Dynamic evolution of thin-skinned fold-and-thrust belts: Field study, magnetostratigraphy and numerical modelling applied to the Zagros and Makran Mountains (Iran)

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“Here the ways of men divide. If you want to achieve peace of mind and happiness, then have faith; if you want to be a disciple of truth, then search.”

Friedrich Nietzsche
For my parents.
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Abstract

Thin-skinned fold-and-thrust belts related to convergent tectonics develop by scraping off a rock sequence along a weaker basal décollement, which is often formed by water-saturated shale layers or low-viscosity salt horizons. Although thin-skinned fold-and-thrust belts are among the best and most investigated geological features on earth, there are many items that need further examination. The interplay between surface processes and the dynamic structural evolution is tested by magnetostratigraphy of progressive unconformities in the onshore Makran accretionary wedge and the Zagros foreland fold-and-thrust belt. The impact of décollement and wedge rheology on the long-term evolution of thin-skinned fold-and-thrust belts is investigated with numerical modelling and applied to the Makran and the Zagros. Comparison of numerical results to analytical solutions has been conducted.

Progressive unconformities are predestined to link sedimentary and deformation processes in tectonically active settings. In the onshore Makran accretionary wedge, growth strata are found in large-wavelength, small-amplitude concentric synclines. The morphology of the growth strata indicates an underfilled basin where fold amplification was faster than sedimentation, forming typical sedimentary offlap structures. Magnetostratigraphy did not reveal any characteristic remanent magnetization. > 50% of the natural remanent magnetization was removed by step-wise thermal heating already above 200°C.

The temporal evolution of deformation in the Zagros Simply Folded Belt is constrained by a magnetostratigraphic sequence containing a progressive unconformity in the Central Fars. The correlation of the magnetostratigraphic section with the geomagnetic polarity time scale constrains the transition from marine to fluvial sediment deposition at ∼ 6 Ma. This transition was accompanied by a change in accumulation rate from ∼ 15 cm/ka to ∼ 40 cm/ka. Alluvial river deposits first occurred at 3.2 – 3.1 Ma. In the Central Fars folding was initiated at 3.7 – 3.5 Ma. Comparing magnetostratigraphic sections
and ages of growth strata in Central and NE Fars indicates a $\sim 1$ cm/a migration rate of the deformation front towards the SW during the Middle and Late Miocene.

A two-dimensional finite element model with a visco-elasto-plastic rheology is used to investigate the structural evolution of fold-and-thrust belts overlying different types of décollements and the influence of multiple weak layers within the stratigraphic column. Model shale décollements are frictional, with lower friction angles as the cover sequence. Model salt layers behave linear viscous, due to a lower viscosity as the cover sequence, or with a power law rheology. Fold-and-thrust belts with a single frictional basal décollement generate thrust systems ramping from the décollement to the surface. The structural evolution of simulations with an additional low-frictional layer depends on the strength relationship between the basal and the within-sequence décollement. Tectonic underplating and antiformal stacking, as observed in the Makran, occur if the within-sequence décollement is weaker than the basal one. Salt décollement with a viscosity of $10^{18}$ Pa·s leads to isolated box folds (detachment folds). Multiple salt layers ($10^{18}$ Pa·s) result in long-wavelength folding.

Latest developments in computer power and computational solutions open new ways to envision complex natural systems. The mechanics and dynamics of thin-skinned compressible thrust wedges with prescribed offsets in the backstop, i.e., transfer zones, are investigated using a three-dimensional finite difference numerical model with a visco-brittle/plastic rheology. The shorter the backstop offset, the earlier the two thrust planes connect, forming a curved frontal thrust along the entire width of the model. Younger, in-sequence thrusts are formed parallel to this curved shape. Long backstop offsets produce strongly curved thrust faults around the indenting corner. These simulations revealed that surface tapers of the wedge in front of the backstop promontory are larger than what the critical wedge theory predicts, whereas the tapers on the other side of the transfer zone are smaller than analytical values. This difference is amplified with increasing length of the backstop offset and/or strength of the décollement. Modelled surface elevation schemes reproduce well the topographic patterns of natural orogenic systems such as the topographic low along the Minab-Zendan transform/transfer fault between the Zagros and Makran.

Advances in high-resolution 3D geodynamic modelling also allow investigating complex tectonic processes such as transpression. We present new answers to long lasting discussions. The Zagros Simply Folded Belt in Iran exhibits spatial variations of structural patterns from the homogeneously shortened, wide fold belt in the Fars to the narrower,
higher elevation area in the Izeh. High-resolution numerical modelling shows how a low-viscosity décollement becoming frictional towards the Dezful embayment influences the exposed fold patterns. Results also emphasize the importance of an oblique backstop to produce en-échelon folds where convergence-related and backstop-controlled folds are mingled in the transpressional orogen.

Furthermore, the effect of frictional weakening of the wedge internal strength is numerically tested. Shear zones are weaker than surrounding rocks, but whether weakening depends on accumulated strain or strain-rate is largely debated. Therefore, the influence of strain and strain-rate (velocity) weakening is investigated. A uniform setup recognizes the effects of each weakening mode on the structural evolution and dynamics of fold-and-thrust belts. A dual setup with two adjacent domains of décollement strength investigates the structural response of laterally varying systems. Strain weakening and velocity weakening lead to remarkably different structural patterns. In contrast to strain weakening, velocity weakening starts with the onset of high strain-rate bands. The influences of shortening velocity, cover sequence thickness and weakening style are tested. The shortening velocity has a minor effect in strain-weakened models. Velocity weakening and a thinner cover sequence enhance the occurrence of strike-slip shear zones. The weakening mode strongly influences the fault patterns and the dynamics of thrust wedges.

Results are consistent with the critical wedge theory and additionally show that surface slope, internal and basal strength and stress orientations within a thin-skinned fold-and-thrust belt can be predicted.
Zusammenfassung


Bessere Resultate ergab die magnetisch-stratigraphische Untersuchung einer progressiven Diskordanz im Zagros, genauer im zentralen Fars. Die Korrelation der magnetischen Polaritäten entlang des untersuchten Abschnitts mit der geomagnetischen Polarität-
szeitkurve belegt einen Wechsel von einem maritimen zu einem flussartigen Ablagerungs-
system vor etwa 6 Ma. Dieser Wechsel ging einher mit einer Änderung der Akkumulati-
sionsrate von Sedimenten von \( \sim 15 \text{ cm/ka} \) zu \( \sim 40 \text{ cm/ka} \). Im zentralen Fars entstanden vor 3.2–3.1 Ma erstmals Schwemmebenen und die Faltung von Gesteinsschichten begann vor 3.7 – 3.5 Ma. Ein zeitlicher Vergleich von progressiven Diskordanzen im zentralen und nordöstlichen Fars deutet auf eine Migrationsrate der Deformationsfront von \( \sim 1 \text{ cm/a} \) in Richtung Südwesten hin, während dem mittleren und oberen Miozän.

Ein zweidimensionales Finites-Elemente-Modell mit einer viskos-elastisch-spröden Rhe-
ologie wurde angewendet, um die strukturelle Entwicklung von Falten- und Überschie-
ungsgürteln mit rheologisch unterschiedlichen Sohlflächen zu untersuchen. Auch der Ein-
fuss von relativ schwachen Horizonten, eingelagert entlang der deformierenden Strati-
graphie, wurde erforscht. In den Modellen haben die Sohlflächen aus Schieferton eine spröde Rheologie, wobei der interne Reibungswinkel kleiner ist als in den überlagern-
den Schichten. Modellierte Salzlagen sind entweder linear oder exponentiell viskos. Falten-
und Überschiebungsgürtel mit einer spröden Sohlfläche entwickeln Rampensysteme, welche von der Sohlfläche bis an die Oberfläche reichen. Zusätzliche spröde Abscherungshorizonte können zu Schuppenbildung in den untersten Schichten führen. Sohlflächen in Form von Salzlagen mit einer Viskosität von \( 10^{18} \text{ Pa} \cdot \text{s} \) führen zur Bildung von isolierten Kofferfalten. Im Falle von mehreren viskosen Salzhorizonten in der Stratigraphie entstehen vorwiegend Falten mit einer grossen Wellenlänge.

Die neuesten Entwicklungen bezüglich Komputertechnik ermöglichen neue Wege, um komplexe natürliche Systeme zu untersuchen und zu verstehen. Die Mechanik und Dy-
namik von deckschichttektonischen, kompressiven Falten- und Überschiebungsgürteln entlang von Transformzonen wurde mit Hilfe eines dreidimensionalen numerischen Models mit viskos-spröder Rheologie erforscht. Die Transformzonen sind definiert durch einen Versatz im Vorgebirge, in welchem keine Deformation stattfindet. Falls dieser Ver-
satz relativ klein ist, entstehen gekrümmte Rampen entlang des gesamten Modells. Bei grösseren Versätzen sind die Rampen um den Winkel des Vorgebirges gefaltet. Die mod-

Weitere Fortschritte hin zu hochauflösenden dreidimensionalen geodynamischen Mod-
ellen erlaubt auch das erforschen von so komplizierten Prozessen wie Transpression.
Im Zagros ändert sich das strukturelle Bild von einem weitreichenden gleichmäßig verkürzenden Faltengürtel im Fars zu einem eingeengten, erhöhten Abschnitt in der Izehzzone. Hochauflösende numerische Modelle zeigen, wie ein Abscherhorizont mit einer tiefen Viskosität, übergehend in spröde Schiefertone in Richtung der Dezful Einbuchung, die Verfaltung beeinflussen. Die resultierenden Modelle unterstreichen auch die Wichtigkeit eines schrägen (aus der Vogelperspektive) Vorgebirges für die Entstehung von En-échelon-falten.


Die erzielten Resultate sind im Einverständnis mit der kritischen Keiltheorie (critical wedge theory) und zeigen zusätzlich, dass Oberflächenneigungen, interne Festigkeit und Festigkeit der Sohlfäche sowie die Orientierung von Hauptspannungsachsen in Falten- und Überschiebungsgürteln ermittelt werden können.
1 | Introduction

1.1 Thin-skinned fold-and-thrust belts

1.1.1 General background

Fold-and-thrust belts form by deformation of predominantly sedimentary sequences in the foreland of developing orogens or along passive or active continental margins. Foreland fold-and-thrust belts are restricted to convergent plate motion. Most referred examples are the Jura Mountains (e.g., Laubscher, 1992; Sommaruga, 1997), the Appalachians (e.g., Hatcher, 1972), the Subandean Ranges (e.g., Baby et al., 1992; Echavarria et al., 2003), the Zagros (e.g., Blanc et al., 2003; McQuarrie, 2004) and others (Poblet and Lisle, 2011). Fold-and-thrust belts are classified into either near-field stress-driven or far-field stress-driven systems (Morley et al., 2011). Near-field stress system is used, when the potential energy arises entirely from uplift or sediment loading, leading to gravity-driven deformation (Rowan et al., 2004). Most examples occur along passive continental margins, such as the West African margins of Gabon, Congo and Angola (e.g., Broucke et al., 2004), the Niger Delta (e.g., Rouby et al., 2011), the Orange Basin (e.g., de Vera et al., 2010) and the Gulf of Mexico (e.g., Rangin et al., 2008). Far-field stress systems define fold-and-thrust belts driven by lithospheric stresses. Typical examples are accretionary wedges like the Makran (e.g., Platt et al., 1985), the South Caspian Sea (e.g., Kazmin et al., 2004), the Cascadia margin (e.g., Gulick et al., 1998) or the Barbados Ridge (e.g., Moore et al., 1998; Westbrook et al., 1988).

There are three modes of horizontal shortening in convergent settings (Coward, 1983; Pfiffner, 2006): i) Thin-skinned deformation is typical in the uppermost levels of the crust; associated thrusts and folds are uncoupled from lower crustal rocks by a basal décollement with a substantial amount of displacement. This style of deformation is common within accretionary wedges and gravity-driven fold-and-thrust belts (Gwinn,
1964; Hsu, 1979, 1981; Morley et al., 2011). ii) Deformation in foreland fold-and-thrust belts can also incorporate underlying basement rocks. If thrust faults cut into the crystalline basement levelling off the basement/cover interface, deformation is defined as basement-involved thin-skinned (Kley, 1996; Miller, 2003). iii) Thick-skinned tectonics implies that faults affect the whole upper crust, i.e. it deforms as one undifferentiated block, and possibly the lower crust (Coward, 1983).

Fold-and-thrust belts related to convergent plate boundaries host ~ 14% of the worldwide discovered hydrocarbon reserves, the Zagros accounting for about half of it (Cooper, 2007). Folding and thrusting within these belts are some of the most observed and best understood deformational features of the earth (Davis et al., 1983; McClay, 1995; McClay and Price, 1981). The following focuses on the widely distributed thin-skinned fold-and-thrust belts driven by lithospheric stress fields (Figure 1.1).

Chapple (1978) specified several fundamental characteristics of such orogenic mountain belts: i) the limiting horizon of folding and thrusting is commonly near or close to the crystalline basement (thin-skinned), ii) the basal décollement is generally composed of rocks relatively weaker than the overlying, folded and thrusted sedimentary sequence, iii) the wedge shaped fold-and-thrust belt is thinner at its toe and thicker towards its
Figure 1.2: Sketch profile of an accretionary wedge in a subduction system. The incoming pile of sediments is scraped off the subducting plate. a) Location of accretionary wedge, fore-arc basin and magmatic arc in a subduction setting (from Hamilton, 1995). b) Deformation takes place between the non-deforming incoming sequence and the non-deforming backstop. LSL: Limiting slip lines. SL: Sea level. (from Stockmal, 1983).

rear and iv) the wedge body has been strongly thickened and shortened due to thrusting and/or shortening strain.

Figure 1.2a illustrates a typical profile through a subduction zone, where oceanic lithosphere is overridden by a continental plate. The in-coming sedimentary sequence lying on the oceanic plate is scraped off by a rigid backstop. The fore-arc basin develops in response to and behind the uplift of the fore-arc ridge. Subduction of water-rich rocks leads to partial melting in the mantle wedge, above the down-going plate; resulting magmas feed the developing magmatic arc on the hanging wall lithosphere (Figure 1.2b).
Deformation within an accretionary wedge takes place between the fore-arc basin and the deformation front, where undeformed material on the subducting plate comes in (Figure 1.2b). In thin-skinned compressional settings, thrust zones are represented by ramp/flat systems, which cause ramp-anticlines and synclines (Figure 1.3a). Growth of accretionary prisms can result from frontal accretion, where incoming sediment is accreted at the toe of the wedge in an imbricate fan (Hoth et al., 2007), or underthrusting and consequential tectonic underplating, where in-coming sediment is accreted on the underside of the accretionary wedge, which leads to antiformal stacking and rearward thickening (Figure 1.3b). These two accretion modes do not exclude each other in natural systems (Figure 1.3c).

The majority of thin-skinned fold-and-thrust belts is comprised of shale and sandstone layers, often deposited into a turbiditic basin or as river and deltaic sediments on former
1.1. THIN-SKINNED FOLD-AND-THRUST BELTS

Figure 1.4: Laboratory measurements of fracture strength and maximum friction of dry sandstones and shales (from Dahlen et al., 1984; after Byerlee, 1978; Hoshino et al., 1972).

Figure 1.5: Wavelength vs. thickness of deformed sequence of referred thin-skinned fold-and-thrust belts characterized by a shale décollement (from Morley et al., 2011).

passive margins. Shale and sandstones have a brittle rheology in near surface temperature and pressure conditions, in particular if they are saturated in fluids (Byerlee, 1978). The rheological behaviour of thin-skinned fold-and-thrust belts can therefore be inferred from the brittle strength of shale and sandstone (Figure 1.4). In several foreland fold-and-thrust belts, in-coming sedimentary sequences contain supratidal sebkha deposits that have a viscous rheology. Reefal and shallow water limestones behave viscous if they are buried deep enough. An example is the Zagros Simply Folded Belt, where several rheologically weak layers control folding as the dominant mode of shortening (Motiei, 1993; Yamato et al., 2011).
CHAPTER 1. INTRODUCTION

Figure 1.6: Distribution of some fold-and-thrust belts. Colours indicate the type of décollement (from Cooper, 2007).

Most of the shortening within compressional fold-and-thrust belts is accommodated by frontward verging thrust faults. Rocks are either compressed by frontal accretion or by out-of-sequence thrusting (Morley, 1988). Frontal accretion forms imbricate fans. The spacing between thrust ramps and fold wavelengths strongly depends on the thickness of the sediments pile involved in the wedge (Figure 1.5). A minor amount of shortening is often achieved by backthrusts, which are usually steeper than frontward thrust ramps (Figure 1.3c). Consequently, backthrusts are more efficient for crustal thickening.

1.1.2 Décollement rheology

A thin-skinned fold-and-thrust belt is characterized by the basement behaving perfectly rigid without deformation and interaction with the overlying strata. Rock sequences, which are deformed in the thin-skinned style, are detached from their basement rocks along a so-called basal décollement. This décollement is usually within a rheologically weak layer, but this is not always the case. Movement along the basal décollement is triggered by stresses that are built up within the overlying, deforming thin-skinned sequence. The type of décollement is crucial for the structural evolution of a thin-
skinned fold-and-thrust belt. Typically, décollement layers consist of salt layers or watersaturated shale (e.g., Morley et al., 2011; Mukherjee et al., 2010). The rheology of these two types of rock, i.e. the material response to imposed stresses, is very different. At stresses and temperatures occurring at depths of typical décollement horizons, salt is viscous (e.g., Nettleton, 1934). This implies that strain-rates within salt décollements are increased as soon as low stresses are acting on it. Whether the relation between stress and strain-rate in rocksalt is linear or follows a power-law is debated (e.g., Carter and Hansen, 1983; Carter et al., 1982; Chemia et al., 2008, 2009; Marques et al., 2011; Senseney et al., 1992; Spiers et al., 1990; Urai et al., 2008). Typical examples of thin-skinned fold-and-thrust belts on salt décollements are found in orogenic forelands, like in the Zagros, the Jura Mountains and the Salt Ranges in Pakistan (e.g., Davis and Lillie, 1994). Fold-and-thrust belts detaching along salt horizons are also known on passive margins. Examples are the Gulf of Mexico or offshore Gabon (Figure 1.6).

Shale exhibits a brittle rheology within upper crustal pressure and temperature conditions (Byerlee, 1978; Kopf and Brown, 2003; Saffer et al., 2001; Takahashi et al., 2007;
Wang et al., 1980; Figure 1.4). Accordingly, stresses build up till they overcome a material specific yield strength, at which the material breaks and stresses drop to yield stress levels. The relative weakness of shale décollements with respect to overlying rocks is often triggered by high fluid pressures (Fischer and Paterson, 1989; Hubbert and Rubey, 1959). High fluid pressures result from the burial of water-saturated sediments capped by impermeable layers. Porosity decreases with depth due to sediment compaction, whereas the amount of water in the pores remains equal as long as overlying layers are not fractured and remain impermeable. Accordingly, fluid pressures increase towards deeper levels and eventually reach close to lithostatic pressures (Figure 1.7). At these conditions, failure is induced by relatively low shear stresses. Shale décollements occur in all kinds of thin-skinned fold-and-thrust belts and are the main type of décollements in accretionary wedges (Figure 1.6). Section 4.2 provides more specific information about the mechanics of the different types of décollements.

1.1.3 Critical wedge theory

The bulk mechanics of thin-skinned fold-and-thrust belts are considered to be similar to the mechanics of a pile of sand or snow developing a wedge shape if pushed by a bulldozer or a plow (Figure 1.8). Upon this assumption, the analytical model of (Davis et al., 1983) has become popular. This model, called non-cohesive critical wedge theory, assumes perfect plasticity and is based on balancing forces over unit segments along strike between \( x \) and \( x + dx \) (Figure 1.9). Important forces acting on a wedge are i) the

![Figure 1.8: Development of a wedge shaped pile of sand in front of a pushing plow (drawing by Jean-Pierre Burg inspired by Gary Larson).](image)
gravitational body force dependent on the thickness $H$ and density $\rho$ of the rock pile, ii) the pressure of water in the rock body, dependent on thickness $D$ and density $\rho_w$ of the fluid column, iii) the frictional resistance to sliding along the décollement in form of shear traction $\tau_b$ and iv) the compressive push that is induced from the side walls. If $dx \to 0$, forces are balanced and the corresponding equation is:

$$\rho g H \sin \beta + \rho_w g D \cdot \sin \alpha + \beta + \tau_b + \frac{d}{dx} \int_0^H \sigma_x dz = 0$$

(1.1)

where $\alpha$ and $\beta$ are the surface slope of the wedge and the slope of the basal décollement, respectively. Applying the standard small-angle approximation ($\sin \alpha \approx \alpha$, $\sin \beta \approx \beta$), the analytical solution for the total taper of a wedge ($\alpha + \beta$) can be derived depending on the wedge internal friction angle $\phi$, basal friction angle $\phi_b$ and the basal slope of the wedge $\beta$.

