experience peak pressures of ca. 2.5 GPa and temperatures of ca. 580°C before exhumation, which is in the range of natural peak pressure and temperature of UHP rocks reported by previous workers (Jamtveit et al., 1991; Andersen et al., 1991; Hacker, 2006; Hacker et al., 2010). It has been suggested that confinement of weaker rocks in a constrained flow between strong materials (that ‘high-pressure cooker’) could lead to non-lithostatic pressure of up to 1-2 GPa (Mancktelow, 2008; Li et al., 2009; Vrijmoed et al., 2009). This would imply that UHP conditions are reached at shallower depth than in a lithostatic pressure environment. The development of dynamic pressure depends on the crustal and mantle rheology and convergence velocity of the
plates (Mancktelow, 2008; Li et al., 2009). Our dynamic collision reference model does not show significant non-lithostatic pressure (order of few MPa in magnitude) and UHP rocks in our models record therefore lithostatic pressure. Our models show that collision systems could experience a relatively large control by their surrounding plates. Forced convergence can cause deep subduction of continental material. Less control by adjacent plates can lead to exhumation of deep subducted rocks and slab break-off.

3.5.4 The influence of the density of the oceanic lithosphere

Many subduction models impose an initial contrast in density between the unsubducted oceanic lithosphere and the underlying asthenosphere. This density contrast is modified during model evolution as the subducted slab warms up (Eq. 3.5). The density difference determines slab pull and changes in its value can therefore effect the dynamics of subduction and collision. Previous numerical studies have used values for the initial density difference between oceanic lithospheric mantle and asthenosphere \( \Delta \rho_{la} \) between 0 and 50 kg m\(^{-3}\) (e.g. Toussaint et al., 2004a; Babeyko and Sobolev, 2008; Warren et al., 2008). Afonso et al. (2007) pointed out that the density difference probably does not exceed 40 kg m\(^{-3}\). We examine the effect of variations in \( \Delta \rho_{la} \) of 20, 30, and 40 kg m\(^{-3}\) by changing the initial density of the oceanic lithospheric mantle from 3270 to 3290 kg m\(^{-3}\). This implies a variation from a weak to a strong slab pull. Our reference models (M2, M27) have a \( \Delta \rho_{la} \) of 30 kg m\(^{-3}\). Our dynamic collision model is not sensitive to density variations of ± 10 kg m\(^{-3}\). Dynamic collision with strong, moderate, and weak slab pull (20, 30, and 40 kg m\(^{-3}\) density contrast) show the same collision style, with break-off being the dominant style of deformation. Results from our kinematically-driven models with the same differences in density contrast also indicate that this order of variation in not powerful enough to make noticeable changes in the style of collision (Fig. 3.13).

3.6 Discussion

We have shown how lithospheric temperature and driving velocity change the deformation style in kinematically-driven collision systems (Fig. 3.8) from RT-dripping at high temperatures (\( T_{le} \geq 1350 \, ^{\circ}C \)), via stable subduction for moderate to high temperatures (\( 1250 \, ^{\circ}C \leq T_{le} \leq 1350 \, ^{\circ}C \)) and for a wide range of driving velocities (\( 1 \, \text{cm yr}^{-1} \leq V_{dv} \leq 8 \, \text{cm yr}^{-1} \)) to folding at low temperatures (\( T_{le} \approx 1200 \, ^{\circ}C \))
and high velocities ($V_{dv} \geq 6 \text{ cm yr}^{-1}$). Our results are consistent with previous studies (Pysklywec et al., 2000; Toussaint et al., 2004a; Burov and Yamato, 2008), except that the boundaries in temperature-velocity space between the different collision styles are somewhat shifted. For example, our models show that folding is likely to occur at the same or cooler geotherm for which stable subduction occurs, whereas previous studies (Toussaint et al., 2004a; Burov and Yamato, 2008) have suggested that folding can also happen in hotter systems. Part of the differences between our study and previous studies is to be expected since differences occur in rheology, density, softening mechanisms, and model setup.

We have illustrated how the collision mechanisms depend on the continental rheological stratification and the initial density contrast between oceanic lithospheric mantle and the asthenosphere. We found that an increase in strength of the upper continental crust facilities subduction of the entire crust to lithospheric depths (Fig. 3.9). An upper crust of moderate to weak strength is more likely to be scraped-off at the trench and contribute to the formation of an accretionary wedge. Accretion of the entire crust is feasible when the lower crust is weak. An intermediate to strong lithospheric mantle leads to subduction, but weakening of the continental lithospheric mantle leads to RT-type of dripping. We found that changes in the initial density contrast between the oceanic lithospheric mantle and underlying mantle on the order of 20 kg m$^{-3}$ do not cause noticeable differences in the style of kinematically-driven collision models (Fig. 3.13).

The transition between lithospheric folding and thickening can be predicted by considering the $Ar^*$ number, which is given as the ratio of stress caused by gravity to

Figure 3.13: Surface topography after 24 Myr convergence for the kinematically-driven collision model with an initial density difference $\Delta \rho_{la}$ between oceanic lithosphere and sub-lithospheric mantle of 20 and 40 kg m$^{-3}$ (the reference model uses $\Delta \rho_{la} = 30$ kg m$^{-3}$).
stress caused by shortening (Schmalholz et al., 2002): 

\[ Ar = \frac{\Delta \rho gH}{2\eta_{eff} \dot{\varepsilon}_{bg}} \]  

\( \Delta \rho \) is the density contrast between lithosphere and asthenosphere, \( H \) and \( \eta_{eff} \) thickness and effective viscosity of lithosphere, and \( \dot{\varepsilon}_{bg} \) back ground strain-rate. Following Schmalholz et al. (2002), folding would dominate over lithospheric thickening if \( Ar < 2.5n \) (using a thick-plate solution). Acknowledging the simplifications inherent in this equation and the values of its parameters, we find that our models can exhibit folding or thickening, with a tendency towards folding. 

To understand the roles of velocity, temperature, rheology, and density we look at the forces in a collision system (Fig. 3.14). The slab pull force can be calculated from the density contrast (\( \Delta \rho_s \)) between subducted oceanic lithosphere and surrounding mantle and the length (\( L_s \)) of the subducted slab (Turcotte and Schubert, 2002): 

\[ F_{sp} \simeq \Delta \rho_s g L_s h \]  

\( h \) is the thickness of the slab. In our models, \( F_{sp} \) is on the order of \( 2 \times 10^{13} \) N m\(^{-1}\) for the oceanic subduction phase. This is consistent with previous studies and analytical calculations that found a slab pull in the range of \( 1-5 \times 10^{13} \) N m\(^{-1}\) (Molnar and Gray, 1979; Bott et al., 1989; Davies and von Blanckenburg, 1995; Turcotte and Schubert, 2002; Funiciello et al., 2003). We computed the density difference \( \Delta \rho_s \) over the entire subducted lithosphere, including crustal material. Slab pull changes with variations in crustal and lithospheric temperature, as in models M3 and M4 (Fig. 3.6a) and for variations in mantle temperature only (models M15 and M16, Fig. 3.6c). However, models in which lithospheric temperature is changed but the temperature at the base of the lithosphere and mantle are shifted by the same amount (models M5-M7 and M10-M12, Fig. 3.6b) have little impact on slab pull, because the relative temperature difference between the slab and surrounding mantle stays
approximately the same. The push force $F_p$ simulates the interaction of the continental plate with adjacent plates in natural settings in a simplified manner and can be calculated from:

$$F_p \approx \int_0^{h_{cl}} \sigma_{xx} \, dy \quad (3.12)$$

$h_{cl}$ is the continental lithospheric thickness and $\sigma_{xx}$ the stress component in the horizontal direction. In our dynamic collision models, the right continent is decoupled from the side boundary and can move more freely in response to slab pull. Only a small 'ridge push' from our simulated ridge acts on the right continent in this case. In kinematically-driven collision models, the right continent is pushed with a constant velocity, which does not imply a constant stress or push force. Therefore, the push force differs most between our kinematically-driven ($1.7 \times 10^{12}$ N m$^{-1}$ toward the trench) and dynamic ($2.8 \times 10^{11}$ N m$^{-1}$ away from the trench) collision models after initiation of continental collision. Our push force magnitudes are within the range of push forces on the order of $1-7 \times 10^{12}$ N m$^{-1}$ found in previous studies (Parsons and Richter, 1980; Toth and Gurnis, 1998; Turcotte and Schubert, 2002; Funiciello et al., 2003). The primary resisting force to subduction is the slab bending force ($F_{be}$). For viscous plates, this force can be calculated from (Turcotte and Schubert, 2002):