Dahlen (1984) presented an exact solution based on the equations of static equilibrium

$$\frac{\partial \sigma_x}{\partial x} + \frac{\partial \tau_{xz}}{\partial z} - \rho g \sin \alpha = 0$$

(1.2)

$$\frac{\partial \tau_{xz}}{\partial x} + \frac{\partial \sigma_z}{\partial z} - \rho g \cos \alpha = 0$$

(1.3)

and showed that the small-angle approximation applied by Davis et al. (1983) is largely valid for all thin-skinned wedges of geological interest. The exact solution provides the
minimum and maximum critical taper angles. The maximum and minimum critical tapers enclose a stable area in a surface slope / base slope diagram (Figure 1.10a-c). Internally strong wedges (large $\phi$) exhibit a larger stable domain and therefore a wider range between the minimum and maximum critical surface slopes (Figure 1.10a). A similar response is obtained with decreasing basal strength (low $\phi_b$). The lower the basal friction, the larger the stable area (Figure 1.10b). If internal and basal friction angles are equal ($\phi = \phi_b$), the stable area shrinks to a line and minimum and maximum critical surface slopes are identical. Steepening or flattening of the wedge results in an immediate response of the surface slope (Figure 1.10a, b). Submarine wedges are water-saturated. The fluid pressure of sea water in the rock pores depends on the height...
of the water column, if there is no over-pressure. In a water-saturated wedge with a rock density of $\rho = 2500\text{kg/m}^3$ and a water density of $\rho_w = 1000\text{kg/m}^3$, the ratio of fluid pressure over confined rock pressure ($P_f/P_r$) is $\lambda = 0.4$. The overall strength of the wedge is lowered and the stable area of the wedge gets smaller with higher $\lambda$ ratios (Figure 1.10c). If a wedge with a given base slope exhibits a surface angle below the critical minimum, then the wedge is subcritical. Theoretically, out-of-sequence thrusting will steepen the wedge surface slope (Morley, 1988). If the wedge is located in the stable area, it can slide along the décollement and accrete new material at its front. This leads to a geometrical lessening of the surface slope. If the wedge with a fixed base slope shows a surface angle above the critical maximum, it is supercritical; normal faulting will flatten the wedge until it reaches the stable area (Figure 1.10d).

The critical wedge theory has been extended to cohesive wedge settings (Dahlen et al., 1984) and to viscous instead of frictional basal shear resistance (Davis and Engelder, 1985). Section 5.3 develops information on the critical wedge theory, including calculation of stress directions and prediction of fault plane orientation.

1.1.4 Questions and aims of the thesis

Although thin-skinned fold-and-thrust belts are among the best and most investigated geological features (e.g., Buiter, 2012; Graveleau et al., 2012; Morley et al., 2011; Poblet and Lisle, 2011), there are many items that have to be examined more closely. One little explored topic is the interplay between surface processes and the dynamic structural evolution of fold-and-thrust belts (e.g., Haghipour et al., 2012; Simpson, 2010a,b; Storti and Mcclay, 1995). Another is the impact of décollement and wedge rheology on the long-term evolution of thin-skinned fold-and-thrust belts (e.g., Bahroudi and Koyi, 2003; Simpson, 2009; Smit et al., 2003). The main goal of this PhD thesis is therefore to better understand the development and dynamics of thin-skinned fold-and-thrust belts in general, with focus on the Makran accretionary wedge and the Zagros foreland fold-and-thrust belt.

A link between surface processes and tectonic deformation in foreland basins is represented by progressive unconformities, i.e. growth strata (Riba, 1976; Suppe et al., 1992; Verges et al., 2002). Within growth strata, the structural evolution and deformation style of single structures, such as folds or faults, is captured. Investigation of such growth strata leads to insights into the relationship between deformation and time. This is achieved with a precise documentation and dating of sequences comprising growth
strata. Different architectures of unconformities are due to different folding and thrusting mechanisms, such as limb rotation vs. hinge migration (e.g., Hardy and Poblet, 1994; Rafini and Mercier, 2002; Figure 1.11a), changes in accumulation (e.g., Poblet et al., 1997) and accelerated vs. slowed amplification of folds (e.g., Anadon et al., 1986; Riba, 1976; Storti and Poblet, 1997; Figure 1.11b). High-resolution dating of progressive unconformities gives results for i) sedimentation rates during formation of the growth strata and ii) rates of fold amplification by comparing the stratigraphic age with dip angle. The most appropriate dating method is, for this purpose, magnetostratigraphy. Combining the results for rates of sedimentation and folding gives some clues about the predominant surface processes at the time and location of the development of growth strata.

Numerical modelling is a practical method to investigate long-term and large-scale tectonic mechanisms in compressional settings. Therefore, focus is placed on forward mod-
elling of the deformation of brittle sedimentary rocks overlying a weak décollement. Sedimentation can be implemented in the code, which allows investigating the influence of sedimentation on the deformation style. Analogue models suggest that syntectonic sedimentation can strongly affect the thrust geometry (Barrier et al., 2002; Storti and Mcclay, 1995).

1.2 Study area

1.2.1 Makran accretionary wedge

One of the largest and still active accretionary wedges is the Makran. It extends about 1000 km between the Minab dextral transform fault to the west in SE Iran and the sinistral, Chaman transform fault to the east, in Pakistan (Figure 1.12). The Makran belongs

![Figure 1.12](image)

**Figure 1.12:** Tectonic setting of the Makran accretionary wedge (hatched area) including linear velocities (arrows with numbers = N–S movement component in mm/a) of the Arabian plate (MUSC = Muscat), the Makran coast (CHAB = Chabahar) and Central Iran (BAZM = Bazman) with respect to stable Eurasia after Vernant et al. (2004). Framed: studied area. Focal mechanism of the 8.1 magnitude earthquake of November 27, 1945, after Byrne et al. (1992). Figure and caption from Haghipour et al. (2012).
to the Alpine-Himalayan orogenic system. It results from the convergence between the 
Arabian and Eurasian plates, which is active since at least the Late Cretaceous (Byrne 
et al., 1992). The Makran accretionary wedge grows both vertically and laterally by 
scraping sediment material off the northwards subducting Arabian lithosphere. Plate 
convergence rates are currently between 2 and 4 cm/a (Bayer et al., 2006; Vernant et al., 
2004; Vigny et al., 2006). The estimated southward migration of the wedge toe is ∼ 
1 cm/a (Platt et al., 1985; White and Louden, 1982). The Makran is divided into an 
active southern part and a more passive northern part, which are separated by a normal 
fault system close to today’s shoreline (Ellouz-Zimmermann et al., 2007a; Grando and 
McClay, 2007)). In contrast to previous work focussing on seismic profiles through the 
offshore Makran (Ellouz-Zimmermann et al., 2007a,b; Fruehn et al., 1997; Grando and 
McClay, 2007; von Rad et al., 2000) and on the Pakistani part of the accretionary wedge 
(Corporation, 1960; Kopp et al., 2000; Platt and Leggett, 1986a,b; Platt et al., 1988, 
1985), not much work has been done on the onshore Iranian Makran (Burg et al., 2008,
1.2. STUDY AREA

Figure 1.14: Typical landscape formed by the Tortonian Olistostrome. Top: 26°08'20.3"N / 59°34'42.5"E. Bottom: 26°30'26.3"N / 60°37'26.4"E (image courtesy by Jean-Pierre Burg).

2013; Dolati, 2010; Dolati and Burg, 2013; Hosseini-Barzi and Talbot, 2003; McCall, 1997, 2002, 2003). This can be explained by the past political situation of Iran and the rugged and not easily accessible landscape of the onshore Makran. Nevertheless, the extremely good outcrop conditions and the excellent sedimentary and stratigraphic record make the Makran an excellent field area to study tectonic and surface processes of an accretionary complex.

The onshore Iranian Makran has been divided into four main segments (Figure 1.13; Dolati, 2010). From north to south, structurally from top to bottom, those are: i) North Makran including mafic to intermediate igneous rocks, tectonic mélanges and Cretaceous deep-water sediments, ii) Inner Makran dominated by Eocene to Lower Miocene turbidite
deposits, iii) Outer Makran exhibiting shallow-water Miocene sandstones and marls and iv) Coastal Makran representing a regressive sequence from slope marls to continental deposits.

Towards the North the imbricate and mélange zone of North Makran is bordered by desert areas (Figure 1.13). The Jaz Murian (Iran) and Meshkel (Pakistan) depressions represent the fore-arc deposition basins (Figure 1.2; Harrison, 1943; McCall, 1997). These two depressions are geographically divided by the NNW–SSE striking Sistan fold-and-thrust belt (Tirrul et al., 1983). To the North of the fore-arc basins, three major and active volcanic centres represent the magmatic arc of the Makran subduction zone (Figure 1.2a): Kuh-e Bazman (Saadat and Stern, 2011) and Kuh-e Taftan (Biabangard and Moradian, 2008; Gansser, 1971) in Iran and Kuh-e Soltan (Nicholson et al., 2010) in Pakistan.

An important characteristic of the Iranian Makran accretionary wedge is the occurrence of a huge, Early Tortonian olistostrome (Burg et al., 2008), which was earlier interpreted as a sedimentary coloured mélange (Gansser, 1960; McCall, 2002; Stocklin, 1968). This gigantic mud-and-debris flow covers today an area of $\sim 10{,}000 \text{ km}^2$ with an incomparable landscape (Figure 1.14).

The offshore Makran has been intensely investigated by seismic profiling (e.g., Grando and McClay, 2007). The Late Oligocene and Miocene sediments are detached over one or more décollement layers formed by water-saturated and shale-rich sediments within turbidites (Figure 1.15). The northern part of the offshore Makran consists of a shelf and a flat slope with high accumulation rates leading to normal faulting seen in Coastal

Figure 1.15: Interpretation of a seismic profile through the offshore Iranian Makran. Vertical exaggeration at the sea floor: $\sim 1:6$ (from Grando and McClay, 2007).
1.2. STUDY AREA

Makran (Dolati, 2010; Dolati and Burg, 2013; Ellouz-Zimmermann et al., 2007a). The southern part of the offshore Makran is a typical, active imbricate fan (Figure 1.15).

Figure 1.16: Mud volcanoes in the Coastal Makran. a) and b) 25°28'01"N / 59°55'59"E. c) 25°30'25"N / 59°59'48"E. d) 25°22'57"N / 61°17'33"E. e) 25°26'35"N / 61°18'41"E. Image courtesy of (a) by Marcel Frehner.
In the Makran, mud volcanoes have been found offshore (Grando and McClay, 2007; Schluter et al., 2002; von Rad et al., 2000; Wiedicke et al., 2001) and onshore Pakistan (Delisle and Berner, 2002; Delisle et al., 2002; Schluter et al., 2002). Along the coast of the Iranian Makran, three regions with mud volcano activity were identified. Several mud volcanoes are located at the South-western edge of the Makran, in the Jask area (McCall, 2002). A second patch of three mud volcanoes is located south of the village of Kahir (Figure 1.16a-c). At least three mud volcanoes are presently located around the village of Rimdan, close to the Iran-Pakistan border (Figure 1.16d and e).

Mud volcanoes in an accretionary wedge environment are presumably sourced in the décollement zones, where fluid pressure is equal or close to lithostatic pressure (Brown, 1990; Mazzini, 2009; Platt, 1990). They rise along faults and are able to penetrate the whole sediment pile of a wedge (Brown and Westbrook, 1988, 1987). This makes them very important for detecting wedge décollement level locations and investigating their properties. It has been stated that extrusions represent only a small part of the material recycled to the surface. Huge amounts intruded into younger formations in form of diapirs (Platt et al., 1985).

Apparently, there are different source levels for the rising mud in the Pakistani part of the Makran. Delisle et al. (2002) interpret the source of extruded onshore mud to be Quaternary to at least Miocene stratigraphic levels at minimum depths of 2.3 km based on considerations of the required overpressures, age- and environment-diagnostic benthic foraminifera and coccoliths. Schluter et al. (2002), on the other hand, favour the idea that offshore mud extrusions are partly sourced by the interpreted Upper Cretaceous to Palaeogene hemipelagic sequence of the Arabian Plate.

In Makran, growth strata appear onshore in Miocene-Pliocene sequences in small amplitude, long wavelength synclines (Burg et al., 2013). In the submarine Makran, progressive unconformities are documented atop rotated thrust sheets of the frontal, imbricate fan (Grando and McClay, 2007).

1.2.2 Zagros foreland fold-belt

As Makran, the Zagros Fold Belt belongs to the Alpine-Himalayan orogenic system. It extends ~ 2000 km between the Minab-Zendan strike slip system in the southeast to the East Anatolian fault in the northwest (Figure 1.17). Zagros formed after closure of the Thetys Ocean, which started during the Late Cretaceous (e.g., Agard et al., 2005;
Berberian and King, 1981). GPS velocities indicate a present-day, northward movement of the Arabian plate at \( \sim 2 \text{ cm/a} \) (Vernant et al., 2004), \( \sim 1 \text{ cm/a} \) of which is absorbed in the Zagros Fold Belt (Masson et al., 2007). From NE to SW, the Zagros is divided into four major tectonic zones (Figure 1.17): i) The Urmieh-Dokhtar magmatic arc, ii) the Sanandaj-Sirdjan zone, iii) the Imbricate zone (or High Zagros) and iv) the Simply Folded Belt. The suture, the Main Zagros Thrust, is placed between the Sanandaj-Sirdjan zone and the Imbricate Zone (Fig. 1.17). This work focuses on the Zagros Simply Folded Belt, which is considered to be a basement-involved thin-skinned fold-and-thrust belt (McQuarrie, 2004; Mouthereau et al., 2006, 2007b; Talebian and Jackson, 2004).

\[ \text{Figure 1.17: Overview map of the Zagros orogen (Map from Cesciello et al., 2009; profile from McClay et al., 2004 after Alavi, 1994).} \]
The Zagros Simply Folded Belt is characterized by large wavelength (∼15 km) NW-trending concentric folds (Figure 1.18; Mouthereau, 2011). The folded stratigraphic sequence consists of limestones, shale and sandstones and evaporitic sediments deposited during rifting (Perm-Trias), at passive (Jurassic-Eocene) and at active (Oligocene-present) margins (Agard et al., 2011; James and Wynd, 1965; Motiei, 1993; Robin et al., 2010; Verges et al., 2011). The deformed thin-skinned sequence has a thickness of 10−12 km (Alavi, 2007; Colmansadd, 1978; Motiei, 1993) and is detached along the 1−2 km thick Precambrian to Cambrian Hormuz salt in the Fars (Bahroudi and Koyi, 2003; Edgell, 1996) and along Cambrian shales in the Lurestan (Sherkati and Letouzey, 2004). Several weak layers are identified within the sedimentary pile (Motiei, 1993). At the top of the stratigraphic sequence, the Neogene Fars group represents a ∼3 km thick regressive cycle representing shallow-marine to shelf limestones and marls, followed by fluvial plane reddish mud-, silt- and sandstones and finally covered by Pliocene alluvial, braided river conglomerates (Falcon, 1974; James and Wynd, 1965; Stocklin, 1968). The Late-Miocene to Pliocene accumulation of the fluvial and alluvial deposits is contemporaneous with the thin-skinned shortening of the Simply Folded Belt. This is inferred from progressive and angular unconformities throughout the Zagros (Figure 1.19; Emami, 2008; Hessami et al., 2001; Homke et al., 2004; Khadivi et al., 2010; Mouthereau et al., 2007b; Saura et al., 2011).
1.3 Methods

1.3.1 Magnetostratigraphy

In the early twentieth century, scientists discovered rocks that were magnetized in a reversed manner to the magnetic field observed today (e.g., Matuyama, 1929; Mercanton, 1926). A magnetization polarity like that today is called normal polarity. If it is in the opposite state, the polarity is reverse. Reversals are documented in rocks as old as the Archean (Strik et al., 2003). The polarity sequence was split up into epochs of dominant normal and reverse polarity (Cox et al., 1964), which are today known as chrons. The two polarity states, normal and reverse, occurred for roughly the same amount of time over the history of Earth (Tauxe, 2010). The migration of magnetic directions from one
CHAPTER 1. INTRODUCTION

state to the other takes some thousand years. Therefore, the magnetic field spends $\sim 1 - 2\%$ of the time in intermediate polarity states. The Geomagnetic Polarity Timescale (GPTS) describes the changes in magnetic earth field polarity through time. The GPTS is well defined for the last 160 Ma (Berggren et al., 1995), while older polarities are more difficult to reveal (Kent and Olsen, 1999). The average frequency of reversals was $\sim 4$/Ma for the last 20 Ma (Constable, 2003).

Surveys with ship-towed magnetometers revealed 'magnetic stripes' on the sea floor, which inspired the model of seafloor spreading (Figure 1.20; Mason and Raff, 1961; Vine and Matthews, 1963). Magnetostratigraphy of sedimentary drill cores were then correlated with the GPTS for the Quaternary (Harrison, 1966; Opdyke et al., 1966) and later back into the Miocene, to $\sim 20$ Ma (Opdyke et al., 1974). Sedimentary rocks deposited in water bodies store the magnetization polarity at the time of their deposition when magnetic particles sink down the water column and orient along the contemporaneous apparent magnetic field. These particles are locked during sedimentation and diagenesis and define the characteristic remanent magnetization (ChRM) of the bearing sedimentary layer. A sequence of sedimentary stratigraphy, offshore or onshore, can therefore be correlated to the GPTS (Kent and Olsen, 1999; Kent et al., 1995), if the ChRM is

**Figure 1.20:** Areas of positive and negative magnetic polarity offshore California inferred from ship-towed magnetometer measurements (from Mason and Raff, 1961).
1.3. METHODS

Figure 1.21: Example of fitting an obtained magnetostratigraphic polarity interval sequence to the Geomagnetic Polarity Time Scale (GPTS) to infer age and accumulation rates (connecting lines between obtained polarity sequence and GPTS; numbers in cm/ka) of the investigated section (from Khadivi et al., 2010).

recorded. The possibility to correlate chrons obtained from magnetostratigraphy with biostratigraphic ages of the same sequence makes it a powerful dating tool, even for relatively short stratigraphic sections with only few reversals (Lourens, 2004). Magnetostratigraphy with a good sampling resolution yields age constraints for the investigated sequence, and thus constrains average accumulation rates as the ratio between sequence thickness and time (Figure 1.21).

Magnetostratigraphy has been shown to be a viable method to date stratigraphic sequences in foreland settings (Burbank and Raynolds, 1988; Burbank et al., 1992; Jordan and Alonso, 1987; Pares et al., 2003; Reynolds et al., 1990; Schlunegger et al., 1997). It was successfully applied to tectonic deformation, such as fold growth and rotation (Bur-
1.3.2 Numerical modelling

Numerical modelling is efficient to investigate the long-term structural behaviour and the influence of rheologies on the evolution of thin-skinned fold-and-thrust belts. A main characteristic of fold-and-thrust belts is the occurrence of faults, which represent brittle failure of the deforming rocks. Therefore, numerical codes suitable to model brittle wedges must be able to develop shear bands depending on a predefined failure criterion. Preferred numerical techniques successfully applied to simulate fold-and-thrust belts are the Discrete (or Distinct) Element Method (DEM), the Finite Element Method (FEM) and the Finite Difference Method (FDM).

The DEM is a particle-based method, where particle elements can theoretically have any shape and size (Cundall and Strack, 1979). These particles only interact directly with each other or a defined wall, depending on their specific distance. The soft-contact DEM allows elements to overlap (Cundall and Hart, 1989). The assembly of particles in a DEM model can mimic granular materials, such as sand (Morgan, 1999; Morgan and Boettcher, 1999). DEM models have been applied to compressional settings to simulate fault initiation and propagation (Benesh et al., 2007; Finch et al., 2003; Hardy and Finch, 2005; Homberg et al., 1997; Hu et al., 2001) and to model fold-and-thrust belts in particular (Burbidge, 2000; Burbidge and Braun, 2002; Miyakawa et al., 2010; Naylor et al., 2005; Wenk and Huhn, 2013; Yamada et al., 2004). The main advantage of the DEM formulation is spontaneous strain localization and large-scale strain accumulation on discontinuous surfaces (Figure 1.22a; Miyakawa et al., 2010).

The FEM provides solutions of various problems by approximating an overall equation by separating it into many simple element equations (Zienkiewicz and Taylor, 1991). To simulate deformation in fold-and-thrust belts, models are based on the continuum equations. The FEM is the most applied technique to model large-, lithospheric-scale structural deformation along compressional settings (Beaumont et al., 1992; Buiter et al., 2009; Selzer et al., 2008; Willett et al., 1993). Especially, FEM suits very well modelling of thin-skinned tectonics related to surface processes. The upper boundary can be prescribed as free-surface, where normal and shear stresses are zero (Braun, 2006). Borja and Dreiss (1989) presented the first mechanical FEM model directly applied to accretionary wedges. Since the 1990’s, several FEM approaches have dealt with brittle/plastic
1.3. METHODS

Figure 1.22: Examples of wedge models obtained with A) DEM (Miyakawa et al., 2010), B) FEM (Stockmal et al., 2007) and C) FDM (Buiter et al., 2006; model by Taras Gerya) of thin-skinned fold-and-thrust belts.

thin-skinned fold-and-thrust belts (Figure 1.22b), focusing on internal or basal rheology, surface processes influencing the overall dynamics, application to the theoretical critical wedge theory, the effect of gravitational load leading to foreland flexure, the influence of weak layers within the deforming sequence, variable shapes of the backstop and comparison to analogue models (Braun and Yamato, 2010; Buiter et al., 2006; Byrne et al., 1993; Ellis et al., 2004; Fillon et al., 2012; Nilfouroushan et al., 2012, 2013; Selzer et al., 2007; Simpson, 2006, 2009, 2010a,b, 2011; Souloumiac et al., 2010; Stockmal et al., 2007; Strayer and Hudleston, 1997; Strayer et al., 2001; Yamato et al., 2011).

The FDM (Patankar, 1980), also a continuum method, was less applied in investigations of thin-skinned fold-and-thrust belts (Figure 1.22c). FD are linear mathematical expressions used to approximate derivatives to a certain degree of accuracy (Gerya, 2010b). Particles that freely advect through a static grid on which the equations are solved can be introduced. In geodynamics, FD is often used to simulate large-scale, large-finite strain problems, such as subduction processes, mantle-lithosphere interaction and mantle convection (e.g., Duretz et al., 2012; Gerya, 2011; Ueda et al., 2012). The major advantage of the FDM compared to the FEM is the smaller amount of matrix entries for an equal
amount of nodal points in a numerical setup. This makes the FDM effective to investi-
gate high-resolution three-dimensional geodynamic processes (Gerya, 2013). Concerning
thin-skinned evolution of fold-and-thrust belts, the problem of having a lack of direct
free-surface boundary condition at the top is solved by introducing a quasi free-surface
in form of "sticky-air", a low-viscosity, low-density material, which mimics successfully
a free-surface at its interface with the underlying, deformed rock pile (Crameri et al.,
2012).

1.4 Outline of the thesis

This PhD thesis presents results obtained from field work and numerical modelling.
In particular, magnetostratigraphic dating of progressive unconformities in the Makran
and the Zagros offer better constraints on the timing of deformation. Numerical models
help understanding the large-scale mechanics and the influence of rheology on brittle
wedges.

Chapter 2 focusses on unconformities within long-wavelength, low amplitude folds in
Upper Miocene layers of the onshore Makran accretionary wedge. Unfortunately, paleo-
magnetic analyses were not successful because of the magnetic properties of the sampled
sedimentary rocks.

Chapter 3 presents new insights on the timing of deformation inferred from magne-
tostratigraphy in the Zagros Simply Folded Belt. A section through the complete suc-
cession of foreland deposits in the Central Fars arc containing a progressive unconformity
at its top was studied. Age constraints of the investigated growth strata and compar-
sion to similar studies in other places of the Simply Folded Belt revealed an in-sequence
growth of the thin-skinned fold belt over $\sim 10$ Ma. This chapter is submitted as Ruh et
al. to Tectonics.

Chapter 4 investigates the influence of single and multiple décollements on the defor-
mation mechanics in fold-and-thrust belts by using a visco-elasto-plastic Finite Element
numerical model. The differences between viscous (salt) and frictional (shale) décolle-
ment rheology are discussed. Results are compared to the critical wedge theory, analogue
sandbox models of brittle wedges and natural examples. This chapter is published as
Ruh et al. (2012) in Tectonics.

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1.4. OUTLINE OF THE THESIS

Chapter 5 contains 3D high-resolution numerical models that test the influence of décollement strength on transfer zones in thrust wedges. A visco-plastic/brittle Finite Difference model suitable for thin-skinned fold-and-thrust belts was developed. This chapter is published as Ruh et al. (2013) in *Geochemistry, Geophysics, Geosystems*.

Chapter 6 focusses on the evolution of an oblique, transpressive fold-and-thrust belt by using a Finite Difference numerical model. In the Zagros Fold Belt, obliquity is given by the strike of the Zagros main suture zone, the High Zagros Fault, to the normal direction of the movement of the Arabian plate (Figure 1.17). In the applied model setup, a backstop geometry oblique to the compression direction is introduced. It is shown that the concentric, elongated folds characteristic of the Zagros Simply Folded Belt result from transpressive compression.

Chapter 7 investigates the effect of plastic weakening of shear zones within thin-skinned fold-and-thrust belts. The same numerical code as in Chapter 5 and 6 is applied to test strain vs. strain-rate weakening. Geometrical setups including changes in décollement strength revealed the impact of plastic weakening on the evolution of transform faults in brittle wedges. This chapter is under review as Ruh et al. in *Tectonophysics*.

Chapter 8 comprises a short discussion and main conclusions of the thesis.
Progressive unconformities are predestined to link sedimentary and deformation processes in tectonically active settings. Growth strata are found in Late Miocene shallow-marine marls and calcareous sandstones within large wavelength small amplitude concentric synclines of the onshore Makran accretionary wedge. The morphology of these growth strata indicates an underfilled basin where fold amplification was faster than sedimentation, forming typical sedimentary offlap structures. Magnetostratigraphy of a ~ 1200 m thick section comprising 43 drill sites has been attempted. > 50% of the natural remanent magnetization is removed by step wise thermal heating already above 200°C. No sample yielded a characteristic remanent magnetization. Low temperature, viscous remanent magnetizations are aligned in the direction of the recent Earth’s magnetic field. Viscosity tests and rock magnetic data indicate that the ferromagnetic (s.l.) mineral is in the grain-size range of superparamagnetic for these sediments. Particles in this size range exhibit short times of magnetic relaxation. Therefore, the Miocene magnetization direction is not preserved in the sampled sequence.
2.1 Introduction

The Makran extends over about 1000 km between the Minab dextral transform fault to the west (e.g., Peyret et al., 2009; Regard et al., 2005) and the sinistral Chaman transform fault to the east, in Pakistan (Figure 2.1; Jadoon and Khurshid, 1996). The Makran is situated in the zone of convergence that is active between the Arabian and Eurasian plates since at least the Late Cretaceous (Byrne et al., 1992). The wedge of essentially Eocene–Holocene sediments results from the on-going subduction of the oceanic lithosphere flooring the Gulf of Oman. Northward subduction is responsible for the Pleistocene-Holocene andesitic volcanic arc and the Jaz-Murian and Meshkel forearc basins to the north (Figure 2.1). Shortening and abundant sediment supply led to the formation of the accretionary prism that grew seawards by frontal accretion and underplating of trench fill sediments (Burg et al., 2008; Ellouz-Zimmermann et al., 2007a; Harms et al., 1997; Platt et al., 1985).

Figure 2.1: Tectonic setting of the Makran accretionary wedge between the Minab (West) and the Chaman (East) fault systems. Red square indicates location of Figure 2.2. (Source: http://maps-for-free.com/)
New results on the stratigraphic and tectono-sedimentary history show that the present-day on-land Makran was a turbidite basin on an active margin between the Late Paleocene - Early Eocene and the Serravallian (Burg et al., 2013; Dolati, 2010). The onshore Iranian Makran has been divided into four main segments: the North Makran including mafic to intermediate igneous rocks, tectonic mélanges and Cretaceous deep-water sediments; the Inner Makran dominated by Eocene to Lower Miocene turbidite deposits; the Outer Makran exhibiting shallow-water Miocene sandstones and marls; and the Coastal Makran representing a regressive sequence from slope marls to continental deposits (Figure 2.2). The deposits became more proximal and the influx of turbidites ceased by the Lower Miocene in the Outer and Coastal Makran. Carbonate reefs and
gypsiferous mudstones indicate a shallow marine and lagoonal environment during the Burdigalian (Dolati, 2010; McCall et al., 1994). The Makran includes a giant catastrophic mud-and-debris flow emplaced in the Lower Tortonian (Burg et al., 2008; Dolati, 2010). The olistostrome includes blocks of ophiolites and oceanic sediments derived from the ophiolite-bearing, imbricate thrust zone to the north, and reworked chunks of the turbidites on which it rests with an erosional unconformity. Its size and internal structure make it a fossil equivalent of the large debris flows found along continental margins and unstable volcanic edifices. An important characteristic of accretionary wedges is the contemporaneous deformation and sedimentation, which generated growth structures (Riba, 1976; Verges et al., 2002). In Coastal Makran, growing anticlines form solitary mountains and ridges that dominate an otherwise flat landscape, filled in by the young deposits of the coastal plain. Landward from the coastal plain, isolated > 100 m high, dish-shaped open synclines have a wavelength > 10 km (Figure 2.3). Growth strata are found in such concentrical synclines folding the Late Miocene, so-called Dar Pahn unit along the northern boundary of Coastal Makran (Figure 2.2; Dolati, 2010; McCall, 1985).
2.2 Growth strata within the Dar Pahn unit

The Late Miocene Dar Pahn unit occurs along the northern edge of Coastal Makran in the direct footwall of the Chah Khan Thrust, which marks the tectono-stratigraphic boundary with Outer Makran, to the north (Figure 2.2). According to Dolati (2010), the Dar Pahn unit is divided into two members: i) a marl-dominated subunit (Figure 2.4) and ii) a calcareous sandstone-dominated subunit (Figure 2.5). The marl-dominated member consists of > 70% light grey, medium to very thick bedded marls and fine to medium grained and thin to medium bedded calcareous sandstones (Dolati, 2010). The calcareous sandstone subunit consists of thick layers of channelized sandstones (Figure 2.5). Coarsening-up, cross-bedding and herringbone structures are documented (Dolati, 2010). The sandstone layers of the two subunits are similar. The difference between these subunits is the abundance and thickness of the sandstone layers.

A magnetostratigraphic study has been conducted along a section through the progressive growth of one of these long-wavelength synclines in order to constrain the age of sedimentation. Detailed rock magnetic experiments were performed to identify the ferromagnetic mineralogy in the sediments.

Figure 2.4: Landscape formed by the marl dominated member of the Dar Pahn unit. A) 25°59′52.7″N / 59°46′26.8″E. B) 26°17′09.4″N / 59°54′15.8″E.
Figure 2.5: Thick-bedded calcareous sandstone of the sandstone-dominated member of the Dar Pahn unit. Cross-bedding indicates currents possibly related to foreshore, deltaic deposition. Location: 25°49′04.3″N / 59°58′34.5″E.

Well-exposed growth strata within an open, concentric, long-wavelength syncline north of the village of Hansoum (Figure 2.2) marks synchronous deformation and sedimentation (Figure 2.3). This part of the sedimentary sequence is dominated by grey marls interlayered with thin to medium bedded calcareous sandstones (Figure 2.6). Samples for the magnetostratigraphic study were taken along a profile across the west-southwest limb of the synclinal structure (Figure 2.3). Along this profile, an angular unconformity is detected from bedding orientations (Figure 2.7). The sequence thickens towards the southeast of the profile, along the bedding strike (Figure 2.3). The fact that the sedimentary layers fade out laterally within the syncline and are thickest in its centre suggests accelerated fold growth in an underfilled sedimentary basin.
2.2. GROWTH STRATA WITHIN THE DAR PAHN UNIT

Figure 2.6: Thin to medium bedded marls alternating with calcareous sandstones of the studied section. Location in Figure 2.3.

Figure 2.7: Profile A - A' (SW - NE) in Figure 2.3. The sequence shows normal younging direction. Each dip symbol refers to a sampling site, symbol inclination indicates bedding dip. Red and blue dots in the Pi-plot refer to bedding dip along the profile. Blue describe pre-growth, red show measurements in post growth sediments. (Interpreted sedimentary contact)
2.3 Paleomagnetic samples and methods

Along the studied section, 43 samples have been drilled at a regular interval in calcareous sandstone layers using a portable gasoline drill. The samples were cut into cylinders of 2.5 cm diameter and 2.3 cm length. The natural remanent magnetization (NRM) of every sample was measured before the samples were stepwise demagnetized. The NRM and the magnetization after every demagnetization step was measured with a 2G Enterprise 755R, 3-axis DC-SQUID rock magnetometer at the Laboratory for Natural Magnetism of the ETH in Zurich. Magnetic susceptibility was measured with an AGICO KLY-2 Kappabridge on all samples before demagnetization. Samples were demagnetized with either thermal or alternating field treatment. Thermal demagnetization was conducted with a Schoenstedt TSD-1 and an ASC TD48 oven with an internal field < 5 nT. Alternating field demagnetization has been conducted on 30 samples with a demagnetizing system integrated into the cryogenic magnetometer. The demagnetization behaviour was plotted on vector diagrams to identify components of magnetization (Zijderveld, 1967) and these were isolated using the Remasoft software (Chadima and Hrouda, 2006).

Acquisition of isothermal remanent magnetization (IRM) has been performed on five representative samples. The IRM was applied with an ASC Impulse Magnetizer (Model 10-30), and measured on the cryogenic magnetometer. A cross-component IRM test was conducted with 1.2 T along the sample z-axis, 0.4 T along the sample y-axis and 0.12 T along the sample z-axis (Lowrie, 1990). The samples were subsequently thermally demagnetized to monitor the thermal unblocking of the three coercivity components. Hysteresis parameters, such as saturation magnetization ($M_s$), saturation remanence ($M_r$), coercivity ($H_c$) and coercivity of remanence ($H_{cr}$), and first-order reversal curves (FORCs; Pike et al., 1999, 2001; Roberts et al., 2000) were measured with a MicroMag 2900 Alternating Force Gradient Magnetometer. Storage tests were carried out to detect whether the magnetization of the specimen acquires a component of viscous magnetization during storage in the ambient field in the laboratory (i.e., the recent Earth’s magnetic field).

2.4 Paleomagnetic results

All measured samples reveal an intensity of the NRM between $12 \cdot 10^{-6}$ and $60 \cdot 10^{-6}$ A/m with an average of $26 \cdot 10^{-6}$ A/m. Thermal demagnetization was carried up in 16
incremental steps until 600°C, at which point the samples were demagnetized. Figure 2.8 illustrates that the samples lose more than half their original magnetization by 200°C. After removal of a random viscous component with the first heating step at 75°C, a magnetization component was removed up to 250 to 400°C (Figure 2.9). Alternating field demagnetization shows that a low coercivity component is removed below approximately 10 mT, and a second component, which is similar to the stable component seen in thermal demagnetization, is removed until 140 mT; samples retain a high-coercivity magnetization above 140 mT (Figure 2.9).

Figure 2.10 shows the distribution of directions that were isolated from vector analysis in geographic coordinates after thermal and AF demagnetization. Specimens only carry a normal polarity, and directions (thermally demagnetized) are better grouped in geographic coordinates ($k = 47.2$, $\alpha_{95} = 3.2^\circ$) than if corrected for bedding tilt ($k = 26.9$, $\alpha_{95} = 4.3^\circ$). Thermal demagnetization shows a tighter grouping of directions than AF demagnetization, but in both cases the mean direction is statistically the same as the present field direction. This suggests that magnetization has been acquired recently, after the sediments have been tilted. Rock magnetic tests were performed to help understanding the origin of this recent magnetization.

IRM acquisition curves indicate a rapid increase in magnetization below 200 mT, followed by a gradual increase, in which the samples are close to saturation by 1200 mT (Figure 2.11). $H_{cr}$ is similar for all specimens (~ 50 mT). The soft (low coercivity) component dominates the cross-component IRM. It is gradually thermally unblocked between room temperature and 600°C (Figure 2.12). The medium coercivity fraction of
Figure 2.9: Representative diagrams of tectonically NRM thermal demagnetization curves in X coordinates. Open and filled circles represent vertical and horizontal projections, respectively. Subfigures a) to c) are tilt-corrected coordinates. d) to f) are geographically corrected.
2.4. PALEOMAGNETIC RESULTS

Figure 2.10: Lower hemisphere, equal area plots showing the stable components of magnetization and their statistical grouping for a) thermal and b) alternating field demagnetization. The mean direction of both groups (magenta circle) is very close to the recent magnetic field in the Makran (blue diamond: Dec: 2°9', Inc: 46°12'. Source: http://www.ngdc.noaa.gov/geomag-web/#igrfwmm).

the cross-component IRM fully unblocks at 600°C. The hard (high coercivity) component demagnetizes between 600 and 700°C (Figure 2.12).

FORC analysis indicates that there is a large superparamagnetic (SP) component to the coercivity spectra of the ferromagnetic minerals. This can be seen from the density peak around $H_c$, $H_u = 0$ (Figure 2.13), which indicates small relaxation times as the particles overcome their energy barrier and switch field orientation (Roberts et al., 2000). There is no substantial multi domain (MD) fraction detectable from the FORC diagrams. Interacting MD particles would result in a broad spreading of FORC densities along the $H_u$-axis, which is not the case (Figure 2.13c). The narrow, elongated density distribution along the $H_c$-axis at $H_u = 0$ indicates the presence of non-interacting single domain (SD) particles (Figure 2.13d). If SD grains would be interacting, the density peak would not be centred on $H_u = 0$ (Roberts et al., 2000).

Storage tests show that the NRM of three samples have a viscous component (Figure 2.14). The NRM was measured in October and then stored in a stagnant position for two months. Measurement of the NRM in December reveal a viscous migration of the
CHAPTER 2. MAGNETOSTRATIGRAPHY IN THE COASTAL MAKRAN

Figure 2.11: Low remanence coercivities defined from five representative normalized IRM acquisition curves.

Figure 2.12: Thermal demagnetization of the soft (< 0.12 T), medium (0.12 - 0.4 T) and hard (0.4 - 1.2 T) coercivity fractions of three-component IRM.
2.5 Discussion

The onshore Makran accretionary wedge exhibits continuous, well preserved sedimentary sequences and an excellent stratigraphic record. Thanks to tectonically undisturbed, kilometre thick Oligocene to Late Miocene sedimentary rock sequences, the Makran is predisposed to investigate the interaction between tectonic and surface processes.

Late Miocene shallow-water stratigraphy in Coastal Makran exhibits unconformable sedimentary layering indicating fold-growth during sedimentation. The studied section cuts a concentric syncline that contains growth strata in its Southern limb (Figure 2.3 and 2.7). Similar synclines developing in a shallow submarine basin are found in the

NRM direction towards the recent Earth’s magnetic field (Figure 2.14b). Intensities of the magnetization in all investigated samples increased during the storage. This is explained by the fact that before stagnant storing of the samples, the NRM was not forced by a static magnetic field but the moving sample was constantly influenced and overprinting earlier magnetizations.

Figure 2.13: a) Series of first-order reversal curves (FORCs). Colour indicates dimensionless measurement density (Harrison and Feinberg, 2008; Roberts et al., 2000). b) Contour plot FORC diagram of data in (a). c) Density profile at $H_c = 0$. d) Density profile at $H_u = 0$. 

2.5 Discussion
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Figure 2.14: Results of the storage test. a) location of the NRM orientations in a Wulff-stereoplot. Circles = MI-30, triangles: MI-31, squares = MI-33. b) Enlargement of square in (a). Diamond = recent magnetic Earth’s field in Zurich. c) Intensity plot showing data of October and September.

frontal imbricate fan of the offshore Makran, where limb rotation in the hanging wall of newly accreted sequences generate growth strata (Figure 2.15).

The investigated syncline exhibits stratigraphic thickening of the unconformity layer towards the southeast (Figure 2.3 and 2.7). This shows that i) the syncline was initially concentric, and ii) the apparent sedimentary basin was underfilled, i.e. the fold growth was faster than the sedimentation rate.