$$F_{be} \approx -\frac{u_{conv} h^3 \eta_s}{r^3} \quad (3.13)$$

$\eta_s$ is the slab effective viscosity, $r$ slab curvature at the trench, and $u_{conv}$ is convergence velocity. The minus sign indicates that this force is a resistive force against subduction. The numerical value for $r$ is obtained by fitting the best circle that accommodate top interface of the subducted slab and separate it from overriding continent and its surrounding mantle. In our models, this force is on the order of $3 \times 10^{12}$ N m$^{-1}$ for the oceanic subduction phase, which is of the same order of magnitude as found by Buffett and Becker (2012). The continental collision phase leads to a bending force which is one order of magnitude smaller. The convergence velocity $u_{conv}$ directly changes $F_{be}$, but temperature variations also play a role through their effect on the effective viscosity of the slab. The viscous mantle resistance force ($F_{vr}$) acts on the top and bottom of the subducted slab and at the base of the continental plate (mantle drag). It can be calculated from the viscosity of the mantle ($\eta_m$) and the mean velocity of the incoming continent ($u_{dv}$) (Turcotte and Schubert, 2002):

$$F_{vr} \approx -\eta_m u_{dv} \quad (3.14)$$

This force can affect our model styles when changes occur in mantle rheology, temperature, or convergence rate. The range of this force in our models is from $3 \times 10^{10}$
N m$^{-1}$ to 1×10$^{11}$ N m$^{-1}$. This is of similar order of magnitude as obtained from simple calculations assuming an upper mantle viscosity of 10$^{19}$ - 10$^{21}$ Pa s (e.g. Steinberger and Calderwood, 2006; Lee et al., 2011) and a velocity of 4 cm yr$^{-1}$, which results in 1×10$^{10}$ to 1×10$^{12}$ N m$^{-1}$. When continental positively buoyant material starts to subduct, the buoyancy force ($F_{bo}$) becomes important. This force is proportional to the density contrast ($\Delta \rho_c$) between continental lithosphere and the surrounding mantle (Molnar and Gray, 1979):

$$F_{bo} \simeq -\Delta \rho_c g L_c h_c$$

(3.15)

$L_c$ is the length of subducted continent, and $h_c$ the thickness of continental crust. Previous studies estimated the buoyancy force at ca. 1-5×10$^{13}$ N m$^{-1}$ (Mckenzie, 1969; Molnar and Gray, 1979) and our models show a buoyancy force of the same order of magnitude in the collision phase. The friction force $F_i$ between the over-riding and subducting plates at the subduction channel resists subduction mainly during initiation of subduction, but becomes one order of magnitude smaller than the slab bending force later in model evolution, if the subduction channel has the same viscosity as the mantle (Funiciello et al., 2003). In our models, the effective viscosity at the plate contact is about 10$^{20}$ Pa s and is on the same order of magnitude as the effective viscosity of the mantle. We therefore ignore this force. Figure 3.15 shows push force ($F_p$) versus effective slab pull ($F_{sp} + F_{vr} + F_{be} + F_{bo}$) for four of our numerical models, representative of continental subduction (M2), folding (M7), RT-type dripping (M10), and slab break-off (M27). After about 7 Myr of convergence, enough subduction took place for slab pull to become active in the models. The effective slab pull increases during the oceanic subduction phase. When the subducted oceanic slab reaches the bottom of the model or the whole oceanic slab is subducted and collision starts (at about 14 ± 1 Myr), the effective slab pull decreases. The effective slab pull is initially the main driving force for colder systems (folding model (M7) with $T_{lc} = 1200 \, ^\circ C$), while the push force is slightly more dominant in warmer systems (RT model (M10) with $T_{lc} = 1400 \, ^\circ C$). The push force grows with time for kinematically-driven models (M2, M7, and M10), but the growth rate is low in hot and slow systems and high in cold and fast systems. The push force in our dynamic collision model follows the same path as our reference collision model until 10 Myr, and it decreases afterward to a very small value. Figure 3.15 shows that stable subduction is the dominant mechanism when the forces are more or less in balance: $F_{sp} + F_{vr} + F_{be} + F_{bo} \approx F_p$. Folding and thickening become more likely when the continent is pushed more than it is pulled: $F_{sp} + F_{vr} + F_{be} + F_{bo} < F_p$. If the effective slab pull is stronger than the push force ($F_{sp} + F_{vr} + F_{be} + F_{bo} > F_p$), slab break-off may occur. Slab break-
off exclusively occurred in our dynamic collision model, because the push force in these models is much reduced (to a small contribution by mantle flowing into our artificial mid-ocean ridge at the top-right corner of the model).

None of the driving forces (effective slab pull and push force) plays a strong role in hot and slow systems. RT-type of dripping develops in these systems when the growth time of instabilities is on the order of the simulation time-scale. The time-scale for the formation of lithospheric instabilities ($\tau_b$) can be estimated by (Canright and Morris, 1993; Houseman and Molnar, 1997):

$$\tau_b = \frac{8\eta_{cl}}{\Delta \rho_c g h_{cl}}$$

(3.16)

$\eta_{cl}$ is the effective viscosity of the lower part of the continental lithosphere. The density contrast and thickness do not change much in our models, when the velocity and temperature are varied. The only parameter that effects the time-scale is therefore the viscosity. A model with 1 cm yr$^{-1}$ driving velocity and 1400 °C mantle temperature (Model M10) has an effective viscosity $\eta_{cl}$ of approximately $10^{21}$ Pa s and a RT growth time on the order of 16 Myr, which makes RT-dripping plausible (Fig. 3.8). But a model with 4 cm yr$^{-1}$ driving velocity and 1200 °C mantle temperature (model M6) with $\eta_{cl} \sim 10^{22}$ Pa s has a RT growth time of $\sim 160$ Myr, which is outside the time-scale of our models.
3.7 Conclusions

We have investigated styles of continental collision with 2-D thermo-mechanical models, which simulate collision after a phase of oceanic subduction. We find that collision style is influenced by driving velocity, lithospheric temperature, rheological strength of continental crust and lithosphere, and the interaction with adjacent plates. We show that:

[1] Stable subduction of continental crust and lithosphere can occur over a wide range of lithosphere temperature (base of lithospheric temperatures of 1250 - 1350 °C) and driving velocity (1 - 8 cm yr⁻¹).

[2] Folding is a favoured mechanism for cold and fast systems, whereas RT-type of dripping is more likely to occur in slow and warm systems.

[3] A strong upper crust facilitates subduction of the entire continental crust and part of the overlying sediments to lithospheric depths and promotes folding of the incoming continent, whereas a weak upper crust promotes off-scaping of the upper crust at the trench and formation of an accretionary wedge.

[4] A continent with a weak lower crust can experience accretion of the entire crust at the trench.


[6] Variation of the initial density contrast between the oceanic lithosphere and the asthenosphere by 20 kg m⁻³ does not effect collision style.

[7] Slab break-off is facilitated in settings in which the subducting plate does not experience a large push from adjacent plates.

Our study illustrates how a simple force balance can be used to differentiate between stable subduction, continental folding and thickening, and slab break-off.

3.8 Acknowledgments

This study was supported by the Norwegian Research Council through NFR project 180449/V30. We would like to thank Joya Tetreault, Matthieu Quinquis, and Zurab Chemia for many constructive discussions. The constructive reviews of Thibault Duretz and Stefan Schmalholz helped our revision. Our models were solved with
3.9 Appendix: The influence of mantle temperature

The numerical code SULEC, which was co-developed between Susanne Buiter and Susan Ellis. Most of our figures were made using GMT (Wessel and Smith, 1991).