Magnetostratigraphy should provide a method for dating these sediments, because the average duration of a normal or reverse polarity interval is $\sim 200$ ka over the last 15 Ma. Although the precision of age constraints is dependent on the spatial resolution of sampling, it should have been possible to isolate a polarity record in the sample profile. Acquisition of a field direction in the laboratory indicates that the original Miocene magnetization may have been completely reset during the past 780’000 years of the Brunhes normal polarity chron. Therefore, these rocks are not suitable for paleomagnetic reconstructions or dating.

In general, samples from the investigated sequence exhibit a relatively low NRM ($M_{av} = 26 \cdot 10^{-6}$ A/m) and are fully demagnetized at 400 to 500°C (Figure 2.8 and 2.9), below the Curie’s temperature of magnetite and hematite. Nevertheless, cross-component IRM demagnetization patterns indicate low-coercivity magnetite (Curie temperature =
2.5. DISCUSSION

Figure 2.15: Detailed seismic profile of a piggy-back, synclinal basin in the Makran imbricate fan in the Oman sea showing the geometric relationship between sedimentation and fold growth. (from Ellouz-Zimmermann et al., 2007b).

Figure 2.16: Relaxation time in magnetite ellipsoids as a function of grain width (from Tauxe, 2010).
$575^\circ C$) and high-coercivity hematite (Curie temperature = $675^\circ C$) to be the main magnetic carrier in the tested samples (Figure 2.12). FORC measurements (Figure 2.13) and the detection of a very viscous magnetic component at room temperature (Figure 2.14) support a SP and non-interacting SD habit of the magnetic minerals. The time of magnetic relaxation is dependent on the size of particles (Figure 2.16). The inability of capturing higher temperature magnetic components in the samples section is due to magnetic grain sizes below 20 nm, which has been interpreted as threshold grain-size for superparamagnetic particles (Tauxe et al., 1996).

### 2.6 Conclusion

Magnetostratigraphy of a growth structure in Miocene sediments of the Coastal Makran did not reveal useful ChRM’s. Therefore, geomagnetic polarities could not be identified. A storage test showed that the NRM is overprinted by the Earth’s magnetic field within the range of months due to a viscous component. The very low coercivities and a three-component IRM thermal-demagnetization test indicated magnetite to be the main carrier with a minor hematite contribution. FORC measurements show that the grain-size is in the range of SP and non-interacting SD. The size of magnetic particles in the studied section is the main reason for the lack of ChRM’s, due to a very low magnetic relaxation time, which is particle size dependent.
Abstract

The temporal evolution of deformation in the Zagros Simply Folded Belt is constrained by a magnetostratigraphic sequence containing a progressive unconformity located on the southern limb of the Kuh-e Ghol Ghol anticline in the Central Fars. The investigated ~1400 m thick sequence exposes a regressive mega-cycle containing, from bottom to top, open and shallow marine marls and sandy limestones, fine- to coarse-grained fluvial deposits and alluvial conglomerates. Correlating the magnetostratigraphic section with the geomagnetic polarity time scale constrains the transition from marine to fluvial sediment deposition at ~6 Ma. This transition was accompanied by a change in accumulation rate from ~15 cm/ka to ~40 cm/ka, as measured on lithified sediments. Alluvial river deposits first occurred at 3.2–3.1 Ma. The the Kuh-e Ghol Ghol anticline began to grow at 3.7–3.5 Ma. A comparison of magnetostratigraphic sections and ages of growth strata in NE Fars indicates a ~1 cm/a migration rate of the deformation front towards the SW during the Middle and Late Miocene.
3.1 Introduction

The Zagros Mountains result from post-collision convergence between Arabia and Eurasia. It extends roughly 2000 km from the Makran subduction zone in the SE to the East Anatolian fault in the NW (Mouthereau et al., 2012). Collision in the Zagros began most probably in Late Eocene to Oligocene times, after subduction of the Tethys Ocean (Agard et al., 2005; Jolivet and Faccenna, 2000; Mouthereau et al., 2012), although different models have been suggested with collision initiation ranging from Late Cretaceous to Upper Pliocene (Berberian and Berberian, 1981; Berberian and King, 1981; Stocklin, 1968).

The Zagros Mountains are subdivided into tectono-stratigraphic zones. From SW to NE, these are the Zagros foreland basin, the Simply Folded Belt, the Imbricate zone (or High Zagros), the Sanandaj-Sirdjan zone and, in a broad sense, the Urumieh-Dokhtar magmatic arc (Figure 3.1; Alavi, 1994; Falcon, 1974; Stocklin, 1968). The boundary between the Arabian and the Eurasian tectonic plates is defined by the north-eastward dipping Main Zagros Thrust (Paul et al., 2010; Stocklin, 1968), separating the Zagros fold-and-thrust belt (Simply Folded Belt and Imbricate zone) from the internal Zagros (Sanandaj-Sirdjan zone and Urumieh-Dokhtar arc; Figure 3.1). The present study is concerned with the Zagros Simply Folded Belt, i.e. structures to the SW of the High Zagros Fault, the boundary to the Imbricated zone (Figure 3.1).

Along strike, the Simply Folded Belt is divided into the Pusht-e Kuh arc (NW Zagros, Lurestan) and the Fars arc (central Zagros, Fars), geographically separated by the Izeh zone and the Dezful embayment (Figure 3.1; Allen and Talebian, 2011; Lacombe et al., 2006; Sherkati et al., 2006). Horizontal shortening of an initially 10 – 12 km thick sedimentary sequence formed hundreds of kilometres long folds (Alavi, 2007; Colmansadd, 1978; Motiei, 1993). The regular fold wavelength (~ 15 km; Mouthereau et al., 2007b) is attributed to thin-skinned deformation. In the Fars, the major basal décollement is placed in the Precambrian Hormuz salt (Bahrudi and Koyi, 2003; Edgell, 1996). In the Lurestan province, salt is replaced by Cambrian shales (Sherkati and Letouzey, 2004). Additional, rheologically weak horizons within the sedimentary sequence (Fard et al., 2011; Motiei, 1993; Sherkati et al., 2005) are the controlling factor for folding to be the dominating deformation mode (Ruh et al., 2012; Yamato et al., 2011). Even though the structural deformation style of the Simply Folded Belt can generally be interpreted as thin-skinned tectonics, large-wavelength (40 – 100 km) topographic steps may root in
active reverse faults rooted in the basement (Leturmy et al., 2010; Mouthereau et al., 2006). Seismotectonic studies support basement-involved faulting (Talebian and Jackson, 2004). Whether deformation of the Simply Folded Belt evolved from thin-skinned to thick-skinned in the Miocene/Pliocene (Molinaro et al., 2005) or was basement-involved from the beginning (Mouthereau et al., 2007b) is still disputed.

A shortening amount of $70 - 110$ km has been estimated across the Zagros Mountains from initial collision to recent times (Agard et al., 2011; McQuarrie, 2004; Mouthereau et al., 2007b). The spatial and temporal distribution of deformation in the Simply Folded Belt.
Belt is still debated. Traditionally, a major unconformity between alluvial conglomerates (Bakhtyari formation) above fluvial sandstones and shales (Agha Jari formation) dated Pliocene/Pleistocene was inferred to denote the time of folding (Falcon, 1974; James and Wynd, 1965). Several authors have dated deformation initiation close to the Eocene/Oligocene transition from angular unconformities to the SW of the High Zagros Fault (Hessami et al., 2001; Khadivi et al., 2010; Saura et al., 2011). Further investigation of unconformities throughout revealed that deformation migrated in-sequence, from the NE to the SW (Hessami et al., 2001). Flat-lying Plio-/Pleistocene conglomerates presently spread over the entire Simply Folded Belt are related to a phase of fold-tightening at $\sim 2 - 3$ Ma (Mouthereau et al., 2007b).

An efficient mean to date deformation in fold-and-thrust belts is the study of progressive unconformities, i.e. growth strata (Riba, 1976; Verges et al., 2002). A suitable tool to accurately derive age and sedimentation rate is magnetostratigraphy (e.g., Burbank and Raynolds, 1988; Schlunegger et al., 1997). Several attempts have been made to date growth strata with this technique in Zagros (Figure 3.1). Homke et al. (2004) investigated a $\sim 3$ km thick section at the front of the Pusht-e Kuh arc, including growth strata and the transition from fine-grained fluvial to conglomerate alluvial deposits. They dated the initiation of the Mountain Front Flexure at $\sim 8.1 - 7.2$ Ma, with subsequent folding lasting for at least 5 Ma. Emami (2008) provided age constraints for folding initiation at 11.8 Ma in the Pusht-e Kuh and $\sim 11$ Ma in the Izeh zone, close to the High Zagros Fault. Khadivi et al. (2010) found that folding initiated no later than 14.8 Ma in the NW of the Fars arc and south of the High Zagros Fault. All magnetostratigraphic sections are close to the geographic boundaries of the Simply Folded Belt, and most of them are located in the Lurestan region (Figure 3.1).

In this study, magnetostratigraphy has been applied on an Upper-Neogene sequence in the central Fars (Figure 3.1). Two sections (Figure 3.2) include a progressive unconformity at their top, which enables constraining the timing of deformation in the central Fars area.

### 3.1.1 Stratigraphy of the Simply Folded Belt in the central Fars

The stratigraphic succession of the Simply Folded Belt lays over the Cambrian Hormuz salt, which acts as the major décollement, and comprises the Palaeozoic and Mesozoic deposits on the Arabian passive margin. The Cenozoic stratigraphy includes several transgressive and regressive cycles that include pelagic marl, shales and limestone ac-
cumulations interlayered with shallow marine limestones and massive dolomites (James and Wynd, 1965; Motiei, 1993). The upper part of the Cenozoic stratigraphy represents a regressive cycle beginning with Early Miocene, interlayered clay and calcareous sandstones (Rasak formation) unconformably covering limestones in the NE Fars (Khadivi et al., 2010). In the central Fars, these clastic sediments are replaced by open shallow marine carbonates (Gachsaran formation; Pirouz et al., 2011). Up-sequence, green marls and thin interlayered limestone and oyster beds are dominating (Mishan formation). Massive reef limestones are present at the bottom of the Mishan formation (Guri member; Pirouz et al., 2011). The limestones are succeeded by marls, red sand- and mudstones and few pebbly sandstone to conglomeratic layers (Agha Jari formation). These marine delta and fluvial river deposits are the erosional products of the Zagros Mountains (Motiei, 1993; Pirouz et al., 2011). The younger deposit consists of massive conglomerates and cross-bedded gravels (Bakhtyari formation) that accumulated in an alluvial system (Motiei, 1993), whose age varies from ~ 15 Ma close to the High Zagros (Fakhari et al., 2008; Khadivi et al., 2010) to Late Pliocene in the outer Fars (Oveisi et al., 2009). This large time variation expresses migration of the deformation front from the NE to SW (Hessami et al., 2001).
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3.2 Methods

3.2.1 Sampling

Two sections that cover the Neogene deposits, \(\sim 40\) km NW of the city of Khonj in central Fars (Figure 3.2), have been drilled for magnetostratigraphic sampling using a gasoline-powered drill. Both sections cut the south-western limb of a large anticline (Kuh-e Ghol Ghol; Figure 3.2). Section 1 cuts from bottom to top: (i) the upper part of the marine Mishan formation (Guri member is not sampled), including thick marl beds, thin (\(\sim 2\) m thick) limestone beds and several oyster layers (Figure 3.3a); (ii) the marine deltaic to fluvial Agha Jari formation with red and grey marls, cross-bedded fine-grained and often pebbly sandstones (Figure 3.3b) and interlayered conglomerate layers (Figure 3.3d); and (iii), the massive alluvial conglomerates of the Bakhtyari formation. The stratigraphic contact between the deltaic and the overlying alluvial formation is continuous, i.e. no major angular unconformity has been observed. Nevertheless, a second arrival of alluvial conglomerates makes and angular unconformity within the alluvial Bakhtyari formation. Several authors separated the conglomerates as Bakhtyari 1 and Bakhtyari 2 formations (e.g., Khadivi et al., 2010; Mouthereau et al., 2007b). Section 1 consists of 104 sample sites over a stratigraphic thickness of 1413 m, with an average sample resolution of 14 m. This sample resolution is comparable to sampling resolution used in similar studies of the same lithologies (Homke et al., 2004; Khadivi et al., 2010).

Section 2 is located \(\sim 6\) km NW of section 1 and cuts the transition from the reddish sand-dominated (deltaic Agha Jari) to the purely conglomerate (alluvial Bakhtyari) formation (Figure 3.2). This section has been drilled along a new (in January 2012) road cut that exhibits unconformities within the Agha Jari formation, with onlapping and progressive flattening towards younger layers, which are growth strata (Figure 3.3e). The contact with the overlying (Bakhtyari) formation is identified by clast lithologies of the conglomerates. The Agha Jari conglomerates contain almost exclusively sub-rounded to rounded limestone clasts < 30 cm in diameter (Figure 3.3d). The Bakhtyari conglomerates are more colorful with clasts of various rocks originating from NE areas of the orogenic belt. Section 2 includes 22 drill sites over a stratigraphic thickness of 189 m with a sample resolution of 9 m.

In both sections, sampling has focussed on fine-grained lithologies such as mud-, silt- and fine-grained sandstones. Where such lithologies were lacking, shelly sandstones, sandy
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Figure 3.3: Photographs of typical lithologies of the studied sections. a) Oyster layer in the Mishan formation. b) Reddish channelized pebbly sandstones of the Agha Jari formation. c) Typical conglomerate from the Bakhtyari formation, with clasts from different lithologies. d) Interbedded sandstones and conglomerate layers with limestone clasts in the Agha Jari formation. e) Onlap and progressive unconformity within the upper Agha Jari and Bakhtyari formations. a) - d) are taken along Section 1, e) along Section 2.

Limestones (Mishan formation) and coarse-grained sandstones (Agha Jari and Bakhtyari formations) were drilled. Six marl and shale samples have been collected along Section 1 for independent calcareous nanoplankton and foraminifera dating.
3.2.2 Magnetic measurements

The natural remanent magnetization (NRM) of every sample was measured before the samples were stepwise demagnetized. The NRM and the magnetization after every demagnetization step was measured with a 2G Enterprise 755R, 3-axis DC-SQUID rock magnetometer (sensitivity of $\sim 3 \cdot 10^{-8}$ A/m) at the Laboratory for Natural Magnetism of the ETH in Zurich. All samples have been demagnetized thermally in 13 steps up to 690°C. Sample heating was conducted with a Schonstedt TSD-1 and an ASC TD48 oven with an internal field < 5 nT. Alternating field demagnetization has been conducted on 10 samples, using an AF demagnetization unit that is part of the cryogenic magnetometer. Demagnetization behaviours were plotted on vector diagrams to identify different magnetic components that unblock at different temperatures (Zijderveld, 1967), and components of magnetization were isolated using Remasoft software (Chadima and Hrouda, 2006). For further magnetic investigation, several additional methods have been applied. Magnetic susceptibility of every sample was measured with an AGICO KLY-2 Kappabridge before demagnetization. Acquisition of isothermal remanent magnetization (IRM) has been performed on 14 representative samples from both sections and all lithologies. The IRM was applied with an ASC Impulse Magnetizer (Model 10-30), and measured on the cryogenic magnetometer. Afterwards samples were given a cross-component magnetization and thermally demagnetized to aid in identification of the ferromagnetic minerals in the samples (Lowrie, 1990). First, different coercivity fractions of IRM were induced along three orthogonal directions in successively smaller fields: a field of 1.2 T was applied in the $z$-axis of the sample, a field of 0.4 T in the $y$-axis and 0.12 T in the $x$-axis. Then, samples were step wise thermally demagnetized with an ASC TD48 oven. Hysteresis parameters such as saturation magnetization ($M_s$), saturation remanence ($M_r$), coercivity ($H_c$) and coercivity of remanence ($H_{cr}$) were measured with a MicroMag 2900 Alternating Force Gradient Magnetometer.

3.3 Results

3.3.1 Paleomagnetic analysis

The intensity of the NRM is larger than $10^{-6}$ A/m for all measured samples (Figure 3.4a). Shelly sandstone and sandy limestone samples (marine Mishan formation) show relatively low intensities in contrast to the overlying clastic sediment samples. In Section
3.3. RESULTS

Figure 3.4: a) Intensity of the natural remanent magnetization, b) bulk susceptibility and c) intensity corrected by susceptibility of sections 1 and 2. Obtained curves match formation contacts observed in the field.

1, intensities decrease slightly upward after the marine/deltaic transition (Figure 3.4a). This trend in intensity is not recognizable in Section 2. Magnetic susceptibility measurements show a trend similar in intensity across both sections with rather low values ($10^{-6} - 10^{-5}$) in the calcareous rocks of the Mishan formation and values between $10^{-4} - 10^{-3}$ in the clastics of the Agha Jari and Bakhtyari formations (Figure 3.4b). The range of susceptibility values agrees with earlier studies in the area (Bakhtari et al., 1998). The ratio of intensity divided by the dimensionless susceptibility (Figure 3.4c) provides a first-order indicator for variations in the magnetic mineralogy. The ratio remains relatively stagnant, indicating similar magnetic mineralogy along both sections.

IRM acquisition curves show that all samples contain both a low and high coercivity component of magnetization (Figure 3.5a). These results support the interpretation of similar magnetic mineralogy throughout the two sections. Coercivity of remanence ($H_{cr}$) is generally around 30 – 40 mT for clastic formations, but can be higher in the
Figure 3.5: Magnetic properties of randomly chosen samples. a) IRM acquisition curves revealing largest remanence coercivities for samples from the Mishan formation. b) Hysteresis plot of sample 12-Z-170 from the Mishan formation indicating paramagnetic behaviour. c) Hysteresis plot of sample 12-Z-184 from the Agha Jari formation indicating pseudo-single domain behaviour combined with a paramagnetic component. d) Hysteresis ratios showing pseudo-single domain for Agha Jari and Bakhtyari sites.

marine, carbonate-dominated Mishan formation. Results of induced magnetization measurements reveal that magnetic particles in Mishan samples exhibit an almost pure paramagnetic behaviour (Figure 3.5b). Samples of the deltaic and alluvial Agha Jari and Bakhtyari formations display hysteresis (Figure 3.5c). After removal of the paramagnetic slope, the ratios of saturation magnetization ($M_s$) to remanent saturation magnetization ($M_r$) and coercivity of remanence ($H_{cr}$) to coercivity ($H_c$) are consistent in the field of pseudo-single domain magnetite on a Day-Dunlop plot (Figure 3.5d; Day et al., 1977; Dunlop, 2002a,b). Coercivity ratios are higher than expected, but this is due to the presence of hematite in the samples.
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Thermal demagnetization curves of the soft (< 0.12 T), medium (0.12 – 0.4 T) and hard (0.4 – 1.2 T) coercivity fractions of the three-component IRM allow to extract different magnetic carrier in tested samples (Figure 3.6). A marine calcareous sandstone of the Mishan formation (12-Z-154) has a large soft (low-coercivity) fraction that unblocks roughly linear with increasing temperature by 600°C (Figure 3.6a). This indicates that the main magnetic carrier in this sample is magnetite. The loss of medium fraction around 200°C could represent maghemite or titanomagnetite. The demagnetization curve of the hard fraction clearly supports the occurrence of hematite with a large coercivity and a high Curie temperature. The sandstone samples are more difficult to interpret. A coarse sandstone sample of the deltaic environment (12-Z-185) also shows a large soft fraction indicating magnetite as the main magnetic carrier (Figure 3.6b). The remaining soft fraction at 600°C that unblocks by 700°C points to low-coercivity hematite, probably due to large grain size. Hematite is further indicated by the hard fraction unblocking at high temperatures. A drop of IRM in the medium fraction is observed at 200 – 250°C, probably according to minor maghemite or titanomagnetite (Figure 3.6b). A sample from a fine sandstone channel in the Bakhtyari conglomerates (12-Z-244) shows a similar pattern as sample 12-Z-185. The major magnetic carrier is magnetite with low- to high-coercivity hematite (Figure 3.6c). A minor amount of maghemite or titanomagnetite is shown by the medium demagnetization curve.

AF demagnetization was not suitable in isolating stable components of magnetization. Thermal demagnetization of the NRM reveals a low-temperature magnetic component in almost all samples, which is removed by 200 – 250°C (Figure 3.7). Most samples lose more than 50% of their initial NRM by 250°C. A single, stable component of magnetization is removed for 77% of the samples, by further progressive heating up to 690°C (Figure 3.7a,b,d,f,h). 23% of the measured samples revealed an additional magnetic component, located mostly between 250 and 500°C (Figure 3.7c,e,g,i). In these samples, the characteristic remanent magnetization ChRM was defined to be the component that was unblocked at highest temperatures, after the intermediate component. Nonetheless, in some cases, the intermediate component was unblocked only at 600°C (Figure 3.7g). In shelly sandstones and sandy limestones of the marine Mishan formation, where magnetic intensities are relatively low (Figure 3.4a), samples were demagnetized at 500 – 550°C (Figure 3.7a,b). In the deltaic Agha Jari and alluvial Bakhtyari formations, complete demagnetization occurs often above 600 or 640°C. This indicates the presence of hematite, whereas magnetite is the main magnetic carrier in the carbonate bearing rocks of the
Figure 3.6: Thermal demagnetization of the soft (< 0.12 T), medium (0.12 - 0.4 T) and hard (0.4 - 1.2 T) coercivity fractions of three-component IRM. a) Calcareous sandstone sample from marine environment (Mishan formation). b) Coarse deltaic sandstone (Agha Jari formation). c) Fine, light sandstone deposited in an alluvial system (Bakhtyari formation). All samples are located in Section 1.
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Figure 3.7: Representative diagrams of tectonically NRM thermal demagnetization curves (tilt corrected). White and black points represent vertical and horizontal projections after Zijderveld (1967). Subfigures a) to f) are representing Section 1. g) to i) Section 2.

Mishan formation. ChRM has both normal and reverse polarity in all sampled formations.

A clear low-temperature component has been found in 116 samples (Figure 3.8). Statistical grouping is better if directions are not corrected for bedding tilt ($k = 10.2$, $\alpha_{95} = 4.3^\circ$) than if they are ($k = 7.1$, $\alpha_{95} = 5.3^\circ$). Although not statistically significant, the grouping in geographical coordinates indicates that the low-temperature component
developed after folding. The mean direction with declination = 353.6° and inclination = 48.7° is similar to the present Earth’s field direction (Dec = 2.6°; Inc = 44.0°).

The ChRM direction of each sample was used to calculate the latitude of the Virtual Geomagnetic Pole (VGP), where positive VGP latitude represents normal polarity and negative VGP reverse polarity. On Section 1, 43 drill sites show a normal and 61 a reverse polarity. Sites in Section 2 have 11 normal and 11 reverse directions. All samples can be divided into three types based on the quality of the vector fit, and location of the VGP. Type 1 demagnetization curves (48% of all samples) refer to ChRMs that were fitted with a component that demagnetized with a clear normal or reverse polarity. Type 2 patterns (38% of all samples) show a clear polarity, but demagnetization curves are imperfectly linear. Type 3 samples (remaining 14% of all samples) show a poor quality in demagnetization, yet yielding a useful direction of the ChRM. Normal and reverse directions after correction for bedding orientation (rotation along strike) are projected on a stereographic net (Figure 3.9). Both normal (Figure 3.9a) and reverse polarities (Figure 3.9b) have mean orientations comparable to the Earth’s magnetic field in Miocene times in this area (Smith et al., 2005).
Figure 3.9: Equal area (Wulff) stereographic projection of the characteristic remanent magnetization (126 samples) after tectonic correction. a) Normal polarity directions derived from VGP (57 samples). Diamond and circle are mean direction and 95% precision. b) Reverse polarity directions derived from VGP (69 samples). Diamond and circle like in (a). (Dec: declination, Inc: inclination, k: precision parameter, $\alpha_{95}$: confidence angle).
3.3.2 Stratigraphic ages

Six marl and shale samples have been collected to obtain stratigraphic fossil ages. Calcareous nanofossils did not yield any firm age constraint. Nanofossils are very rare in all samples, depleted by detritic material and probably destroyed during diagenesis. Only one sample (13-Z-S2) could be ascribed to the lower Serravallian (lower NN6). However, this age is taken with caution due to the bad preservation of the fossils (written and oral communication with Carla Mueller). The investigation of benthic and planktonic foraminifera indicate a shallow water environment (< 100 m) with high nutrient availability, but with no age constraint for samples 13-Z-S1 to S3, which belong to the marine Mishan formation. Sample 13-Z-S5 contains Asterorotalia venusta and Asterorotalia schroeteriana, which both indicate a Quaternary age (written and oral communication with Silvia Spezzaferri). No erosional interface or hiatus has been observed in the ∼ 300 m stratigraphic thickness between 13-Z-S2 and 13-Z-S5. Therefore, it is highly implausible that both fossil ages represent true deposition ages of their host sediment (10 Ma of time difference within 300 m of stratigraphy).

3.4 Correlation to the Geomagnetic Polarity Timescale (GPTS)

The relative fossil ages are uncertain. There are three possibilities to use these ages, having in mind that samples 13-Z-S2 and 13-Z-S5 cannot be included into the correlation to the GPTS (Cande and Kent, 1992, 1995). These possibilities are: i) the Serravallian age from calcareous nanofossils of sample 13-Z-S2 is correct, ii) the Quaternary age of foraminifera in 13-Z-S5 is correct, or iii) none is correct. For this reason, an independent correlation of the magnetostratigraphy is made to the GPTS. The optimum solution was achieved by a simple fit of the normal and reverse polarity intervals to the GPTS, ignoring the fossil ages (Figure 3.10). Samples from the Mishan formation at the bottom of Section 1 show mainly normal polarities. A strong majority of samples from the overlying Agha Jari formation exhibit a reverse polarity with short normal intervals (Figure 3.10). Towards the transition to the Bakhtyari formation, a normal polarity zone is dominant (Section 1; 1150 m). Section 2 contains a short normal period at the bottom, followed by a reverse and again a normal polarity interval. Three reverse polarized samples top the section (Figure 3.10). The contact between the deltaic Agha
Jari and the alluvial Bakhtyari formation is located in the upper normal interval in both sections, which provides a stratigraphic tie point.

The fit to the GPTS has been made with Section 1. Section 2 was then pinned to Section 1 at the stratigraphic contact between the Agha Jari and Bakhtyari formations. A best fit was achieved by pinning the bottom normal interval of Section 1 to the combined chrons C4n and C3B (Figure 3.10). A potentially lower sedimentation rate reduces the stratigraphic resolution and increases the possibility of missing magnetic intervals.

Figure 3.10: Correlation of the obtained magnetic polarity curve to the Geomagnetic Polarity Timescale (GPTS). Fine dashed lines indicate pinning points, thick dashed lines represent transitions of sedimentological formations observed in the field. Accumulation rates (in cm/ka) are achieved by connecting pin points. Gray area indicates growth of strata.
Above this reverse chron C3Ar is only represented by a single (but type 1 quality) pole at 170 m in Section 1 (Figure 3.10). The two successive normal intervals both belong to chron C3An (Figure 3.10). The stratigraphic contact between the Mishan and the Agha Jari formation is located in the top normal interval of this chron. The lower 200 m of Agha Jari formation represent a reverse interval, which is connected to chron C3r. A succession of several short normal intervals ending at 1050 m in Section 1 shows the complete chron C3n (Figure 3.10). Chron C2Ar is linked to the top reverse interval of the Agha Jari formation. The two normal polarity intervals at the top of Section 1 are interpreted to be the lower part of chron C2An (Figure 3.10). The contact between the Agha Jari and Bakhtyari formation is therefore in the bottom normal interval of chron C2An. Linking the stratigraphic contact to Section 2 suggests that the lower reverse interval of Section 2 represents chron C2Ar and the top normal interval (Figure 3.10).

An alternative fit to the GPTS can be made by fixing sample 13-Z-S2 to chron C5AA, representing the lower NN6 nanostratigraphic zone, which is lower Serravallian. In this case, the top normal polarity phase has to be connected to chron C5n, which starts at \( \sim 11 \) Ma. This would yield improbably high deposition rates (up to 50 cm/ka) for the partly marine Mishan formation. Furthermore, the visual fit of chrons to measured polarity intervals would be worse if nanofossils are taken into consideration. Pinning the site of sample 13-Z-S5 to the Quarternary is also not feasible because there are 1000 m of overlying stratigraphy and the top sediments are tilted \( \sim 30 - 40^\circ \). The contact from marine Mishan to deltaic Agha Jari formation would then link to chron C2n at \( \sim 2 \) Ma.

### 3.5 Discussion

#### 3.5.1 Quality of paleomagnetic and fossil data

Magnetite is the main carrier of magnetic components throughout the studied area. The carbonates of the Mishan formation contain high-coercivity hematite. Clastic lithologies show the occurrence of a minor fraction of low- to high-coercivity hematite (Figure 3.6). Whether the medium-coercivity fraction demagnetization at low temperatures represents maghemite, a form of oxidized magnetite, or titanomagnetite, potentially originating from the Internal Zagros, is not answered here.
85% of the measured samples showed demagnetization patterns of quality type 1 and 2 with a clear ChRM converging towards the origin (Figure 3.7). The resulting inclination and declination of the ChRM does not necessarily coincide with the Earth’s magnetic field orientation at the time of sediment deposition. Nevertheless, the clustering of normal and reverse polarities of the ChRM yield mean orientations similar to Earth’s field orientations (Figure 3.9). Statistical values for the clustering of ChRM \( k \approx 4; \alpha_{95} \approx 11^\circ \) and the percentage of good thermal demagnetization data (type 1 and 2) is comparable to values of similar studies in the Zagros (Homke et al., 2004; Khadivi et al., 2010).

The calcareous nanofossils and planktonic and benthic foraminifera did not provide indisputable ages. Compared to the good quality of the paleomagnetic data, the fossil ages were disregarded during construction of the polarity curve (Figure 3.10).

### 3.5.2 Time constraints of the Agha Jari and Bahktyari formations

Correlating the magnetostratigraphic section with the GPTS constrains the time of sedimentation. The contact between the marine Mishan and the deltaic Agha Jari formation is observable in the field and distinguishable by a change in NRM intensity and susceptibility. The increase of NRM intensity and susceptibility from sample 12-Z-171 upwards (Figure 3.4) reflects the change towards the fluvial deposits of the Agha Jari formation. The contact is located between sample 12-Z-170 and 171 at 363 m within the profile of Section 1, which ties the bottom of the Agha Jari formation to the upper normal interval of chron C3An, between 6.14 and 5.89 Ma. This stratigraphic contact also marks a change in sedimentation rate (Figure 3.10). The marine Mishan sediments were deposited at an averaged (compacted) sedimentation rate of 16.4 cm/ka, which is consistent with rates calculated for the comparable Rasak formation in the NE Fars by Khadivi et al. (2010; 15.1 – 21.3 cm/ka). The sedimentation rate (after compaction) increases slightly upwards towards the initiation of fluvial sediment input. The Agha Jari formation was deposited with a mean rate of 32.4 cm/ka. This rate is comparable to rates of Agha Jari lithologies in the NW Zagros (Homke et al., 2004). The boundary between the Agha Jari and the Bakhtyari formation is clear in the field. There is little constraint from intensity and susceptibility data (Figure 3.4). It lays between sample site 12-Z-234 and 235 in Section 1 and between 12-Z-260 and 261 in Section 2. Both sections exhibit a normal polarity interval at this stratigraphic level (Figure 3.10). In section 1, the Agha Jari - Bakhtyari contact lays at 1250 m from bottom of the sampled profile,
which indicates a thickness of \(\sim 900\) m for the Agha Jari formation in the studied section (Figure 3.10). The normal polarity interval containing the contact can be linked to the intermediate normal interval of chron C2An, which covers an interval between 3.22−3.11 Ma. The alluvial Bakhtyari formation atop was deposited with a mean accumulation rate of 26 cm/ka. An alternative correlation of the top of Section 1 to the GPTS can be obtained due to the possibility of very large accumulation rates in alluvial systems (e.g., Khadivi et al., 2010). If sedimentation rates were as large as 52.5 cm/ka, the Agha Jari - Bakhtyari contact could be placed within the lower normal polarity interval of chron C2An between 3.58−3.33 Ma (Figure 3.10). In Section 2, the normal polarity interval hosting the Agha Jari - Bakhtyari contact is also pinned to the intermediate interval of chron C2An. This results in lower accumulation rate of approximately 13 cm/ka. This low rate can be explained by the stratigraphic onlap within the Agha Jari formation before the bedding dips progressively decrease towards the section top (Figure 3.3e). Around this unconformity, the calculated sedimentation rates could be underestimated due to erosion of stratigraphic layers.

The stratigraphic transition from marine (Mishan) to fluvial, partly continental (Agha Jari) deposits took place at \(\sim 6\) Ma in the studied Section 1. In the Fars, the Mishan lithologies are geographically more spread than in the Dezful and the Push-e Kuh arc (Figure 3.1; Pirouz et al., 2011), where the shallow open marine marls, shelly sandstones and limestones of the Mishan formation are replaced by clay, sandstones and conglomerates (Rasak) and supratidal and sabkha deposits (Gachsaran). This implies that in the Middle Miocene, the NW Zagros was a shallow water evaporitic basin, while a shallow to open marine system was present in the Fars. Throughout the Zagros, the Mishan formation is thought to start in the Lower to Middle Miocene (Alavi, 2004; Hessami et al., 2001; James and Wynd, 1965). Stratigraphic columns of drill logs from the Coastal Fars show that the Mishan formation can be more than 1600 m thick (Pirouz et al., 2011). Assuming a similar thickness below its upper stratigraphic boundary with the Agha Jari formation in Section 1 and extrapolating the calculated accumulation rates, an upper boundary at around 6 Ma is plausible. The fact that the Simply Folded Belt and its deformation front migrated towards the SW during orogenic growth entails that timing of facies deposition in the Zagros is diachronous (Pirouz et al., 2011). The transition from Agha Jari to Bakhtyari formation is \(\sim 3.2\) Ma old. This age is comparable to Late Pliocene or younger ages obtained in earlier and recent studies for the alluvial Bakhtyari formation (Falcon, 1974; Homke et al., 2004; James and Wynd, 1965; Motiei, 1993; Oveisi et al., 2009). Published ages based on nano- and microfossils for the Mishan, Agha Jari
and Bakhtyari formations along the sampled sections agree very well with the obtained results of the present work (Yousefi, 2005).

3.5.3 Folding of the Zagros foreland

The onlap of strata followed by a progressive unconformity along Section 2 (Figure 3.3e) allows constraining: i) the time of folding of the Kuh-e Ghol Ghol anticline and ii) the deformation distribution in the Simply Folded Belt. According to the magnetostratigraphic sequence, the base of the growth strata is at 3.7 – 3.5 Ma (Figure 3.10). The erosion surface marking incipient fold growth may be interpreted as the change from an accelerated to a reduced diastrophism, indicating high tectonic activity (Figure 1.11). Furthermore, the erosion and removal of strata at the base of the progressive unconformity could explain the low accumulation rates calculated on Section 2, compared to rates in Section 1, where no erosional contact has been observed. In order to obtain a similar sedimentation rate for the Bakhtyari formation in both sections, erosion of about 50 m is required in Section 2. This does not influence the dating of the initiation of folding as the footwall of the erosional contact is well constrained.

This work is the first magnetostratigraphic study dating deformation in the Outer Fars of the Zagros Folded Belt. It is generally accepted that the Simply Folded Belt developed by in-sequence folding. A combined magnetostratigraphic section that incorporates all growth strata along the Simply Folded Belt allows clarifying the temporal evolution of the thin-skinned folding in Zagros (Figure 3.11).

The oldest progressive unconformity within Neogene stratigraphy is reported by Khadivi et al. (2010) close to the High Zagros Fault, in NE Fars (Figure 3.1). There, initiation of deformation is contemporaneous with the beginning of the accumulation of braided river deposits of the Bakhtyari formation between 15 – 14 Ma (Figure 3.11). The temporal difference in deformation initiation from the NE to the Central Fars (3.7 – 3.5 Ma) is therefore > 10 Ma for a distance of > 100 km. This yields a ~ 1 cm/a frontal migration of the Folded Belt. A slightly faster (1.7 cm/a) advance of the deformation front has been inferred for the Quaternary in the SE Zagros (Mann and Vita-Finzi, 1988).

Magnetostratigraphy of growth strata close to the High Zagros Fault revealed deformation ages of 11 Ma in the Izeh and 11.8 Ma in the Pusht-e Kuh arc (Figure 3.1; Emami, 2008), indicating initiation of folding later than in the Fars arc (Figure 3.11). A progressive unconformity along the Mountain Front Flexure (MFF) in the Pusht-e Kuh arc
dates the activity of the MFF at $8.1 - 7.2$ Ma (Figure 3.1; Homke et al., 2004). The age difference of folding is $\sim 4$ Ma (Figure 3.11) over a distance of $\sim 100$ km across the Pusht-e Kuh arc (Figure 3.1). If deformation propagated in-sequence in the Pusht-e Kuh arc, a migration rate of $\sim 2.5$ cm/a is necessary.

### 3.6 Conclusion

Magnetostratigraphy of a $\sim 1400$ thick sequence including growth strata in the Central Fars was used to constrain the temporal evolution of stratigraphy and timing of folding. According to magnetic polarity correlation to the GPTS, the transition from a mainly marine to fluvial sedimentation occurred at $\sim 6$ Ma. Alluvial braided river deposits commenced at $3.2 - 3.1$ Ma. Accumulation rates (calculated from compacted sediments) of $13 - 15$ cm/ka during marine deposition increased up to $42$ cm/ka in fluvial and alluvial systems during the Pliocene. Growth strata, and therefore the beginning of folding in the Fars, are dated at $3.7 - 3.5$ Ma; this is $\sim 10$ Ma later than along the High Zagros Fault. This result supports an in-sequence evolution instead of contemporaneous,
homogeneous folding throughout the Simply Folded Belt. In the Upper Miocene the deformation front migrated from NE to SW with a rate of $\sim 1 \text{ cm/a}$. A composite section of magnetostratigraphic studies in the Zagros Mountains shows that shortening and subsequent deformation of the Simply Folded Belt took place over $> 10 \text{ Ma}$.

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Numerical investigation of deformation mechanics in fold-and-thrust belts: Influence of rheology of single and multiple décollement

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Abstract

Thin-skinned fold-and-thrust belts related to convergence tectonics develop by scraping off a rock sequence along a weaker basal décollement often formed by water-saturated shale layers or low-viscosity salt horizons. A two-dimensional finite element model with a visco-elasto-plastic rheology is used to investigate the structural evolution of fold-and-thrust belts overlying different types of décollements. In addition, the influence of multiple weak layers in the stratigraphic column is studied. Model shale décollements are frictional, with lower friction angles as the cover sequence. Model salt layers behave linear viscous, applying a lower viscosity as the cover sequence, or with a power-law rheology. Single viscous décollement simulations have been compared to an analytical solution concerning faulting versus folding. Results show that fold-and-thrust belts with a single frictional basal décollement generate thrust-systems ramping from the décollement to the surface. Spacing between thrust ramps depends on the thickness of the cover sequence. The structural evolution of simulations with an additional low-frictional layer depends on the strength relationship between the basal and the inter-sequential décollement. Tectonic underplating and antiformal stacking occur if the within-sequence décollement is weaker. In the frontal part of models, deformation is restricted to the upper part and imbrication occurs with a wavelength depending on the depth of the intermediate weak layer. “Salt” décollement with a viscosity of $10^{18}$ Pa·s leads to isolated box-folds (detachment-folds). Multiple salt layers ($10^{18}$ Pa·s) result in long-wavelength folding. Our results for both frictional and viscous décollements are in bulk agreement with the Mohr-Coulomb type, critical wedge theory.

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4.1 Introduction

Fold-and-thrust belts have been within the scope of intense industrial and academic interest for several decades (e.g., Boyer and Elliott, 1982; McClay, 1992). They are reported in contrasting tectonic settings such as passive margins, where they are gravity-driven, and convergent zones such as mountain forelands and submarine accretionary wedges. In this study, we focus on the thin-skinned deformation of fold-and-thrust belts related to convergent plate boundaries. They develop by scraping off crustal material along a major décollement over a relatively rigid basement/subducting plate. Typically, such décollement zones are weaker than the bulk of the deforming rock pile and often consist of a viscous salt layer or more or less water-saturated, occasionally over-pressured shale (e.g., Cooper, 2007; Morley et al., 2011). Sliding of an evolving fold-and-thrust belt on a décollement depends on the stresses acting on the weak layer and its shear resistance. Therefore, the style of thin-skinned deformation is strongly dependent on the physical properties of the weak, major décollement layers.