Figure 3.16: a) kinematically-driven collision models showing the effects of variations in mantle adiabat for a cold \( T_{lc} = 1200 \degree C \), intermediate \( T_{lc} = 1300 \degree C \), and hot lithosphere \( T_{lc} = 1400 \degree C \). The mantle adiabat is varied between 0.2, 0.3, and 0.4 \( \degree C \text{km}^{-1} \). The panels show material and velocity fields after 24 Myr of convergence, with surface topography at the top. b) Initial continental geotherms.
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4. Numerical modelling of the role of salt in continental collision: An application to the southeast Zagros fold-and-thrust belt
Numerical modelling of the role of salt in continental collision: An application to the southeast Zagros fold-and-thrust belt

4.1 Abstract

The Zagros fold-and-thrust belt formed in the collision of Arabia with Central Iran. Its sedimentary sequence is characterised by the presence of several weak layers that may control the style of folding and thrusting. We use 2-D thermo-mechanical models to investigate the role of salt in the southeast Zagros fold-and-thrust belt. We constrain the crustal and lithospheric thickness, sedimentary stratification, convergence velocity, and thermal structure of the models from available geological and geophysical data. We find that the thick basal layer of Hormuz salt in models on the scale of the upper-mantle decouples the overlying sediments from the basement and localises deformation in the sediments by trench-verging shear bands. In the collision stage of the models, basement dips with +1° toward the trench. Including the basal Hormuz salt improves the fit of predicted topography to observed topography. We use the kinematic results and thermal structure of this large-scale model as the initial conditions of a series of upper-crustal-scale models. These models aim to investigate the effects of basal and intervening weak layers, salt strength, basal dip, and lateral salt distribution on deformation style of the simply folded Zagros. Our results show that in addition to the Hormuz salt at the base of the sedimentary cover, at least one intervening weak layer is required to initiate fold-dominated deformation in the southeast Zagros. We find that an upper-crustal-scale model, with a basal and three internal weak layers with viscosities between $5 \times 10^{18} - 10^{19}$ Pa s, and a basement that dips +1° towards the trench, best reproduces present-day topography and the regular folding of the sedimentary layers of the simply folded Zagros.

4.2 Introduction

4.2.1 Salt in the Zagros fold-and-thrust belt

The low mechanical strength of salt layers can play a fundamental role in controlling the deformation style of fold-and-thrust belts (FTBs). Where salt forms a weak

\[ \text{In review with Tectonophysics as: Ghazian R.K. and Buiter S.J.H.} \]
base to a FTB, its low strength reduces surface slope, promotes deformation far into the foreland, and leads to a more symmetric style of thrusting characterised by forward and backward thrusts (Davies and Engelder, 1985; Dahlen, 1990; Cotten and Koyi, 2000; Costa and Vendeville, 2002; Bahrouri and Koyi, 2003; McQuarrie, 2004; Buiter, 2012). Internal salt layers may control the style of folding and faulting, depending on the thickness and strength of the salt layers (Bahrouri and Koyi, 2003; Yamato et al., 2011; Frehner et al., 2012; Moutheau et al., 2012). Salt layers occur in several fold-and-thrust belts around the world, such as Zagros, Jura, the southern Pyrenees, and the Salt Range and Potwar Plateau (Davies and Engelder, 1985; Alavi, 1994; Cotten and Koyi, 2000; Alavi, 2004). Since the Zagros is one of the largest salt diapir provinces in the world, we here use this fold-and-thrust belt as an inspiration to investigate the role of salt in continental collision.

The Zagros FTB (Fig. 4.1a) extends over a vast area, with a width of 200-300 km (NE-SW) and a length of ca. 1800 km (NW-SE). This orogenic system formed when Arabia separated from Africa in the late Cretaceous and converged towards Eurasia, resulting in closure of the Neo-Tethys ocean. The Neo-Tethys Ocean opened in the late Permian (Sengör, 1988; Stampfli, 2000; Muttoni et al., 2009) and at the time of collision the oceanic crust was therefore old. The Main Zagros Thrust Fault (MZTF) (Fig. 4.1b and 4.2b) is assumed to represent the suture of the collision of Arabia with Central Iran. Collision-associated deformation initiated at the MZTF and propagated southwestward through time (Hatzfeld et al., 2010). The timing of continental collision in the Zagros is still debated (Dewey et al., 1973; Falcon, 1974; Scott, 1981; Dercourt et al., 1986; McQuarrie et al., 2003; Alavi, 2004; Agard et al., 2005). Agard et al. (2005) infer that collision started at about 25-23 Ma from dating Blueschist facies rocks found in the Sanandaj-Sirjan zone (Fig. 4.1b), which separates the Zagros FTB from Central Iran. In contrast, arc-related magmatism in Central Iran would support initiation of collision at 34 Ma (Ballato et al., 2011; Moutheau et al., 2012). Since the onset of collision, ca. 500-800 km of convergence has occurred and GPS measurements show that convergence is still ongoing today (Vernant et al., 2004; Hessami et al., 2006; Hatzfeld and Molnar, 2010; Hatzfeld et al., 2010).

Tomography data suggest steepening or break-off of the subducted slab in the NW Zagros, which is likely concurrent with slab break-off in Anatolia at about 12 Ma (Regard et al., 2005; Authemayou et al., 2006; Faccenna et al., 2006; Hafkenscheid et al., 2006; Agard et al., 2011; Verges et al., 2011). However, more recent tomographic inversions imply a continuous underthrusting of the Arabian plate with a low dip angle beneath the Central Iranian block in the SE Zagros (Chang et al., 2010; Simmons et al., 2011).
Figure 4.1: a) Surface topography of the Zagros fold-and-thrust belt. Surface shaded from resampled SRTM data to 250 m resolution (Jarvis et al., 2008). b) Main structural elements of the region inside the white rectangle in a). MZTF = Main Zagros Thrust Fault, HZF = High Zagros Fault, MFF = Mountain Frontal Fault, KF = Kazerun Fault, DEF = Dezful Embayment Fault, BRF = Balarud Fault, and ZFF = Zagros Foredeep Fault. Blue circles show salt diapirs from Bahroudi and Koyi (2003). Our models can be seen as a simplified representation of a cross-section along the red line (AB). c) Present-day topography along the red line (AB) in the SE Zagros from resampled SRTM data to 250 m resolution (Jarvis et al., 2008).
Figure 4.2: a) Sedimentary stratification representative for the southeast Zagros (Alavi, 2004). Frictional materials in grey and ductile (weak) materials in red. The most prominent weak layer is the Hormuz salt which separates basement from the overlying sediments. b) Cartoons illustrating continental collision between the Arabian plate and the Iranian micro-plate. Panel on top shown enlargement of the sedimentary folding on the incoming plate at 14±3 km wavelength (Yamato et al., 2011). MFF = Mountain Frontal Fault, HZF = High Zagros Fault, MZTF = Main Zagros Thrust Fault, SSZ = Sanandaj-Sirjan Zone.
Using 3D numerical models, van Hunen and Allen (2011) show that in a subduction-collision system slab break-off could start ca. 16 Myr after the onset of collision near the edge of a continental block and propagate along-strike. Their results can explain slab break-off under the NW Zagros and continuous slab underthrusting under the SE Zagros.

The NW Zagros and SE Zagros are separated by the north-south trending Kazerun fault (Hatzfeld et al., 2010). Across this fault the surface expression of salt-related deformation changes. Several salt layers were deposited on a highly metamorphosed Proterozoic basement in the SE Zagros (Alavi, 2004; Hatzfeld et al., 2010). The presence of these weak layers between other (mechanically stronger) sedimentary layers has been shown to have a significant impact on the style of deformation in the SE Zagros by introducing detachment layers and complex salt diapir structures (Colman-Sadd, 1978; Davies and Engelder, 1985; Talbot and Alavi, 1996; Bahroudi and Koyi, 2003; Alavi, 2004; Yamato et al., 2011). In the NW Zagros such salt diapirs are absent. The Neoproterozoic-Cambrian Hormuz salt with a thickness of 1-2 km is the oldest and most important weak layer in the Zagros FTB. It rests directly on the crystalline basement (Talbot and Alavi, 1996; Alavi, 2004). Previous studies pointed out that this layer behaves as a viscous basal decollement which changes the evolution of the orogenic belt, by forming a gentle taper, and facilitating progression of the deformation front away from the suture (Davies and Engelder, 1985; Koyi et al., 2000; Bahroudi and Koyi, 2003; Storti et al., 2007).

Using numerical models, Yamato et al. (2011) show that a single viscous decollement at the base of the sedimentary layers would result in thrust-dominated deformation. This finding does therefore not agree with the regular 14±3 km wavelength folding that is observed in the SE Zagros (Yamato et al., 2011). To produce such upper-crustal-scale folds, internal weak layers are required, combined with relatively weak uppermost crust and sediments (Yamato et al., 2011).

### 4.2.2 Our study of salt in the Zagros FTB

Many previous studies have investigated the role of salt in FTBs with numerical and analogue models at the scale of the upper crust (e.g. Cotten and Koyi, 2000; Bahroudi and Koyi, 2003; Clark et al., 2005; Emami et al., 2010; Yamato et al., 2011; Buiten, 2012; Graveleau et al., 2012). The advantage of numerical crustal-scale models is that a high resolution can be reached and that the models allow a direct control of initial geometries and boundary conditions. At a larger scale, numerical models have been used to investigate the effects of slab detachment in the Zagros (Yamato et al., 2007; van Hunen and Allen, 2011; Bottrill et al., 2012).
So far, no models have reconciled scales between the upper crust and the upper mantle. Here we aim to bridge between shallow (sandbox-like) studies of salt in fold-and-thrust belts and upper mantle studies of continental collision. We first use a continent-ocean-continent collision model to investigate the effects of salt layers on crustal deformation in a larger (upper-mantle-scale) framework. The driving mechanism for these models is a combination of a constant lateral push velocity and slab pull. These result in a convergence velocity after collision, which is of the same order of magnitude as the present-day convergence between the Arabian plate and the Iranian micro-plate obtained from GPS measurements. We then use higher resolution small-scale models to examine the deformation of a 200 km wide region of the sedimentary sequence overlying the Arabian plate during collision. We use these models to investigate the effects of salt rheology, basal dip angle, and the distribution of the salt layers on deformation styles. The driving velocity, dip of the top basement, and basal heat flow are constrained by values from the larger collision model.

Our models highlight the importance of choosing the appropriate scale for models. First-order aspects of the Zagros FTB development can be investigated with models on the scale of the upper-mantle, whereas studies that aim at investigating salt and sediment deformation can focus on the scale of the upper crust. Upper crustal models may benefit from kinematic and thermal constraints derived from large-scale models. We have chosen to use constraints from the large-scale model at 5.5 Myr after the onset of collision as the starting point for our upper crustal-scale models. But models may also couple between scales at each time step, as in the nested model approach of Beaumont et al. (2009). Our models confirm the finding of Yamato et al. (2011) that at least one internal salt layer is required in addition to the basal Hormuz salt to reproduce fold-dominated deformation. We use this result to explore the sensitivity of the models to the viscosity values of salt, top basement dip, and the lateral distribution of salt on fold spacing and surface topography in the SE Zagros.

4.3 Geological and geophysical data used in the modelling

Our models are as far as possible constrained by geological and geophysical observations of the sedimentary layers, crustal and lithospheric thicknesses, crustal thermal structure, and convergence velocity. Alavi (2004) compiled a sedimentary stratigraphy of the Zagros FTB based on field observations, sedimentological and petrographic information, and well data. This compilation shows 4-7 km of
Paleozoic and Mesozoic sedimentary successions overlain by 3-5 km of Cenozoic siliciclastic and carbonate rocks (Fig. 4.2a). These sediments rest on highly metamorphosed Proterozoic Pan-African basement (Alavi, 2004). In the SE Zagros, the sedimentary column also contains four intervening weak layers. These layers consist of halite, gypsum, anhydrite, dolomite, and shale (together referenced as ‘salt’ in this paper, although we recognise that shales are not evaporitic depositions as the others). The Hormuz salt with a thickness of 1-2 km is the oldest and most significant weak layer in SE Zagros.

The temperature gradient in the Zagros and surrounding areas is so far poorly constrained. The first heatflow study in Zagros showed that surface heatflow varies between 21-51 mW m$^{-2}$ with an average value of 36 mW m$^{-2}$ (Coster, 1947). Borehole data (two wells) from the National Iranian Oil Company suggest a crustal geotherm in the range of 10-23 °C km$^{-1}$ (Bird, 1976). These values are supported by recent studies of 3 wells which indicate a crustal geotherm of 15-24 °C km$^{-1}$ in the Zagros (Gavillot et al., 2010; Homke et al., 2010; Khadivi et al., 2012).

The crustal thickness map based on Bouguer gravity anomalies of Dehghani and Makris (1984) shows that the crust has a thickness of ca. 40 km beneath the Arabian plate and Central Iran, but increases to about 50-55 km under the MZTF (Fig. 4.2b). This increase is likely caused by thickening of the crust during collision. Receiver function analyses of a teleseismic earthquake (single station in central Iran, (Hatzfeld et al., 2003) and dense stations along two transects that cross the NW and SE Zagros (Paul et al., 2006, 2010)) show similar average crustal thicknesses, but a deeper Moho of ca. 69 km under the MZTF in the southeast.

The global thermal model of Artemieva (2006) shows a lithospheric thickness of 100 km for the Zagros region. However, only few borehole heatflow measurements were available to constrain the model in Iran. Using surface wave data and fitting temperature converted from shear wave speeds to a geotherm, Priestley et al. (2012) obtained a lithospheric thickness less than 120 km for the surrounding undeformed region increasing to as much as 225 km beneath the Zagros. This indicates that shortening is accommodated by lithosphere thickening in the Zagros. From S-receiver functions, Mohammadi et al. (2013) also find a north-eastward increase in lithospheric thickness towards the Zagros, from 130 km beneath the simply folded Zagros to 150 km beneath the Sanandaj-Sirjan zone (Fig. 4.1b).

GPS measurements show a present-day convergence rate between Arabia and Eurasia of 9-22 mm yr$^{-1}$. A regional-scale survey, with a station spacing larger than 150 km, found an average convergence rate of about 22 mm yr$^{-1}$ (Nilforoushan et al., 2003; Vernant et al., 2004). The higher resolution (on the order of 100 km) measurements of Hessami et al. (2006) in SE Zagros show a current shortening rate
of 9±3 mm yr⁻¹ accommodated along a zone of active thrusting and at the MZTF.

4.4 Numerical approach

4.4.1 Modelling method

We use the Arbitrary Lagrangian-Eulerian (ALE) finite element code SULEC-2D v3.4 to solve the equations for conservation of mass, momentum, and energy for incompressible slow creeping flows:

\[ \nabla \cdot \mathbf{u} = 0 \]  
\[ \nabla \cdot \sigma' - \nabla P + \rho \mathbf{g} = 0 \]  
\[ \rho c_p \left( \frac{\partial T}{\partial t} + \mathbf{u} \cdot \nabla T \right) = k \nabla^2 T + Q \]

\( \mathbf{u} \) is the velocity vector, \( \sigma' \) deviatoric stress tensor, \( P \) dynamic pressure (mean stress), \( \rho \) density, \( \mathbf{g} \) gravitational acceleration (\( g_x = 0 \), \( g_y = -9.8 \) m s⁻²), \( c_p \) specific heat, \( T \) temperature, \( t \) time, \( k \) thermal conductivity, and \( Q \) heat production. We solve for pressure using the iterative penalty formulation (Uzawa method) (Pelletier, 1989).

\[ P^t = P^{t-1} - \lambda \nabla \cdot \mathbf{u} \]