Submarine accretionary wedges like Barbados (Westbrook et al., 1988), the Makran (Grando and McClay, 2007; Platt et al., 1985) and the South Caspian Sea (Berberian, 1983) are typical examples for fold-and-thrust belts with shale as décollement horizons in convergent zones. Shale-related fold-and-thrust belts are also known from retro-arc settings like the Subandean ranges in Argentina (Echavarria et al., 2003). Fold-and-thrust belts with décollements within salt layers often occur in front of orogenic belts. The Jura Mountains (Laubscher, 1992; Sommaruga, 1999), the Salt Range and Potwar Plateau (Baker et al., 1988; Greland et al., 2002; McClay et al., 2004), the Pyrenees (Bourrouilh et al., 1995), the Parry Island Fold Belt (Harrison, 1995) and the Zagros Folded Belt (Mouthereau et al., 2006) are typical examples. Cross-sections illustrate the geometrical variability of these fold-and-thrust belts (Figure 4.1), which all involve thin-skinned tectonics in unmetamorphosed to low-grade metamorphic sedimentary sequences (Poblet and Lisle, 2011).

Although the large-scale mechanics of accretionary wedges and foreland fold-and-thrust belts are comparable (sediments are decoupled from and pushed over a rigid basement by a moving backstop), the difference in structural evolution and deformation style strongly depends on whether a salt or a shale layer forms the décollement level, as replicated in many analogue modelling studies. In such studies, shale décollements with a frictional boundary condition are mimicked by glass beads or a foil between the rigid basement...
Figure 4.1: Profiles of selected fold-and-thrust belts with different décollement types. a) One frictional décollement: Hikurangi subduction margin, NE New Zealand (adapted from Barnes et al. (2010)), where ramps splay up to the surface. b) Two or more frictional décollements: Makran accretionary wedge, SE Iran (adapted from Burg et al. (2013)). With duplex structures. c) Multiple shale décollements: sub-Andean thrust belt, Argentina (adapted from Echavarria et al. (2003)). Spaced ramp anticlines. d) One salt décollement: Faltenjura, NW Switzerland (adapted from Buxtorf (1916)), with narrow symmetrical anticlines and broad synclines. e) Two salt décollements: Parry Island fold belt, Arctic Canada (adapted from Harrison (1995)). f) Two or more salt décollements: southwestern simply folded zone in Zagros, SW Iran (adapted from Hessami et al. (2001)).
and the deforming sand pile. Viscosity salt layers are simulated by low-viscosity silicon or honey (Bahroudi and Koyi, 2003; Bonini, 2007; Costa and Vendeville, 2002; Storti and McClay, 1995). In numerical modelling studies, weak shale décollements are simulated by a layer with a lower Mohr-Coulomb failure criterion than the wedge-forming material (Selzer et al., 2007; Simpson, 2011; Stockmal et al., 2007). Ings and Beaumont (2010) treat shale layers as visco-plastic Bingham fluids (Bingham, 1922). Salt décollements are defined by a layer of much lower viscosity than the overlying material with linear (Simpson, 2010b; Yamato et al., 2011) or non-Newtonian rheology (Chemia et al., 2009; Li et al., 2012).

Besides the type of décollement, other important characteristics influence the structural evolution of fold-and-thrust belts. The presence of weak layers within the rock sequence affects the thin-skin deformation style, independent of whether salt or shale makes the décollement. In submarine accretionary wedges, multiple, often over-pressured shale layers are common. An example is the Makran accretionary wedge (Figure 4.1b), where mud volcanoes collect over-pressured mud horizons at shallow depth in the wedge sedimentary pile (Burg et al., 2013; Schluter et al., 2002). In the Parry Island and Zagros fold belts (Figures 4.1e and f), where the basal décollement is in evaporite, additional thin evaporite and shale layers occur within the stratigraphy (Harrison, 1995; Motiei, 1993). These could favour folding rather than faulting because intermediate décollements allow flexural slip (Yamato et al., 2011).

One goal of this study is to constrain numerically our understanding of the influence of the décollement type on the structural evolution of a compressional fold-and-thrust belt. Numerical studies of the evolution of fold-and-thrust belts already exist, but they focused either on purely viscous or purely frictional décollements, rarely comparing these two types (Simpson, 2009). Comparison is however important because the previous studies used different implementations and setups, which impedes direct assessment of the respective results. Another aim of this paper is to investigate the impact of multiple weak layers subsequently forming duplex structures within the thin-skinned stratigraphy. The role of several weak horizons has been investigated in few analogue (e.g., Konstantinovskaya and Malavieille, 2011) and numerical studies (e.g., Yamato et al., 2011). Building on these studies, ours is one of the first to implement frictional intermediate décollements (Stockmal et al., 2007) and the first to investigate the difference between single and multiple, viscous and frictional décollements using the same setup and analytic procedures. To achieve these aims, an incompressible 2D numerical code, based on the Finite Element Method with visco-elasto-plastic rheology, is employed.
Finally, we use analytical approximations, based on the critical wedge theory, to explain the overall characteristics of the model results related to both décollement types, frictional and viscous.

4.2 Mechanics of shale and salt décollements and structural characteristics of related fold-and-thrust belts

4.2.1 Frictional shale décollement

We treat shale as a purely frictional material, in contrast to a visco-plastic Bingham material, to make this study applicable to the boundary conditions of the critical wedge theory (Kopf and Brown, 2003; Saffer et al., 2001; Takahashi et al., 2007; Wang et al., 1980). This implies a frictional resistance to sliding along a potential shale décollement. The relative weakness that prompts shale to act as décollement is largely due to high pore pressures resulting from burial below impermeable layers. An increase of pore pressure reduces the yield strength of rocks (Fischer and Paterson, 1989). Consequently, the frictional resistance along such a décollement is defined by the yield strength of the over-pressured shale. The yield strength depends on the friction angle $\varphi$, the cohesion $C_0$, the stress in normal direction to the fault plane $\sigma_N$ and the fluid pressure $P_f$.

$$
\tau = (\sigma_N - P_f) \cdot \tan \varphi + C_0 \tag{4.1}
$$

To simplify equation (4.1), one can eliminate the fluid pressure term and introduce a term for the decreased friction angle of the base layer, $\varphi_b$, which is used to calculate the basal yield stress $\tau_b$.

$$
\tau_b = \sigma_N \cdot \tan \varphi_b + C_0 \tag{4.2}
$$

The mechanics of a wedge evolving over a frictional base are classically and extensively explained by the critical wedge theory (Dahlen et al., 1984; Davis et al., 1983). The evolution of a fold-and-thrust belt is compared to the evolution of a pile of sand or snow pushed in front of a bulldozer along a plane with an inclination angle $\beta$. The exact solution of the non-cohesive critical wedge equation yields the total taper ($\alpha + \beta$, where $\alpha$ is the surface angle), which is a function of the base dip angle $\beta$, the internal strength of the deforming rock pile and the strength of the basal layer (Figure 4.2). According to
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Figure 4.2: Flounder diagrams of the stability of frictional wedges with an internal friction angle $\varphi = 30^\circ$ and basal friction angles $\varphi_b$ from 5$^\circ$ to 25$^\circ$ (Davis et al., 1983) relating minimal and maximal critical surface tapers to dip angle $\beta$ of the basal décollement (every point within the flounder-shaped area is stable, points outside are unstable). Dots show minimum critical surface angles for a horizontal décollement ($\beta = 0^\circ$) for different friction angles in the basal décollement. The bottom left inset defines angles $\alpha$, $\beta$, $\psi_b$, $\psi_0$ in a critical wedge and related principal stresses $\sigma_1$, $\sigma_3$ in two dimensions.

the analytical solution, every wedge with given internal and basal strength and a defined base angle exhibits two critical total taper angles. Between these two angles, the wedge is considered to be stable. Below the minimum critical taper, thrusting thickens the rear of the wedge in order to increase the surface slope towards a stable taper. When the minimum critical taper is reached, the wedge accretes new material at its toe to sustain its stable surface angle. If the total taper exceeds the maximum critical taper, the wedge fails along normal faults to decrease the surface slope angle towards a stable attitude. For wedges with a frictional décollement, the cross-sectional taper and the internal deformation are independent of both the velocity of the basement relative to the rigid backstop and the thickness of the weak frictional layer.

The structural characteristics of fold-and-thrust belts with a frictional boundary at their base can be best understood by studying accretionary wedges. Total tapers of accretionary wedges reported in the literature vary strongly, from 2.9$^\circ$ for the Makran to 13.5$^\circ$
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for the Chile margin (Saffer and Bekins, 2006, and references therein). This variability is attributed to the relation between the internal and basal strength and the fluid pressure.

Thrusts are usually verging towards the toe (Figure 4.1a). Thrust sheets are developing sequentially (i.e. forward with respect to the sense of general thrusting) as the wedge grows by accumulating new material at its toe and thus building a new frontal thrust. The size of the thrust sheets strongly depends on the thickness of the incorporated rock sequence. Backthrusts appear at the rear of accretionary wedges when deformation reaches the internal parts (Poblet and Lisle, 2011). Many accretionary wedges exhibit large-scale underplating, i.e. sediments are thrust below an evolving wedge (e.g., Platt et al., 1985). Underplating leads to increasing confining pressures due to increasing load of the thickening wedge; the subsequent formation of duplex thrust sheets (Glodny et al., 2005) leads to faster surface uplift at the rear of the wedge (Ellouz-Zimmermann et al., 2007a).

4.2.2 Viscous salt décollement

If the main décollement is formed by a salt layer, it is viscous (Nettleton, 1934). A wealth of studies investigated the rheology of salt, defining its viscosity to be Newtonian (linear relation between stress and strain rate) or non-Newtonian (power-law relation between stress and strain rate) (Chemia et al., 2008, 2009; Marques et al., 2011; Spiers et al., 1990; Urai et al., 2008). In this study, we take both rheologies into account to model salt décollements.

The resistance to shearing of a purely viscous material is defined by the shear stress of a viscous fluid $\tau_b$, which is (for a Newtonian fluid) linearly proportional to the velocity gradient within the viscous layer and to its viscosity, and inverse proportional to the layer thickness (Turcotte and Schubert, 2002). This is expressed by:

$$\tau_b = \eta \cdot \frac{\Delta u}{h} \quad (4.3)$$

where $\eta$ is the dynamic viscosity and $\Delta u$ the difference in horizontal velocity along the vertical thickness $h$ of the viscous layer. In contrast to frictional décollements, there is no failure since a low viscosity fluid is easily sheared and flows. Equation (4.3) shows that the shear stress in a viscous layer, which influences the wedge formation, depends on the thickness, the viscosity and the difference of compression rate between the top
Another difference verified by equations (4.2 and 4.3), where $\sigma_N$ defines the normal pressure state, is the influence of the overlying rock sequence: in the case of a frictional décollement, the strength of the décollement layer and the strength of the overlying sequence both depend on the mass of the hanging wall. This leads to the fact that theoretical surface taper angles of wedges evolving on frictional décollements are independent of the thickness of the moving rock sequence (see equation (17) in Dahlen (1984)). In the case of a viscous décollement, the strength of the basal layer is pressure independent (see equation (4.3)), in contrast to the pressure dependent strength of the overlying rock sequence. As a consequence, the overburden thickness influences the wedge shape. Applying equation (4.3) from Davis and Engelder (1985) illustrates the theoretical surface slope angles of a wedge on a horizontal viscous décollement depending on the thickness of the cover sequence and the shear stress for the viscous layer, calculated by our equation (4.3) (Figure 4.3).
Several parameters defining the shear stress in a viscous décollement can be estimated empirically: the thickness of salt layers is known from drilling logs and seismic profiles and by GPS measurements constrain the velocity of active plate convergence (e.g., Sommaruga, 1997; Vernant et al., 2004). The viscosity of salt is more debatable, due to the wide range of published values, from $2.7 \cdot 10^{10}$ to $10^{23}$ Pa·s for Newtonian and non-Newtonian rheology, where most values fall between $10^{17}$ - $10^{20}$ Pa·s (Mukherjee et al., 2010, and reference therein).

Shear stresses defined by equation (4.3) can be very low in weak layers, so that fold-and-thrust belts on a viscous décollement have cross-sectional tapers much narrower than for other wedges (Davis and Engelder, 1985). Therefore, a wide fold belt can evolve as minor fold growth at the rear already results in a surface slope larger than the minimum critical taper. The deformation front is pushed further away from the backstop. In the Jura (Figure 4.1d) for example, narrow anticlines are clustered in the South (Faltenjura) and form box-folds separated by flat-bottomed, broad synclines in the North (Plateaujura) (Pfiffner, 2009; Sommaruga, 1997). An example for a salt-related fold-and-thrust belt with several weak zones is the Zagros Folded Belt (Motiei, 1993). The Simply Folded Zone of Zagros (Figure 4.1f) is characterized by periodic, symmetrical and open anticlines and synclines with an average wavelength of $\sim 14.5$ km (Yamato et al., 2011).

### 4.3 Model setup

#### 4.3.1 Governing equations

The mechanical model is based on the equations for conservation of momentum

$$- \frac{\partial P}{\partial x_i} + \frac{\partial \tau_{ij}}{\partial x_j} = \rho g_i$$

and the conservation of mass for an incompressible case

$$\frac{\partial u_i}{\partial x_i} = 0$$

where $P$ is pressure ($P = -\frac{\sigma_{ii}}{3}$), $\sigma_{ij}$ the stress tensor, $\tau_{ij}$ the deviatoric stress tensor ($\tau_{ij} = \sigma_{ij} + P$), $u_i$ the velocity ($u_1 = u_x$, $u_2 = u_z$), $x_i$ the spatial coordinates ($x_1 = x$, $x_2 = z$), $\rho$ the density and $g_i$ the gravitational acceleration ($g_1 = 0$, $g_2 = 9.81$ m/s$^2$).
The model follows a Maxwell visco-elasto-plastic shear rheology, defined by

$$
\dot{\varepsilon}_{ij} = \frac{1}{2\eta} \tau_{ij} + \frac{1}{2G} \frac{D\tau_{ij}}{Dt} + \lambda \frac{\partial Q}{\partial \sigma_{ij}}
$$

(4.6)

where $\eta$ is the effective viscosity, $G$ the elastic shear modulus, $t$ time, $\lambda$ the plastic multiplier and $Q$ the plastic flow potential (Moresi et al., 2007). The corotational derivative of the deviatoric stress tensor after Jaumann is defined by

$$
\frac{D\tau_{ij}}{Dt} = \frac{\partial \tau_{ij}}{\partial t} + u_k \frac{\partial \tau_{ij}}{\partial x_k} - W_{ik} \tau_{kj} + \tau_{ik} W_{kj}
$$

(4.7)

where $W$ is the vorticity ($W_{ij} = \frac{1}{2}(\frac{\partial u_i}{\partial x_j} - \frac{\partial u_j}{\partial x_i})$).

If differential stresses exceed the yield stress, rocks fail plastically according to the Mohr-Coulomb failure criterion. After Vermeer and De Borst (1984), the yield function in 2D is defined as

$$
F = \tau^* - \sigma^* \sin \varphi - C \cdot \cos \varphi
$$

(4.8)

and the plastic flow $Q$ as

$$
Q = \tau^* - \sigma^* \sin \psi
$$

(4.9)

where $\tau^* = \sqrt{(\sigma_{xx} - \sigma_{zz})^2 + \sigma_{zz}^2}$ and $\sigma^* = -\frac{1}{2}(\sigma_{xx} + \sigma_{zz})$. $C$ is the cohesion, $\varphi$ the friction angle and $\psi$ the dilation angle, which is zero in incompressible systems.

### 4.3.2 Numerical implementation

The governing equations described in the previous section are solved numerically by discretizing the time derivative of equation (4.6) in an implicit manner. The rheological behaviour is initially visco-elastic. If stresses exceed the yield stress ($F(\tau_{ij}) > 0$), effective viscosities are decreased until the maximum stresses are at the yield stress ($F = 0$). The discretized system of equations is solved with the finite element code MILAMIN_VEP (e.g., Kaus, 2010), which employs techniques described in Dabrowski et al. (2008) to speed up the matrix assembly. Plasticity is solved iteratively until either the error is below a critical value or after a given amount of iteration steps. Quadrilateral elements, bilinear shape functions for velocity and a constant shape function for pressure ($Q_1P_0$) in combination with direct solvers were used in all simulations of this study.
4.3.3 Boundary conditions and initial geometry

The same velocity boundary conditions have been applied to all simulations presented here. They simulate the mechanics of convergent plate boundaries where a rigid backstop scrapes upper levels of the crust off a rigid moving “plate” (Figure 4.4a). This is comparable to setups of analogue models, where a rough sheet lying below sand layers is pulled out below a fixed and rigid backstop (e.g., Konstantinovskaya and Malavieille, 2011). This setup also matches the boundary conditions used in the analytical critical wedge theory (Dahlen et al., 1984; Davis et al., 1983). Numerically, we apply free surface conditions on the top layer and a no-slip boundary at the left-side, non-deformable backstop. At the bottom and the right side, a negative constant horizontal velocity ($v_x = \text{compression velocity}$) and zero vertical velocity ($v_z = 0$) are applied, compressing the model in the horizontal direction. A singularity point is included at the left bottom node, the base of the backstop (Figure 4.4b). At this point, the code solves the system of equations with the bottom horizontal velocity. The resulting new coordinates of this point are reset afterwards to their initial value.

Figure 4.4: a) Model setup in a larger tectonic context. Rigid backstop appears as overriding plate at convergent subduction zone with accretionary wedge or as existing orogen or Molasse basin in foreland-type fold-and-thrust belts. b) Model setup. Parameters defined in Table 4.1. Boundary conditions are no-slip at the left side, constant velocity in the horizontal direction at the right side and at the bottom and free surface at the top boundary.
The initial geometric values of the simulations are listed in Table 4.1. Thicknesses of 2500 and 4500 m of scraped-off rock sequence are implemented for single salt décollement setups only. Variation of basal salt layer thickness (500 m, 1000 m, 2000 m) are applied for both single and multiple décollement simulations.

The characteristic rock parameters for the overburden sequence and the different décollements are listed in Table 4.2. All simulations contain an identical visco-elasto-plastic rock pile representing scraped-off upper crustal levels. Basal and intermediate shale décollements exhibit the same viscosity and shear modulus values as the overlying “rocks”, but have a lower failure criterion with a friction angle of $5^\circ < \varphi_b < 25^\circ$ and no cohesion. The values for $\varphi_b$ in relation to $\varphi = 30^\circ$ stand for the range of existing total taper angles of natural accretionary wedges Table 4.3. Viscosities implemented for basal and intermediate Newtonian salt layers range from $10^{17}$ to $10^{20}$ Pa·s. For non-Newtonian salt, a power-law coefficient of 5 has been used for a salt temperature of 50°C (Li et al., 2012).
4.4 Results

We discuss the influence of décollement architecture and rheology on the structural evolution of fold-and-thrust belts. The results are split in simulations with either salt or shale décollement types and with either multiple weak layers or only a single basal décollement. In all presented simulations, frictional shale décollements are displayed with green (basal) and blue (intermediate) colours and viscous salt layers with greyscale (linear viscous) and violet (power-law viscous) colours. Beige and brown colour bands display the initial horizontal layering of the rock stratigraphy. The colour difference is a passive marker for visualisation, marking no difference in rock properties. A velocity heterogeneity always occurs at the bottom left singularity point because we employed identical boundary conditions in all simulations. This implies that the first shear band

\[
\frac{\partial h_s}{\partial t} = \kappa \frac{\partial^2 h_s}{\partial x^2}
\]

where \( \kappa \) is the diffusion constant, and \( h_s \) the surface topography. The equation was solved numerically using the Finite Differences first order backwards diffusion approximation for the top nodes of the model. In all simulations \( \kappa \) has a value of \( 10^{-20} \) m\(^2\)/s. Numerical sedimentation is accomplished by only updating the coordinates at locations where diffusion increased topography.