\( \lambda \) is penalty or compressibility factor (10⁻³⁰ Pa s) and \( t \) indicates time step. Each time step, we first compute viscosities and velocities until the velocity field is converged, and then compute pressure until \( \nabla \cdot \mathbf{u} \) reaches to a specified tolerance (order of 10⁻¹⁴). We discretize the model domain with quadrilateral elements with a continuous linear velocity field and constant, discontinuous pressures. The Eulerian grid is vertically stretched or shrunk to accommodate displacements of the free surface and surface processes (Fullsack, 1995). We include a stabilization term to suppress numerical overshoot across interfaces with large density contrasts, such as the top surface (Kaus et al., 2010; Quinquis et al., 2011). Density depends on temperature following the Boussinesq approach:

\[ \rho = \rho_0 (1 - \alpha (T - T_0)) \]

\( \rho_0 \) is the density at temperature \( T_0 \) and \( \alpha \) is the coefficient of volume expansion. Our materials deform by viscous or plastic behaviour. Viscosity \( \eta \) has a power-law
relation with strain-rate, pressure, and temperature:

$$ \eta = s_{c} \frac{1}{2} A^{-\frac{1}{n}} \dot{\varepsilon}_{II}^{\frac{1}{n} - 1} d^{\frac{p}{n}} C_{OH}^{-\frac{r}{n}} e^{(E + PV) \frac{RT}{n}} $$

(4.6)

$s_c$ is a scaling factor, $A$ a constant, $\dot{\varepsilon}_{II}$ effective strain-rate ($\dot{\varepsilon}_{II} = \frac{1}{2} \dot{\varepsilon}_{ij} \dot{\varepsilon}_{ij}$), $n$ stress exponent, $d$ grain size, $p$ grain size exponent, $C_{OH}$ water content, $r$ water content exponent, $E$ activation energy, $V$ activation volume, and $R$ the gas constant.

Sub-lithospheric material deforms following dislocation ($\text{dis}$) and diffusion ($\text{diff}$) creep in parallel and the composite ($\text{comp}$) viscosity is computed from (van den Berg et al., 1993):

$$ \eta_{\text{comp}} = \frac{\eta_{\text{dis}} \eta_{\text{diff}}}{\eta_{\text{dis}} + \eta_{\text{diff}}} $$

(4.7)

We simplify the rheology of salt in our models and consider salt as a weak layer, which we approximate with a linear viscous (Newtonian) rheology in most of our models. Only one model uses the creep law for halite (Heard, 1972). We use a minimum viscosity cut-off of $10^{17}$ Pa s in all our models. Material fails in a plastic manner when deviatoric stress exceeds a critical yield stress described by the Drucker-Prager relation:

$$ \sigma'_{II} = P \sin \phi + C \cos \phi $$

(4.8)

Where $\sigma'_{II}$ is the effective deviatoric stress ($\sigma'_{II} = \frac{1}{2} \sigma_{ij} \sigma'_{ij}$), $\phi$ angle of internal friction, $C$ cohesion, and $P$ dynamic pressure. We use a linear softening of the angle of internal friction between $\phi_1$ and $\phi_2$ (Table 4.1) over an interval of finite strain of 0.5-1.0 (e.g. Beaumont et al., 1996). This mimics the softening of shear zones because of foliation development, mineral transformations, and grain size changes (Mandl, 1988; Rice, 1992; Bos and Spiers, 2002). Material properties are stored on tracers and advected at the end of each time step. We use harmonic averaging of tracers to elements for viscosity, and arithmetic averaging for density. All material parameters used in our models are listed in Table 4.1. SULEC uses the direct solver PARDISO (Schenk and Gärtner, 2004) to solve the mechanical and thermal equations.

**4.4.2 Model set-up**

To explore the role of salt in continental collision, we start with models on the scale of the upper mantle and then examine smaller, upper-crustal-scale models. Our large-scale model is similar to the kinematically driven collision model of Ghazian and Buiter (2013). The model is 1900 km wide by 660 km deep and contains two continents separated by a 450 km wide ocean (Fig. 4.3a). The continents consist
Table 4.1: Rheological and thermal parameters of the models.

<table>
<thead>
<tr>
<th>Layers</th>
<th>Continental</th>
<th>Continental</th>
<th>Continental Sediment</th>
<th>Oceanic</th>
<th>Oceanic Mantle Upper seed</th>
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<td>Flow law</td>
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<th>Linear viscous</th>
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<td>dislocation dislocation dislocation dislocation diffusion dislocation</td>
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- \( \sigma_{ij}^{(0)} \) in Pa
- \( n \) 4 3 3.5 2.1 3.4 3.5 1 3.5 3.4 3.5 5.5 1
- \( E \) (J/mol) 137e3 345e3 520e3 210e3 497e3 480e3 335e3 480e3 497e3 480e3 98.3 0
- \( V \) (m^3 mol^-1) 0 38e-6 11e-6 0 0 11e-6 4e-6 11e-6 0 11e-6 0
- \( d \) (m) 0 0 0 0 0 0 5e-3 0 0 0 0
- \( p \) 00 0 0 0 0 3 0 0 0 0
- \( C_{OH} \) (H/106Si) 0 200 50 0 0 200 200 200 0 200 0
- \( r \) 0 1 1.2 0 0 1.2 1 1.2 0 1.2 0
- \( \varphi_1 \) 15° 15° 20° 10° 5° 20° 20° 5° 5° 0° 0°
- \( \varphi_2 \) 7.5° 7.5° 10° 5° 2° 10° 10° 2° 2° 0° 0°
- \( C \) (Pa) 20e6 20e6 20e6 20e6 5e6 20e6 0 2e6 2e6 1018
- \( \rho \) (kg m^-3) 2800 2900 3240 2700 3100 3280 3250 3100 3240 2200
- \( T \) (°C) 0 339 1300 0 0 1300 1300 0 1300 0
- \( \alpha \) (K^-1) 2e-5 2e-5 2e-5 2e-5 2e-5 2e-5 2e-5 2e-5 2e-5 2e-5
- \( k \) (W m^-1 K^-1) 2.5 2.5 2.8 2.5 2.3 2.3 83.33 2.3 2.3 2.5
- \( C_p \) (m^2 K s^-2) 750 750 750 750 750 750 750 750 750 750
- \( Q \) (W m^-3) 0.58e-6 0.58e-6 0 0 0 0 0 0 0 0

- \( a \) Gleason and Tullis (1995),
- \( b \) Rybacki et al. (2006),
- \( c \) Hirth and Kohlstedt (2003),
- \( d \) Turcotte and Schubert (2002),
- \( e \) Wilks and Carter (1990),
- \( f \) Heard (1972).

\( g \) Adjusted to a general state of stress (Ranalli, 1995).

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of an upper crust and lower crust of 20 km thickness each which overlie an 80 km thick continental lithospheric mantle. The intervening ocean consists of a 70 km thick oceanic lithospheric mantle overlain by a 6 km oceanic crust. The oceanic plate is about to subduct underneath the right continent. A weak inclined region (referred to as weak seed) of ca. 4-5 elements wide ensures that strain localises along the ocean-continent interface. The seed extends from the surface to the base of the continental lithosphere and has a dip angle of 38°. The Eulerian grid has a fine resolution defined by elements of 0.5 km height in the upper crust, coarsening vertically to 12 km towards the base of the model. The horizontal elemental size is also 0.5 km near the trench and increases to 12 km towards the sides. Our models have 57200 elements and 514800 tracers (9 tracers per element) initially. The number of tracers may change during model evolution by tracer injection and deletion which keeps tracers density between 9 and 20 for each element.

The left continental lithosphere is pushed with a constant inward velocity of 3 cm yr⁻¹, which together with slab pull ensures that the effective convergence rate across the trench region after collision is ca. 9±3 mm yr⁻¹, consistent with the present-day convergence in SE Zagros. The velocity boundary condition is applied from the surface down to the last four continental lithospheric elements over which the velocity reduces linearly to zero at the base of the lithosphere. The right continental lithosphere is held fixed at the side boundary over the lithosphere thickness. The lithospheric influx is balanced by an outward velocity at both sub-lithospheric side boundaries. The velocity increases linearly over the first four sub-lithospheric elements to a constant outward velocity (0.32 cm yr⁻¹). Our model has a vertically fixed basal boundary at 660 km depth and a true free surface boundary at the top. The vertical component of velocity at the side boundaries and horizontal component of velocity at the base are free.