### Table 4.2: Parameters for Rock Sequence and Décollements

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Description</th>
<th>Rock Sequence</th>
<th>Shale Layers</th>
<th>Salt Layers</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \varphi )</td>
<td>Friction angle (^{\circ})</td>
<td>30</td>
<td>5 – 25</td>
<td>30</td>
</tr>
<tr>
<td>( C_0 )</td>
<td>Cohesion (MPa)</td>
<td>20</td>
<td>0</td>
<td>20</td>
</tr>
<tr>
<td>( \eta )</td>
<td>Viscosity (Pa·s)</td>
<td>1e25</td>
<td>1e25</td>
<td>1e17 – 1e20</td>
</tr>
<tr>
<td>( \rho )</td>
<td>Density (kg/m(^3))</td>
<td>2700</td>
<td>2200</td>
<td>2200</td>
</tr>
<tr>
<td>( G )</td>
<td>Elastic shear modulus (Pa)</td>
<td>2e9</td>
<td>2e9</td>
<td>2e9</td>
</tr>
<tr>
<td>( n )</td>
<td>Power-law coefficient</td>
<td>1</td>
<td>1</td>
<td>1, 5</td>
</tr>
</tbody>
</table>

A diffusional sedimentation process depending on the surface curvature has been implemented to consider surface processes:
CHAPTER 4. DEFORMATION MECHANICS IN FOLD-AND-THRUST BELTS

Table 4.3: Parameters for Simulations With Single (F) and Multiple Frictional (DF) Décollement Layers

<table>
<thead>
<tr>
<th>Simulation</th>
<th>$\varphi_b$ (°)</th>
<th>$\varphi_m$ (°)</th>
<th>Figure</th>
</tr>
</thead>
<tbody>
<tr>
<td>F1</td>
<td>5</td>
<td>-</td>
<td>5, 6, 9</td>
</tr>
<tr>
<td>F2</td>
<td>10</td>
<td>-</td>
<td>4, 5</td>
</tr>
<tr>
<td>F3</td>
<td>15</td>
<td>-</td>
<td>5</td>
</tr>
<tr>
<td>F4</td>
<td>20</td>
<td>-</td>
<td>5</td>
</tr>
<tr>
<td>F5</td>
<td>25</td>
<td>-</td>
<td>5, 6</td>
</tr>
<tr>
<td>DF1</td>
<td>5</td>
<td>10</td>
<td>10, 11</td>
</tr>
<tr>
<td>DF2</td>
<td>10</td>
<td>10</td>
<td>10</td>
</tr>
<tr>
<td>DF3</td>
<td>10</td>
<td>5</td>
<td>10, 11</td>
</tr>
</tbody>
</table>

originates from this point in all simulations, independent of salt or shale, single or multiple décollements.

4.4.1 Single frictional (shale) décollement

Snapshots of a typical simulation with a single frictional décollement show that deformation starts at the rear of the model and migrates away from there while forming in-sequence, forward-verging thrust sheets (Figure 4.5). The plots below the snapshots show the effective (black) and yield (grey) differential stresses in the lowermost element row. For all snapshots, stresses in the décollement exceed yield stresses ∼40 km further away from the backstop than the actual deformation front in the overburden sequence. This happens because along the ∼40 km of décollement failure the hanging wall is not at its yield stress; therefore, the equivalent rock sequence undergoes pure shear. This explanation is supported by the fact that the length of failure at the base, in front of the wedge, decreases with an increasing strength of the basal décollement.

Simulations with a single frictional décollement were investigated with a wide range of yield strengths (Figure 4.6; simulations F1-5). Total tapers of the thrust wedges are larger for larger basal friction angles and roughly correlate with the theoretical minimum critical taper angles derived from the critical wedge theory (Figure 4.2; plotted lines in Figure 4.6).
Figure 4.5: Temporal evolution of fold-and-thrust belt (beige and brown, $\varphi = 30^\circ$) with a frictional décollement (green, $\varphi_b = 10^\circ$). Thrust wedge grows horizontally by in-sequence thrusting and vertically by reactivating thrusts within the wedge. Most of the horizontal shortening is accommodated by the active frontal thrust. Plots show the differential stress (black line) and the differential yield stress (gray line) in the lowest element row (no vertical exaggeration).
For all simulations with frictional décollement, the frontal thrust is active until the basal yield stress in front of the wedge toe is exceeded. Then, the toe of the wedge jumps by a distance $L_W$ according to the following equation,

$$L_W = 2 \cdot \frac{H_s}{\tan \theta}$$

where $L_W$ depends on the sequence thickness $H_s$ and the inclination $\theta$ from horizontal of the shear band defining the new ramp (Figure 4.7a), which in our numerical simulations
Figure 4.7: Second invariant of the strain rate tensor. Jump of deformation in a frictional wedge with basal friction angles $\varphi_b$ of (a) 5° and (b) 25°. Dip angles of the developing shear bands in the undeformed rock pile are identical in Figures 4.7a and 4.7b. Jump of deformation $L_W$ (equation (4.11) in text) depends on the sequence thickness $H_s$ and the angle $\theta$ between the shear band and the décollements. Stress orientations (converging white arrows) have analytically derived values according to the critical wedge theory (Dahlen, 1984). Dotted lines indicate Coulomb (C), Arthur (A), and Roscoe (R) angles derived from the theoretical stress field (no vertical exaggeration).

Initially follows Arthur rather than Coulomb orientations (Kaus, 2010). The initial inclination $\theta$ of the shear bands at the wedge toe is independent of the décollement strength (Figure 4.7) because $\sigma_1$ is close to horizontal in the undeformed rock pile, in front of the wedge toe. Accordingly, new shear bands are conjugate. However, these shear bands do not necessarily mark the location of the next ramping thrust. This is the case only if the friction angle in the basal décollement is relatively low ($\varphi_b < 10^\circ$). For larger basal friction angles, the deformation front is more dynamic and constantly generates new shear bands dipping towards the wedge (Figure 4.7b). Such shear bands do not necessarily produce recordable offsets and strain rates decrease very fast as long as a new active frontal ramp has not formed.

The orientation of the principal stresses within the wedge produces ramp thrusts that are shallower-dipping with increasing basal friction angle (Figure 4.7b). The largest principal stress $\hat{\sigma}_1$ is close to horizontal for compressional wedges with a very weak décollement. For wedges with a higher basal friction angle, i.e. with a larger total taper, $\sigma_1$ is steeper, raking away from the rear. This is due to a stronger asymmetric gravitational potential.
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Figure 4.8: Horizontal evolution of fold-and-thrust belts overlying (a) single shale and (b) salt décollements with different strengths. Average gradient of kinked lines indicate rate of migration of the deformation front away from the rear. In Figure 4.8a, abrupt jumps in deformation front: \( \phi_b = 5^\circ \). If \( \phi_b > 5^\circ \): shift of the deformation front through time. In Figure 4.8b, decreasing salt viscosity results in a faster horizontal growth of the wedge. Wide fold belts with single-standing structures: \( \eta_b < 1 \times 10^{19} \text{ Pa s} \).

in large-tapered wedges. Backthrusts occur predominantly in simulations with \( \phi_b > 10^\circ \) at the rear of the wedge. Backthrusts are steeper than forward verging ramps depending on the direction of the principal stresses (Figure 4.7b). Our model results (Figure 4.7) agree with the angles \( \psi_b \), between the horizontal décollement and \( \sigma_1 \) predicted by the critical wedge theory (Figure 4.2; Dahlen, 1984). Ramp and backthrust dipping angles within the wedge follow Coulomb orientations (Figure 4.7). This stands in contrast to the initial thrust orientation mentioned above.

Figure 4.8 shows the distance of the deformation front to the backstop versus time. These plots have been computed at every tenth time step by extracting the \( x \)-coordinate of the surface boundary node, which has an elevation difference (\( z \)-coordinate) of 5 m to the closest node in \( x \)-direction and is furthest away from the backstop. The elevation difference of 5 m implies a slope of 1.5\(^{\circ}\) for an undeformed grid. This value was found to be appropriate for our purpose.

The step-like shape of the graphs for low basal friction angles (Figure 4.8a) indicates a sudden jump of the deformation front (increase of distance to backstop over a short amount of time). The forward shift of the deformation front in models with high friction angles at the base can be observed in the graphs for higher basal shear resistances (\( \phi_b > 10^\circ \); Figure 8a). The steps become smaller in space and time and thus increase in numbers. Slightly larger steps every \( \sim 1.8 \text{ Ma} \) indicate the formation of a new frontal
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ramp with visible offset. The rate of forward migration of the deformation front decreases with increasing basal strength (Figure 4.8a). For the setup with $\varphi_b = 5^\circ$ the wedge toe migrates at a rate $\sim 2 \text{ cm/a}$ over the first 3 Ma and then reaches an average rate of $\sim 1 \text{ cm/a}$ for the rest of the simulation time. With $\varphi_b = 25^\circ$, the frontal thrust migrates away from the rear at of $\sim 1 \text{ cm/a}$ over the first 2 Ma and $\sim 0.5 \text{ cm/a}$ for the next 10 Ma (Figure 4.8a). The fact that the distance from the toe to the backstop decreases during the activity of a single frontal thrust, i.e. the slopes of the graphs are negative, indicates internal, horizontal shortening of the wedge. This internal strain is not related to the frontal thrust zone.

4.4.2 Single linear viscous (salt) décollement

As stated earlier, the shear resistance to sliding along viscous décollements is influenced by more factors than along a frictional base. For sake of clarity, all presented results were obtained with a shortening velocity of 1 cm/a. Simulations with higher rate were performed as well, but the influence of this rate on the shear stress in the décollement is minor compared to uncertainties in the values of effective viscosity (see equation (4.3) and Table 4.2).

Shear stresses in salt décollements follow equation (4.3). This implies that shear stresses in the salt increase with increasing viscosity and decreasing thickness. It also results in a decrease of velocity difference $\Delta u$ between the bottom and the top of the salt. Since the velocity at the salt bottom is constant (compression velocity), the decrease of $\Delta u$ results in a drag of upper crustal material towards the backstop. If the resulting shear stresses are low enough (critical taper $\sim 1^\circ$; $< 2.5 \text{ MPa if } H_s = 2.5 \text{ km}$, $> 7.5 \text{ MPa if } H_s = 4.5 \text{ km}$; Figure 4.3), $\Delta u$ can reach the value of the shortening rate and there is no drag of the rock sequence towards the backstop. To better understand the consequences of this, simulations with identical shortening rates and sediment and salt thicknesses have been performed. Different shear stresses were achieved with different viscosity values. For low salt viscosities ($\leq 10^{18} \text{ Pa-s}$), i.e. low basal shear stresses, critical taper angles of $\sim 0.2^\circ$ lead to a distributed deformation away from the backstop and eventually to folding and thrusting spatially unrelated to any previously developed structure. Symmetrical box folds or conjugate thrusts develop because the main principal stress acts horizontally. For simulations with a salt layer thickness of 500 m, a rock pile thickness of 2.5 km and a shortening rate of 1 cm/a, a viscosity $< 10^{19} \text{ Pa}\cdot\text{s}$ is required to yield the formation of single-standing box folds (Figure 4.9a). Higher basal
viscosity causes a drag of overburden towards the backstop due to lower shear strain rate within the décollement, resulting in wedge-shaped fold-and-thrust belts (Figure 4.9a).

For models with a rock thickness of 4.5 km, migration rates of the deformation front are \( \sim 0.4 - 5 \) cm/a (Figure 4.8b). The thickness of a basal salt layer influences the basal shear stress proportionally (see equation (4.3)). Furthermore, it is also influencing the deformation mechanisms of thrusting and folding. The overburden can sag more easily on a thicker viscous basal layer and, therefore, thrusts may have larger offsets and more asymmetry (Figure 4.9b).

Frictional layers fail when shear stresses exceed their yield strength. The wedge slides on the failing décollement up to the active frontal thrust. In the very front of it, the rock sequence undergoes pure shear over the length where the décollement is at its yield stress (Figure 4.5). Therefore, stresses can disperse into the wedge as far as pure shear appears (Figure 4.10). In contrast, a low viscosity salt décollement cannot fail since there is no yield criterion. This leads to increased strain rates even at low stresses. High
strain rates in basal viscous layers propagate long distances away from the deformation front, leading to high differential stresses in the layered sequence in front of the wedge, i.e. that part which is not yet incorporated in the fold-and-thrust belt (Figure 4.10b).
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Figure 4.10: Second invariant of the deviatoric stress tensor. Black arrows point to the location of the deformation front. a) Wedge with a frictional décollement ($\varphi_b = 5^\circ$). In front of the wedge toe, the basal layer is not at the yield stress. Therefore, no failure occurs along the décollement and stresses do not disperse far into the undeformed layered sequence. b) Wedge over a salt décollement ($\eta_b = 1e19 Pa\cdot s$). Initial low viscosity results in high strain rates in the viscous layer. Stresses are dispersed throughout the layered sequence (no vertical exaggeration).

4.4.3 Two frictional (shale) décollements

Adding a frictional décollement layer within the wedge sequence yields important structural differences with simulations with a single basal décollement (Figure 4.11a). The depth of the second weak layer has been conveniently chosen for all simulations to be at two thirds from the basal décollement. This is consistent with the fact that weak layers in frictional fold-and-thrust belts are commonly made of over-pressured shale, which needs a certain depth of burial so that the water pressure is sufficient to reach the failure criterion (Platt, 1990).

Fold-and-thrust belts with two frictional décollements grow sequentially like simulations with a single basal décollement (Figure 4.5). However, instead of thrust sheets and fault planes normally reaching from the surface down to the single décollement, shear bands and folds are interrupted downward by the intermediate weak décollement and develop a wedge-scale duplex structure.

Yet, the cross-sectional taper of the fold-and-thrust belt depends on the strength of the deepest active décollement. This means that failure of the basal décollement results in deformation of the whole layered sequence independent of the stress state within the weak intermediate layer. Deformation involves interplay between structures above and below the weak intermediate layer. Consequently, a failing intermediate layer and a stable base within the same vertical profile lead to underthrusting (tectonic underplating) of material below the deforming upper sequence (Figure 4.11a; simulation DF3).
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Figure 4.11: Simulations with multiple décollements. a) Basal and intermediate décollements consist of shale (green and blue) and are shown after 6.34 Ma. Straight lines indicate analytically derived surface tapers from the critical wedge theory. Simulation DF1: Intermediate décollement is stronger than basal one. The total wedge taper depends on frictional strength of basal décollement. Simulation DF2: Intermediate and basal décollement have the same strength. Simulation DF3: Basal décollement is stronger than intermediate one. Underplating leads to antiformal stacking at the rear. Short-wavelength imbrication appears at the toe in the rock pile. b) Simulations with two salt layers (gray scale) compressed during 4.98 Ma. Simulation DV1: High salt viscosity produces drag toward the backstop. Simulation DV2: Fault propagation folds with no preferred vergence. Very low surface taper. Simulation DV3: Low salt viscosity and increased salt thickness result in open folds with wavelengths comparable to Zagros folded belt (Moutheau et al., 2007a,b; Yamato et al., 2011) (no vertical exaggeration).
Three simulations with different relationships between the basal and the within-sequence décollements are discussed (Figure 4.11a). If the basal décollement is weaker than the intermediate one ($\varphi_b < \varphi_m$), the overall cross-sectional taper is controlled by the strength of the basal weak layer (simulation DF1). The theoretical minimum critical taper angle of 1.7° is consistent with the simulation result (Figure 4.11a; simulation DF1). Two structural wavelengths develop according to equation (4.11): (i) A short wavelength restricted to the uppermost pile, which is a function of the depth of the upper décollement and (ii) a broad wavelength, throughout the wedge, which depends on the depth of the basal décollement. The long wavelength structures initially arise in the upper part of the rock pile, with fault-propagation folds (Figure 4.11a; simulation DF1) evolving into ramp faults affecting the whole pile (Figure 4.12a). The fast growth of upper-partial fault-propagation folds at the toe of the wedge locally overcomes the critical taper related to the intermediate décollement and controls the short wavelength structures in the upper pile. In contrast, the forward jump of a ramp cross-cutting the whole sequence depends on the critical taper of the complete wedge, which grows much slower. Therefore, a relatively long wavelength in relation to the complete pile is produced.

The same strength for the basal and the intermediate décollements ($\varphi_b = \varphi_m$), intensifies tectonic underplating below the upper décollement (simulation DF2). New frontal thrusts, soling to the base, occur at regular intervals (Figure 4.11a; simulation DF2). The theoretical minimum taper angle predicted by analytical solutions for a frictional strength $\varphi_b = \varphi_m = 10^\circ$ and an internal friction angle $\varphi = 30$ is in this case 3.4° (Figure 4.2). The numerical models show a slightly steeper angle (5°). This very small discrepancy could be due to underplating and surface uplift of the rear part by antiformal stacking (Figure 4.11a; simulation DF2). Wavelengths on the surface depend on the thickness of the upper sequence after equation (4.11). Ramping of the lower sequence occurs with no preferred distance between thrusts.

Simulations in which the intermediate décollement is weaker than the basal décollement ($\varphi_b > \varphi_m$) are characterized by an unambiguous decoupling between the upper and lower parts of the wedge (Figure 4.11a; simulation DF3). As previously discussed, the taper angle increases with increasing basal strength while the forward migration rate of the frontal thrust decreases (Figure 4.6). Accordingly, the critical taper angle related to the intermediate décollement is smaller than that resulting from the basal décollement. Therefore, the deformation front of the upper sequence migrates faster than in the lower part. This leads to deformation propagating in the upper sequence without affecting the underthrust lower part. The ramp spacing depends on the initial sequence thickness.
Figure 4.12: Second invariants of the strain rate tensor for simulations with two frictional décollements. a) The intermediate décollement is stronger (ϕ_b = 10°) than the basal décollement (ϕ_b = 5°). b) Intermediate décollement is weaker (ϕ_b = 5°) than the basal décollement (ϕ_b = 10°). Stress profiles of distal and proximal parts for both cases. Empirical σ_1 (full line); theoretical main principal stresses σ'_1 needed to fail the layered rock sequence and the weak layers (dashed lines) (no vertical exaggeration).
according to equation (4.11), i.e. only the thickness of the upper sequence. The surface slope of the distal part of the model agrees with the expected minimum critical taper (1.7°), whereas the surface of the rear part, which is also influenced by underplating, is steeper (6 - 7°) than the theoretical value (3.4°). This difference results from strong underplating and the influence of the weak intermediate layer on the stress state of these underlying, underplating rocks.

In summary, the style and location of deformation in wedges with several frictional décollements vary strongly. In a fold-and-thrust belt with a weak intermediate layer, yet stronger than the basal décollement, critical stresses and therefore deformation occur predominantly at the front of the taper, with ramps climbing from the décollement to the surface (Figure 4.12a). If an intermediate layer is weaker than the basal décollement, a décollement takes place along this layer, preventing high stresses in the lower rock sequence (Figure 4.12b). This allows underthrusting. Deformation mainly acts in the upper part, at the wedge toe, whereas underplating and eventually antiformal stacks occur in the lower part, near the rear of the wedge (Figure 4.12b).

4.4.4 Two linear viscous (salt) décollements

In simulations with viscous décollements, strain rates increase within these layers compared to the surrounding layers, even at low stresses. For simplicity, we focus on cases in which the viscosity in the basal and in the intermediate décollement is the same. This stands in contrast to simulations with multiple frictional décollements, where the relative strength of the weak layers with respect to each other is the factor determining the deformation style of an evolving fold-and-thrust belt.

Three simulations with the basal shear stress decreasing by a factor of ten are presented (Figure 4.11b). The maximum basal shear stress is adjusted by means of the layer thickness and viscosity (see equation (4.3)). Simulations with two viscous weak layers grow sequentially with the same rate of frontal migration as the equivalent single décollement simulations. In two-décollement simulations with décollement viscosity of $5 \cdot 10^{19}$ Pa·s and a basal salt thickness of 500 m, the rock sequence is dragged towards the backstop and builds a wedge-shaped taper (Figure 4.11b; simulation DV1). The presence of an intermediate viscous layer concentrates high strain rates between the lower and the upper wedge pile, where faulting is the dominant mode of deformation. Forward verging imbrication independently evolves in the lower and the upper parts. Thrust sheet dimensions depend on the respective thicknesses. Folding occurs at the toe of the wedge in
form of long-wavelength fault-propagation folds in the upper part and in form of isolated box-folds $\sim 30$ km in front of the toe in the upper part.

If the salt viscosity is divided by five and the salt thickness is doubled, folding becomes more important (Figure 4.11b; simulation DV2). The lower part of the wedge is mainly deformed by thrusting, backthrusts and forward verging ramps appearing equally. The upper part shows three types of structures: (i) long-wavelength (10 km) fault-propagation folds, (ii) thrust ramps that reach down to the basal décollement and (iii) ramps contained within the upper wedge sequence (at a width of $\sim 90$ km).