The initial temperature of the continental lithosphere is computed from the steady-state conductive heat equation (Chapman, 1986) while the plate cooling model is used for the oceanic plate (Turcotte and Schubert, 2002) (Fig. 4.3b). We use a 60 Ma old ocean, which combined with a > 10 Myr subduction period before collision, results in a 70 Ma old ocean at the onset of collision. The Neo-Tethys Ocean was likely much older at collision though the exact age is difficult to determine now. Since oceanic lithosphere converges to a similar thermal structure for ages over 70 Ma and density is temperature-dependent in our models, an older slab than our model age of 70 Ma at collision, would result in a similar density structure and similar amount of slab pull. Our thermal parameters result in a 15 °C km⁻¹ continental crust geotherm which is within the range observed for the Zagros (Coster, 1947; Bird, 1976; Gavillot et al., 2010; Homke et al., 2010; Khadivi et al., 2012).
Figure 4.3: a) Set-up of the large-scale collision model showing two continents with an intervening 60 Ma ocean. The box on the left continent shows the schematic location of the upper-crustal-scale model. M = mantle, CUC = continental upper crust, CLC = continental lower crust, CL = continental lithosphere, OL = oceanic lithosphere, O = oceanic crust, WS = weak seed. b) Initial thermal profiles of large-scale model. O = ocean, C = continent. c) Set-up of the model on the scale of the upper crust. HZF = High Zagros Fault (modelled as a lithological contrast in our models, not as a pre-existing fault).
As our models do not have vigorous enough mantle convection to maintain the mantle adiabat, the sub-lithospheric mantle is assigned a high thermal conductivity ($k = 83.33 \text{ W m}^{-1} \text{ K}^{-1}$) which ensures an adiabat of $0.3 \, ^\circ\text{C km}^{-1}$, following the approach of Pysklywec and Beaumont (2004). During model evolution, the temperature is held fixed at $0 \, ^\circ\text{C}$ at the surface and free at other boundaries. Heat flux is held fixed at $25 \, \text{mW m}^{-2}$ at the base of the model, $0 \, \text{mW m}^{-2}$ at the side boundaries, and free at the top.

Our small-scale model is a 200 km wide slice of our large-scale model, consisting of the uppermost 7.48 km of sediments (Fig. 4.2b and 4.3c). The elements are 250 m wide and 34 m high over the entire domain. The model has 720000 tracers initially. Convergence is driven by a constant 9 mm yr$^{-1}$ from the left side and base of the model while the right side boundary is held fixed. This velocity follows the observed convergence between the continents after collision in our large-scale model and also agrees with present-day GPS measurements of $9 \pm 3 \, \text{mm yr}^{-1}$ east of the Kazerun fault in SE Zagros (Hessami et al., 2006; Hatzfeld et al., 2010). The vertical component of velocity on the side boundaries is free. Our small-scale model has a true free surface at the top. The steady-state heat equation is solved to achieve a $15 \, ^\circ\text{C km}^{-1}$ initial continental geotherm. The surface temperature is again held fixed at $0 \, ^\circ\text{C}$ and free at other boundaries during model evolution. Heatflow is set to $37 \, \text{mW m}^{-2}$ at the base (following values from the large-scale model) and $0 \, \text{mW m}^{-2}$ at the side boundaries.

4.5 The role of salt in the evolution of collisional upper-mantle-scale models

We present three upper-mantle-scale models (models M1-3), which differ from one another in the lithological layering of the upper crust and sediments of the incoming continent. The upper crust of the left (west) continent in the first model consists of 20 km of felsic crust with wet quartzite rheology, the second model has 8 km of sediments resting on 12 km of felsic crust with wet quartzite rheology, and the third model is composed of 6.5 km of sediments resting on 1.5 km of salt, both overlying 12 km of felsic crust with wet quartzite rheology. This salt layer represents the Hormuz salt in the Zagros (weak layer nr. 1 in Fig. 4.2a). We include only this salt layer (and not nrs. 2-4 of Fig. 4.2a) because the other three layers are too thin for the resolution of our upper-mantle-scale model. These models are aimed at investigating the effects of a sedimentary package and the presence of a weak layer at the base of the sediments on the style of deformation in a continent-continent collision
Figure 4.4: Three large-scale collision models. The figures show material (large panel) and viscosity (200 km wide slice of the incoming continent) after 20 Myr of convergence. Collision starts after 14.5 Myr of convergence (so the figures are at 5.5 Myr after the onset of collision). The models have an upper crust of 20 km thick of different composition: a) Model M1 with upper crust of wet quartzite rheology. b) Model M2 with 8 km of sediment on the incoming continent (and 12 km of upper crust). c) Model M3 with 6.5 km of sediment overlying 1.5 km of salt (and 12 km of upper crust). M = mantle, SE = sediment, O = oceanic crust, CL = continental lithosphere, OL = oceanic lithosphere, CUC = continental upper crust, CLC = continental lower crust, WS = weak seed, S = salt.
Table 4.2: List of models.

<table>
<thead>
<tr>
<th>Model</th>
<th>Upper-mantle scale</th>
<th>Upper-crustal scale</th>
<th>Sediment</th>
<th>Salt layers$^a$</th>
<th>Salt viscosity ($\mu$Pa s)</th>
<th>Basal dip</th>
<th>Salt distribution</th>
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<td>×</td>
<td>×</td>
<td>×</td>
<td>×</td>
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</tr>
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<td>0°</td>
<td>0°</td>
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</tr>
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</table>

$^a$ Salt layers as numbered in Fig. 4.2a.

Figure 4.4 shows the material and viscosity fields after 20 Myr of convergence. Because of the continuous push on the incoming continent and the strong slab pull due to the subducted oceanic slab, continental subduction is the dominant deformation style in the models (see also Ghazian and Buiter, 2013). In the models without salt (Figs. 4.4a, 4.4b), the lower continental crust and lithospheric mantle subduct together, but the upper crust decouples from the lower crust and accumulates at the suture zone. The viscosity plot clearly shows this decoupling with a low viscosity band formed at the base of the upper continental crust. Stress is distributed almost evenly within the upper crust without localisation into shear bands (Figs. 4.4a, 4.4b). A weak layer at the base of the sediments changes the deformation pattern (Fig. 4.4c). The weak layer causes a decoupling of the overlying sediments from the basement. The remaining part of the upper continental crust has therefore less negative buoyancy and can subduct more easily. The weak layer also promotes localisation of deformation in the sediments with trench-verging shear bands. The shear bands first form near the suture zone and then propagate progressively westward. This model shows the impact that a weak layer can have on continental collision. It also illustrates that most deformational effects are concentrated in the crust, allowing us to achieve a higher numerical resolution in the area of interest by limiting the size of our model domain (see next section).
4.6 The role of salt in the evolution of collisional upper-crustal-scale models

Upper-mantle-scale models usually do not resolve thin intervening weak layers (of a few hundreds of meters thickness) in sediments or crust very well, because it is computationally expensive to achieve the required resolution at this scale. We therefore use the upper-mantle-scale models to show the larger-scale effects of the thick Hormuz salt and to extract boundary conditions for our upper-crustal-scale models. The driving velocity ($9 \text{ mm yr}^{-1}$) and basal heatflow ($37 \text{ mW m}^{-2}$) are taken from model M3 at 20 Myr. Our first models (M4-M13) have a horizontal base, but large-scale model M3 indicates a top basement dip $+1^\circ$ and we therefore separately investigate the effects of basal dip angle on the evolution of the upper crustal-scale models.

Our upper-mantle-scale models show that sediments resting on a ductile layer decouple from the underlying crust and lithosphere and accumulate near the trench (Fig. 4.4c). A few million years after onset of collision, the sediments on the incoming continent are locked near the suture zone and pushed by a constant boundary condition at their left (westward) side. This is the setting we consider for our upper-crustal-scale models (Fig. 4.3c). We view the area of our model to represent the sediments between HZF and MFF in the SE Zagros (red line in Fig. 4.1b).

The aim of these models is to investigate the effects of basal and intervening weak layers, salt strength, basal dip, and lateral salt distribution on deformation style. In models with a basal salt layer, the initial surface of the basal layer is perturbed with a sine function (with a wavelength of 14.4 km and amplitude of 200 m) to initiate folding. This ensures that folds initiate with the wavelength that agrees with the observed dominant wavelength in the SE Zagros (Molinaro et al., 2005a,b; Mouthereau et al., 2007; Yamato et al., 2011). Although we explicitly introduce the initial wavelength, it is important to note that folds finally develop with a dominant wavelength that depends on the viscosity contrast between weak and frictional layers and their thicknesses.

Figure 4.5 shows the evolution of a sedimentary layer with a basal and three intervening weak layers (weak layers 1-4 in Fig. 4.2a), which is shortened at its left-hand side (model M4). Folding starts symmetrically close to the HZF (Fig. 4.5a) and propagates left(west)-wards as shortening proceeds. Previously formed folds become asymmetric, verging westwards (Fig. 4.5a-f). Shear bands in the top frictional layer initially form as regular conjugate sets at 14.4 km spacing, measured along the surface as the distance between shear bands of the same vergence. Later in model
evolution, folding becomes the dominant deformation mechanism and controls the spacing of the shear bands (Fig. 4.5a-f). The wavelength of the folds in model M4, is initially 14.4 km and becomes 11.5 km after 5.5 Myr of convergence, following effective shortening with 25% of the original model width (49.5 km convergence over 200 km model width). These wavelengths are in agreement with the observation of 11-17 km wavelengths in the SE Zagros and with numerical results of Yamato et al. (2011).

4.6.1 The effects of basal and intervening salt layers

Weak layers facilitate outward propagation of deformation, formation of fold-controlled shear bands in the overlying frictional layer, and low surface slope (Fig. 4.5, 4.6a-d). We separate the effects of the three intervening weak layers from the basal layer, by progressively removing the weak layers from the model (models M4-8 in Fig. 4.6). A purely frictional model, without weak layers, deforms in a localised manner and builds a wedge-shaped topography close to the right-hand (east) side (HZF), while the region far from HZF is almost undeformed (Fig. 4.6e). The wedge achieves a surface slope of 13.3° after 5.5 Myr of convergence (measured as a line through the thrust valleys). For our material properties and model setup (Table 4.1, Fig. 4.3b), we compute an analytical upper critical taper angle of 12.7° (Zhao et al., 1986). Our model is therefore close to the expected analytical value. A model with only a basal weak layer does not experience folding, but deforms by thrusting (Fig. 4.6c). Intervening weak layers are required for the model to deform by folding and shear zone formation, which confirms the results of Yamato et al. (2011) (Fig. 4.6a-c). As Yamato et al. (2011), we also find that the transition from thrusting to folding occurs when one internal weak layer is presented in addition to the basal weak layer.

4.6.2 The effect of salt strength

We investigate the effects of viscous material strength for the model with a basal salt and three intervening salt layers (Fig. 4.7). We use linear viscous materials with a viscosity of $10^{17}$, $10^{18}$, $5 \times 10^{18}$, $10^{19}$, and $10^{20}$ Pa s and a power-law material with a halite rheology (models M4 and M9-13). Figure 4.7 shows that models with a weak layer viscosity in the range of $10^{17}$-$10^{19}$ Pa s, deform in a fold-dominant mechanism (Fig. 4.7a-d), whereas thrusting is the dominant deformation mechanism for higher viscosities (Fig. 4.7f). Although folding occurs over a wide range of viscosities (two orders of magnitude), the wavelength and amplitude of the folds change with the strength of the weak layers.
Figure 4.5: Strain-rate plots of upper-crustal-scale model M4 with four weak layers (layers 1-4 in Fig. 4.2a) at: a) 1 Myr, b) 2 Myr, c) 3 Myr, d) 4 Myr, e) 5 Myr, and f) 5.5 Myr of convergence. Black line represents the High Zagros Fault, which is included as a lithological transition to a domain without salt in our models.
Folds have higher amplitude and are irregularly distributed in the models with lower viscosity values, but are evenly distributed (with a wavelength of 11.5 km) with lower amplitude for higher viscosity values (see Fig. 4.7a-d). This observation can be explained by the interplay between gravitational and compressional forces, where for lower viscosities (higher Argand numbers) gravity dominates over compression and gravitational growth of the weak layers is the dominant controlling mechanism, while for higher viscosities (lower Argand numbers) folding is the dominant structure forming mechanism (Burg et al., 2004). We find that for our set-up, the effective viscosity of weak material with a halite rheology is similar to a linear viscous material of $5 \times 10^{18}$ Pa s, and that these two models show almost identical deformation patterns and viscosity structures (Fig. 4.7c,f). This series of models indicates that viscosity values between $10^{17}$ and $10^{19}$ Pa s are low enough (compared with our sediments with $\varphi=5^\circ$ and $C=20$ MPa) for the salt layers to act as weak layers and lead to folding of the sediments, rather than thrusting.

4.6.3 The effect of basal dip

Our upper-crustal-scale models so far had a horizontal base, but the large-scale models have a dip angle of +1° at the base of the sediment layers (dipping downward towards the trench) (Fig. 4.4c) after 5.5 Myr after onset of collision. Changes in basal slope may affect the surface slope and the deformation style in a fold-and-thrust belt (Davis et al., 1983; Dahlen et al., 1984; Smit et al., 2003).
Figure 4.7: Strain-rate plots of upper-crustal-scale models with a basal weak layer and three intervening weak layers (layers 1-4 in Fig. 4.2a) after 5.5 Myr of convergence. a) Viscosity of the weak layers is $10^{17}$ Pa s (model M9). b) Viscosity of the weak layers is $10^{18}$ Pa s (model M10). c) Viscosity of the weak layers is $5 \times 10^{18}$ Pa s (model M4). d) Viscosity of the weak layers is $10^{19}$ Pa s (model M11). e) Viscosity of the weak layers is $10^{20}$ Pa s (model M12). f) Viscosity of the weak layers follows halite flow-law of Heard (1972) (model M13).
Crustal models from gravity observations, seismological data, and balanced cross-sections also indicate that the base of the sedimentary layer in the SE Zagros dips with +0.5° to 1° (Snyder and Barazangi, 1986; Hatzfeld et al., 2003; McQuarrie, 2004). We investigate the effect of variations in basal dip angle on our upper-crustal-scale models by tilting the gravitational force between -3° to 3° for the model with a basal and three intervening weak layers (layers 1-4 in Fig. 4.2a) (models M4, M14-19). Figure 4.8 shows that basal dip changes lead to different folding and thrusting deformation styles. In the model with a flat base (M4, Fig. 4.8d) folding is distributed homogeneously over the entire model. Models with a negative basal dip (dipping away from the HZF) show reduced surface topography and more salt extrusion structures near the HZF, and higher topography and folding-dominated deformation away from it (M14-16, Fig. 4.8a-c). This is not consistent with observations in the SE Zagros, where most of the extruded salt diapirs are located north-east of the MFF (Fig. 4.1b) (which would be in the left part of our models) and topography decreases away from the HZF. Models with a positive basal dip (dipping towards the HZF) give more consistent results with observations in SE Zagros. 

Figure 4.8: Material plots of upper-crustal-scale models with different basal dips after 5.5 Myr of convergence. All models have a basal weak layer and three intervening weak layers (layers 1-4 in Fig. 4.2a with a viscosity of $5 \times 10^{18}$ Pa s). a) Model with $-3^\circ$ basal dip (M14). b) Model with $-1^\circ$ basal dip (M15). c) Model with $-1^\circ$ basal dip (M16). c) Model with $0^\circ$ basal dip (M4). d) Model with $1^\circ$ basal dip (M17). e) Model with $2^\circ$ basal dip (M18). f) Model with $3^\circ$ basal dip (M19).
Zagros (M17-19, Fig. 4.8e-g). They show highest topography and fold-dominated deformation close to the HZF. Surface elevation decreases away from the HZF. Folding is the main deformation style until far from the HZF where the frictional layer in the shallow part of the model thins to the degree that the intervening weak materials can open a channel to the surface and extruded salt diapirs are more frequent. We find therefore that a base that dips with 1° towards the HZF best explains surface slope and salt extrusion patterns observed in the SE Zagros.

4.6.4 The effect of the spatial distribution of salt layers

The distribution of salt varies both temporally and spatially in the Zagros FTB. The spatial variation has been shown to play a significant role in the style of deformation and strain partitioning in the FTB (Bahroudi and Koyi, 2003), as domains without salt are thrust-dominated and form a steep taper. To investigate the effects of the spatial distribution of salt on our models, we use four models with different initial salt distributions (models M4, M20-21, M7) (Fig. 4.9). The strain-rate fields of the models confirm that deformation is fold-dominated in the region where salt is present, while regions without salt experience thrust-fault-dominated deformation (Fig. 4.9). Models with a partial salt distribution (models M20-21) show a prominent thrust fault, which separates the highly deformed area where salt is present from a relatively undeformed area where salt is absent. The formation of this type of fault during the early evolution of the Zagros FTB has formerly been attributed to reactivation of old normal faults in the basement (Jackson, 1980). Our results agree with the analogue results of Bahroudi and Koyi (2003) in suggesting that these structures can also be formed by initial variation of the mechanical properties.

4.7 Discussion

We used upper-mantle-scale and upper-crustal-scale models to investigate the role of salt in the Zagros collision. We find that the presence of the basal and intervening weak layers as well as their strength and spatial distribution play a crucial role in controlling the style of deformation in the SE Zagros. We show that the dip of the top basement affects the style and distribution of deformation and has a significant impact on the predicted surface topography in the SE Zagros.

Figure 4.10 shows how a simple tuning of the sedimentary stratification improves the fit of predicted topography to observed topography.
Figure 4.9: Strain-rate plots of upper-crustal-scale models with different initial salt distributions after 5.5 Myr of convergence. All models have a basal weak layer (layer 1 in Fig. 4.2a). a) Intervening weak layers are absent (M7). b) Internal salt layers initially cover half of the model width (M20). c) Internal salt layers initially cover two thirds of the model width (M21). d) Internal salt layers cover entire model width (M4).
Upper-mantle-scale models without Hormuz salt (M1-2) predict a steep surface slope, while the model with Hormuz salt (M3) predicts a topography that is more consistent with the observation despite its low resolution in the crust (Fig. 4.10a). This clearly shows the importance of constraining numerical models by available observational data. We used geological and geophysical data to better constrain the crustal and lithospheric thickness, sedimentary stratification, convergence velocity, and thermal structure of our models. Our best large-scale model (M3) indicates that Hormuz salt decouples overlying sediments from the basement, and localises deformation in the sediments by trench-verging shear bands. The model also shows that the sediments are overlying a basement which dips with 1° toward the trench, which is in a good agreement with the predicted basal dip from gravity observations, seismological studies, and balanced cross-sections (Snyder and Barazangi, 1986; Hatzfeld et al., 2003; McQuarrie, 2004). Our upper-mantle-scale models also show that the Hormuz salt layer at the base of the model successfully reproduces thrust fault dominated deformation between MZTF and HZF in the SE Zagros. However, the presence of the basal weak layer alone is not sufficient to reproduce fold-dominated deformation between HZF and MFF in the SE Zagros.

Since our large-scale model with Hormuz salt (model M3) indicates that deformation is concentrated in the shallow part of the upper crust, we used upper-crustal-scale models to investigate this region with higher resolution. We used the velocity fields, thermal structure, and basal dip at 5.5 Myr after the onset of continental collision of our large-scale model M3 to constrain the initial set-up of our small-scale models.
The small-scale models illustrate how, following Yamato et al. (2011), internal weak layers are required to reproduce fold-dominated deformation between HZF and MFF in the SE Zagros. We illustrate that folding at regular intervals only occurs for viscosities of the weak layers between $5 \times 10^{18} - 10^{19}$ Pa s. As the weak layers occur at shallow depths in the SE Zagros where temperature is low, a linear approximation of their rheology is sufficient for these upper-crustal-scale models (Fig. 4.7). We find that basal dip is an important parameter to explain the observed high topography and fold-dominated deformation near the HZF and lower topography and extruded salt diapirs near the MFF in the SE Zagros. Figure 4.10b shows that a $+1^\circ$ basal dip gives the best fit to the observed topography. This result agrees with crustal models constrained by gravity and seismological observations and balanced cross-sections (Snyder and Barazangi, 1986; Hatzfeld et al., 2003; McQuarrie, 2004) and with the top basement dip we obtained from our large-scale model M3.

Our models suggest that the deformation style in the SE Zagros collision zone is best explained by a 7-8 km thick sedimentary sequence with the following properties: a) a thick (ca. 1.5 km) basal layer of weak salt, b) three intervening weak layers, c) a viscosity of $5 \times 10^{18} - 10^{19}$ Pa s for weak layers that are embedded in sediments with $\phi=5^\circ$ and $C=20$ MPa, d) the presence of weak layers in the sedimentary sequence over the entire region between the HZF and MFF, and e) a basement that dips $+1^\circ$ towards the High Zagros Fault. Most of these constraints are derived from upper-crustal-scale models as these have high enough resolution to resolve the internal weak layers of 100-1000 m thickness. However, our upper-mantle-scale models gave us the kinematic and thermal background required for the smaller-scale models. The large-scale models also illustrate the potential important role of weak layers along which sedimentary sequences can decouple and accrete near the trench, allowing the remaining continental crust to subduct. An interesting avenue to explore for future studies would be high-resolution 3D models of continental collision at the upper-mantle scale with several internal weak layers and lateral lithological variability on the incoming continent.

4.8 Conclusions

We investigated the role of salt in the SE Zagros fold-and-thrust belt with 2-D thermo-mechanical models. We first examined upper-mantle-scale models of ocean-continent convergence followed by continent-continent collision in order to obtain the first-order evolution of the region. These models were then used to constrain higher resolution, upper-crustal-scale models. We find that:
[1] The Hormuz salt plays an important role in controlling the deformation style of the SE Zagros FTB, by decoupling overlying sediments from the basement and localising deformation in the sedimentary layers.

[2] In addition to Hormuz salt, at least one internal weak layer in the sedimentary sequence is required to reproduce fold-dominated deformation in the simply folded Zagros (the region between HZF and MFF in Fig. 4.1b), confirming the findings of Yamato et al. (2011).

[3] Since salt in the SE Zagros occurs at shallow depths where temperature is low, linear viscous weak layers with a viscosity of $5 \times 10^{18}$ Pa s reproduce the behaviour of models with a halite creep law.

[4] Models with low viscosity weak layers ($5 \times 10^{18}$-$10^{19}$ Pa) deform by folding at wavelengths starting ca. 14.4 km initially and decreasing to ca. 11.5 km after 5.5 Myr of convergence. Models with higher viscosity values for the weak layers deforms in a thrust-dominated style.

[5] The dip angle of the top basement controls surface topography, the transition from thrust fault-dominated deformation near the trench to fold-dominated deformation away from the trench, and the distribution of extruded salt diapirs.

[6] Discontinuous salt layers may explain the formation of thrust faults that separate deformed regions from undeformed areas.

[7] Large (upper-mantle) scale models can be useful to constrain the initial kinematic and thermal conditions of crustal-scale models, whereas such smaller-scale models are best suited to lithological sensitivity studies.

We propose that deformation in the SE Zagros FTB was controlled by the Hormuz salt and three intervening weak layers with a viscosity of ca. $5 \times 10^{18}$ Pa s. The sediments and salt overlie basement that dips at $+1^\circ$ towards the High Zagros Fault. As long as crustal resolution resolves the intervening weak layers inside the sediments, upper-mantle-scale models can provide a reasonable picture of the Zagros FTB development albeit of long computation times. We show that studies that focus on sediment and salt deformation can safely be constructed on the scale of the upper crust only, without necessity for including the lithosphere and mantle. However we suggest that (lower resolution) upper-mantle-scale models prior to crustal-scale studies may help constrain the initial geometry and (kinematic and thermal) boundary conditions of the smaller scale models.
4.9 Acknowledgments

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