Decreasing the salt viscosity tenfold leads to a fold-dominated structural pattern (Figure 4.11b; simulation DV3). The folds are open and show a regular wavelength of $\sim 15$ km. Although folding dominates, thrusting still accommodates some shortening. Fault planes appear as forward verging ramps and as backthrusts and cut through the complete rock sequence (Figure 4.11; simulation DV3). The weak material has a high flow potential and is able to fill the hinge of a growing anticline. The effect of an additional intermediate décollement on the deformation mode is best demonstrated by comparing single décollement with two-décollement simulations. Setups without an intermediate décollement show faulting as dominant deformation mode (Figure 4.9b; simulation V2), whereas an identical setup with an intermediate décollement is fold-dominated (Figure 4.11b; simulation DF3). Physically, this reflects the fact that the folding growth rate of a multilayer sequence is larger than that of a single layer (Schmid and Podladchikov, 2006; Yamato et al., 2011). If the overburden had a purely viscous rheology, the single layer case would deform by pure-shear thickening. As it is brittle in our simulations, faulting dominates. The case with an additional décollement layer, however, produces folds in a sufficiently rapid manner to install folding as the dominant deformation mode.

4.4.5 Simulations with non-linear viscous décollements

Urai et al. (2008) argued that the deformation mechanism for coarse grained salt is defined by dislocation creep with a power-law coefficient of $n \approx 5$. Accepting this argument, simulations with power-law viscous décollements were carried out using the rheology of Li et al. (2012), which describes the relationship between differential stress and strain rate. The initial geometry and defined parameters for the corresponding simulations are summarized in Table 4.4. Throughout basal décollements, a constant temperature of 50°C (following Li et al., 2012) is implemented for every setup.
Figure 4.13: Simulations with nonlinear viscous salt décollements. a) Single basal décollement. b) Basal and intermediate décollements. Plots show the effective viscosity of the lowermost element row in the basal salt (black) and the lowest viscosity in every element column in the intermediate décollement (gray). See Table 4.4 for initial geometry and rheological parameters.

Figure 4.13a shows two simulations with a single non-linear viscous ($n = 5$) basal décollement (simulation P1, P2). The plots below the horizontally compressed stratigraphy illustrate the effective viscosity in the bottom element row at the same time step. Simulation P1 has the same initial geometry as simulations in Figure 4.8a. Structurally, it is comparable with simulation V3, where the basal décollement has a viscosity of $10^{19}$ Pa·s. This is featured by the effective viscosity plot of simulation P1 in Figure 4.13a, which shows similar values as V3 below the deformed sequence (0 – 60 km). The drag of overburden material for simulations with a linear viscous basal décollement is not as effective as for non-linear cases, since the effective viscosity drops at higher stresses. Nevertheless, a wedge-shaped fold-and-thrust belt evolves with forward-vergent thrusts.
4.4. RESULTS

(Figure 4.13a; simulation P1). Simulation P2 has an identical geometric setup as simulation V6 (Figure 4.8b). The resulting effective viscosities for the bottom element row of simulation P2 are rather higher than the constant viscosity used in simulation V6. Accordingly, the wavelength of the deformation pattern is smaller, but forward- and backward-verging thrusts tend to develop equally (Figure 4.13; simulation P2).

Figure 4.13b shows two simulations with multiple non-linear viscous \( (n = 5) \) décollements (simulation DP1, DP2) with the related effective viscosity plots below. Besides the viscosity of the lowest element row, the effective viscosity of the intermediate décollement layer is plotted. This has been done by taking the lowest values for viscosity of the intermediate layer for every element column. Simulation DP1 and DP2 are geometrically identical to simulations DV1 and DV2, respectively (Figure 4.11b). Structurally, they resemble to linear viscous simulations with a décollement viscosity of \( 10^{19} \) Pa·s (Figure 4.11b; simulation DV2). The lower structural level of simulation DP1 forms forward-verging fault-bend anticlines. The upper level shows thrusts merging with lower level ramps and detachment faults without any connection to the lower part. Simulation DP2 has a thicker basal décollement. The ramp spacing in the lower level is consistently larger than in simulation DP1. The upper structural level is predominantly deformed by fault-propagation folds cored by the lower ramps (Figure 4.13b).

4.4.6 Limitations of the model

The numerical models presented in this study are very simplified. They concern a horizontally layered sedimentary pile pushed over a rigid "plate" by a rigid, perfectly vertical backstop. The advantage of this simplified setup is the possibility to compare results with existing analogue models, which generally use sand boxes (e.g., Konstantinovskaya and Malavieille, 2011). Furthermore, the simplified setup yields results that can be explained by analytical solutions such as the critical wedge theory, which uses similar geometrical boundary conditions (Dahlen et al., 1984; Davis et al., 1983). The chosen geometrical simplifications make it impossible to investigate several important factors known in natural fold-an-thrust belts, for example the influence of basement deformation, although thin-skinned and thick-skinned tectonics may coexist (e.g., Zagros; Mouthereau et al., 2007b). Also, the initially undeformed rock pile of the models neglects effects of inherited fault systems (e.g., Pyrenees; citealpTavani:2012). Furthermore, the overburden is set as an isotropic viscosity; the models are therefore unable to address layer parallel slip as reported in turbidites. Yamato et al. (2011) investigated the effect of flexural-slip by
implementing 100 to 200 m thick weak layers. However, the resolution of their and our simulations cannot integrate weak horizons on the centimetres scale of turbidite layers. Numerical models with an anisotropic viscosity are in principle capable of simulating this effect. This was previously shown to increase folding growth rates and produced chevron folds (Kocher et al., 2006).

In this study, imbricate shear bands begin to develop as conjugate thrusts and back-thrusts bifurcating from a point in the décollements layer. Out of sequence backthrusts additionally appear in the rear to maintain the critical taper. Although backthrusting is known in natural fold-and-thrust belts, it is less common at the wedge toe than models suggest. This discrepancy may be explained by the homogeneity of the modelled sedimentary pile. Yet, the frontal shear bands of models do not accumulate much strain, and we conjecture that they would be unnoticeable in the field.

4.5 Discussion

The results demonstrate that the rheology of the basal décollement and the presence of weak intermediate layers in the compressed rock pile play key roles in the evolution of a fold-and-thrust belt. Comparing the rheology of shale and salt décollements explains the structural differences of fold-and-thrust belts overlying either of these. If the décollement takes place along over-pressured shale, it follows a frictional failure criterion and propagates when stresses overcome the yield stress. If a fold-and-thrust belt is soled by a viscous salt layer, its structural evolution depends on the shear resistance of this layer.

Frictional fold-and-thrust belts grow horizontally in-sequence, independent of the strength of the décollement (Figure 4.5). The resulting total taper angle (Figure 4.6) and stress fields (Figure 4.7) are both consistent with the critical wedge theory. The stable frictional wedge is sliding along a basal décollement and accommodates shortening on an active frontal ramp in front of which the décollement does not extend. Therefore, differential stresses cannot build up further away in front of this frontal thrust (Figure 4.10a).

Fold-and-thrust belts related to salt décollements can develop a wedge-like shape if shear stresses are high; then material is dragged towards the backstop (Figure 4.9a; simulation V3). Related models exhibit a frontal ramp. Internally, the wedge is constantly compressed as the drag continues because no basal failure occurs. This stands in contrast to frictional décollements, which are at failure. Like this, salt décollements can distribute
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high strain rates far from the backstop and generate high stresses over a large horizontal
distance (Figure 4.10b). This leads to the development of wide fold-and-thrust belts
with little finite shortening, e.g., the Zagros Folded belt.

4.5.1 Folding versus faulting analysis for viscous décollement
simulations

Simpson (2009) provided an analytical solution for whether a horizontally compressed
elastic rock sequence overlying a viscous substratum is deformed by folding or faulting. In
order to achieve this solution, the time required for fast fold amplification $t_f$ is compared
with the time needed to fault the rock sequence $t_p$.

The equation for the folding time $t_f$ is derived from the maximum growth rate analysis
and the equation for the time required to produce an amplitude due to folding.

$$t_f = \frac{9}{16} \frac{\eta_b(1 - v^2)H_b^3 \ln A}{E\varepsilon(1 + v)^2H_b^3}$$

where $A$ is the amplitude factor (e.g., $A = 1000$ for explosive folding, Biot, 1961), $v$ is
the Poisson ratio, $E$ is Young’s Modulus, $\varepsilon$ is strain, $\eta_b$ is the décollement viscosity, $H_b$
is the décollement thickness and $H_s$ the overburden thickness.

The plastic time $t_p$ is calculated by setting the equation for horizontal elastic stress
against horizontal stress needed to build up the Mohr-Coulomb envelope everywhere in
the rock sequence. Then this equation is solved for strain and divided by strain rate:

$$t_p = \frac{(1 + v)(2v - 1 - \sin \varphi) \rho g H_s}{\dot{\varepsilon} E (\sin \varphi - 1)} + \frac{\cos \varphi(1 + v)(2C_0 v - 2C_0)}{\dot{\varepsilon} E (\sin \varphi - 1)}$$

where $\dot{\varepsilon}$ is strain rate. At variance to Simpson (2009), the maximum lithostatic pressure
is calculated from the thickness of the wedge sequence only instead of the total model
thickness. This is due to the fact that the décollement is not behaving plastic but purely
viscous.

If $t_f < t_p$, fast amplification engenders folding before stresses are at the yield. For
$t_f > t_p$, the base of the overburden is at failure before an amplitude required for folding
has evolved. The case $t_f = t_p$ designates the boundary between folding and faulting de-
formation modes. In Figure 4.14, similar to Figure 12 in Simpson (2009), this boundary
is shown for two cases ($H_s = 2500$ m and $H_s = 4500$ m) as a function of the ratio of
CHAPTER 4. DEFORMATION MECHANICS IN FOLD-AND-THRUST BELTS

Figure 4.14: Domains of folding/faulting depending on the décollement/overburden ratio and the normalized viscosity by equalizing equations (4.12) and (4.13) for overburden thickness of 2500 m (solid line) and 4500 m (dashed line). Thick line segments represent the range of nondimensional values from numerical simulations V1–V3 (solid line) and V4–V5 (dashed line). Circles are projected horizontal axis values for the dominant strain rate of labeled simulations.

The décollement thickness $H_b$ and the rock sequence thickness $H_s$ and the normalized viscosity of the décollement. Thick horizontal lines correspond to simulations V1-3 (full) and V4-5 (dashed) and illustrate the normalized viscosity using the range of horizontal strain rates observed at the décollement / overburden interface from the numerical results (Figure 4.14).

The folding versus faulting analysis indicates that folding dominates deformation in simulations V1, V5 and V6. The range of values for normalized viscosity of V2 and V4 trespasses the folding/faulting boundary and simulation V3 plots in the area, where faulting is dominant. For simulations V1-3, the change from folding to faulting with increasing décollement viscosity can be observed in Figure 4.9a. The structural pattern of simulation V2 can be compared with that of simulation V4 (Figure 4.9), both plotting on the boundary of faulting versus folding (Figure 4.14). The most controversial cases are simulations V5 and V6, which analytically yield folding as the dominant deformation mode (Figure 4.14), although numerical simulations illustrate that faulting takes place during shortening (Figure 4.9b). We conjecture that this anomaly is a result of the fact that our setup implies a velocity bottom boundary condition, whereas the amplification theory is rather applicable to free-slip condition.
4.5.2 Comparison to natural fold-and-thrust belts

Results indicate that structural styles of fold-and-thrust belts vary according to the type and strength of the main décollement. The occurrence of additional décollements within the thin-skinned rock sequence also influences the structural evolution. Now, several examples of natural fold-and-thrust belts are compared to our modelling results to illustrate the mechanics of their evolution.

The Hikurangi subduction margin, NE New Zealand (Figure 4.1a), exhibits a thrust imbricated accretionary wedge, up to 150 km wide (Lewis and Pettinga, 1993), and a taper of \(\sim 4^\circ\) (Barnes et al., 2010). Barnes et al. (2010) characterized the ancient décollement zone as a sheared matrix of mudstones from the upper plate containing basalt and chert blocks from the oceanic lower plate. This indicates that the basal décollement is close to the bottom of the incoming sedimentary sequence. The re-drawn seismic section of this accretionary wedge (Figure 4.1a) shows that there is no major intermediate décollement within the deformed sedimentary sequence and thrust ramps merge into the basal décollement. Imbricated thrusts are preferentially verging towards the front, in the SE (Figure 4.1a). Barnes et al. (2010) mention an extensive zone of proto-thrusts beyond the principal deformation front. Such proto-thrusts locally appear as conjugate systems and usually have an offset of few tens of meters with dips of 40 \(\pm 5^\circ\). These features are consistent with simulations performed with a single frictional décollement (Figure 4.6). The taper of \(\sim 4^\circ\) suggests an internal to basal strength relation similar to that of simulation F2 \((\phi = 30^\circ, \phi_b = 10^\circ)\) with a theoretical taper of 3.4\(^\circ\).

The Makran, SE Iran and SW Pakistan (Figure 4.1b), is one of the largest accretionary wedges worldwide with more than 350 km in cross-sectional width and more than half of its area exposed on land (McCall, 2002). The total taper is \(\sim 3.6^\circ\) (White and Ross, 1979). Although the incoming sedimentary pile is \(\sim 7\) km thick, only the upper 4 km are involved in the frontal imbricate fan (Grando and McClay, 2007). These upper levels are thrust along a weak horizon within the sedimentary sequence and the lower part is underplated. Platt et al. (1985) estimated that up to 50\% of the incoming sediments have been underplated so that the lower sedimentary sequence was tectonically thickened by a factor of two to three. Underplating in the rear part of the wedge leads to surface uplift of the today onshore part (Ellouz-Zimmermann et al., 2007a). These observations can be explained by simulations with multiple frictional décollements, where the roof décollement is weaker than the basal one (Figure 4.11a; simulation DF3). Active roof-
duplex and antiformal stacking both push up the surface at the back of the simulation, producing a surface slope steeper than on areas where underplating is absent. This is consistent with the cross-sectional surface taper of the Makran, which, at its rear, is slightly steeper (\(\sim 1^\circ\)) than in the coastal area (\(\sim 0.2^\circ\)). The lower strength of the basal décollement relative to the wedge sequence can be explained by the exceptionally thick (7 km) incoming sediment sequence, as at a depth of 5 km, up to 90% of available smectite in shale is converted into illite (Bekins et al., 1994). This process is accompanied by dehydration, which increases fluid pressure within the shale layers at deeper levels and thus weakens them. The probable absence of an intermediate décollement towards the rear explains the different wavelengths between the onshore and offshore Makran as well as recently active fault zones like the Ghasr Ghand Thrust (Figure 4.1b), which reaches down to the basal décollement as illustrated in simulation DF1 or DF2 in Figure 4.11a.

The Subandean fold-and-thrust belt, NW Argentina (Figure 4.1c), is one of the active thin-skinned fold-and-thrust belts in a retroarc setting (Echavarria et al., 2003). Décollement horizons follow Silurian shale at the base of the belt and upper Devonian shales within the stratigraphic sequence (Baby et al., 1992; Echavarria et al., 2003). The intermediate décollement separates upper from lower structural levels. In the lower structural level, eastward verging in-sequence thrusts generate fault-bend anticlines with gentle backlimbs and steeper forelimbs. The upper structural level is characterized by folds with steep flanks and narrow crests (Figure 4.1c; Echavarria et al., 2003). These structural features are replicated in multiple frictional décollement simulations, where the basal décollement is weaker than the intermediate one (Figure 11a; simulation DF1). The lower stratigraphic level in simulation DF1 forms forward verging ramps with fault-bend anticlines. The upper level deformation is controlled by the intermediate décollement and builds steep flanking box-fold shaped fault-propagation folds.

The Jura, NW Switzerland and E France (Figure 1d), is an Alpine fold belt displaced towards the NW over a salt décollement in middle and upper Triassic evaporites (e.g., Pfiffner et al., 1997). This décollement can be tracked along the base of the Molasse Basin down to the crystalline roots of the Alps. During the late Alpine orogeny, the Molasse Basin acted as a stiff unit transferring shortening to its front, thus building the Jura Mountains (Pfiffner, 2009). The south-eastern part of the Jura, the Faltenjura, is characterized by closely grouped anticlines, whereas anticlines in the north-western part, the Plateaujura, are separated by wide flat synclines (Pfiffner, 2009). The thin-skinned sequence is 800 to 2000 m thick and the décollement evaporites had an initial thickness of < 500 m (Sommaruga, 1997). The structure of the Jura is best fitted by a single linear
viscous décollement simulation with a viscosity of $10^{18}$ Pa·s (Figure 4.9a; simulation V2). This simulation produced closely gathered anticlines at the rear of the evolving fold belt and solitary anticlines with flat, wide synclines in the distal part. These solitary anticlines can develop because the stresses can be easily transmitted horizontally through the sedimentary sequence, setting the whole pile under a similar state of stress (Figure 4.10b).

The Parry Island fold belt in the western arctic islands, Canada (Figure 4.1e), is an ancient foreland fold-and-thrust belt attached to the Franklinian Mobile Belt, which results from the Late Devonian - Early Carboniferous collision of the Canadian and the Anabar-Aldan shields (Condie and Rosen, 1994; Fox, 1985; Harrison, 1995). A lower décollement has been detected in evaporite and rock salt layers whose thickness is 60 - 2200 m; an upper shale layer has a thickness of up to 950 m. The overall width of the fold belt is $\sim$ 200 km after bulk shortening of < 11%, which indicates the relative weakness of the bottom décollement. The lower plastic sequence exhibits forward- and backward-vergent ramp thrusts, locally building thrust stacks filling anticline cores of the upper sequence. The upper part shows upright anticlines and broad synclines, sporadically scattered by back- and frontward-verging thrusts (Fox, 1985; Harrison, 1995). Although the intermediate décollement level is formed of shale, the geological structures are consistent with simulations with two viscous décollements with a relatively high viscosity of $10^{19}$ Pa·s (Figure 4.11b; simulation DV2). A decreased thickness of the basal salt décollement would have a similar effect. Basal thrusts would appear because the available salt volume would be too small to fill potentially growing anticlines.

The Zagros belt, SW Iran (Figure 4.1f), results from post-collision shortening between the Arabian plate and the Iranian continental blocks (Mouthereau et al., 2007b). The Zagros Simply Folded zone is characterized by a $\sim$ 200 km wide folded belt with mostly symmetrical open folds. The dominant wavelength of these folds is $15.8 \pm 5.3$ km (Mouthereau et al., 2007a,b). Emami et al. (2010) defined two mobile levels within the stratigraphy. The lower mobile level lies right above the basement in the $\sim$ 1 km thick Hormuz evaporites. The upper mobile level is the $\sim$ 800 m thick Gachsaran evaporite. The measured wavelength and the fact that folding is the dominant shortening process is consistent with simulations exhibiting multiple viscous décollements with a salt viscosity of at most $10^{18}$ Pa·s (Figure 4.11b; simulation DV3). The thickness of the viscous layer in this simulation matches the estimated thickness of the Hormuz evaporites (Emami et al., 2010). An increase of the salt viscosity would decrease the ability of flexural slip between the upper and lower stratigraphic levels.
4.5.3 Comparison to previous modelling studies

The structural evolution of fold-and-thrust belts has been investigated in several ana-
logue and numerical modelling studies. Our numerical results are consistent with scaled
analogue models simulating the evolution of single décollement accretionary wedges.
Regularly spaced thrust ramps develop in a forward-vergent manner, each with minor
backward-vergent backthrusts at the leading edge, suggesting an initially conjugate fault
system (e.g., Mulugeta, 1988; Schreurs et al., 2006; Storti and Mcclay, 1995). We have
determined a forward jump of the deformation front and documented steepening of
thrust ramps during their integration into the growing wedge, in accordance with the
Figure 2 of Storti and Mcclay (1995). They used sand with friction angles of $\sim 29^\circ$ and
$23^\circ$, respectively. Their resulting cross-section looks alike our simulation F5 (Figure 4.6)
with $\varphi = 30^\circ$ and $\varphi_b = 25^\circ$.

Several analogue modelling studies investigate the influence of a viscous décollement on
the structural evolution of a frictional rock sequence (e.g., Bahroudi and Koyi, 2003;
Smit et al., 2010, 2003). In contrast to purely frictional models, where deformation is
dominated by forward-verging in-sequence ramp systems, backthrusts and deformation
gaps occur very frequently in modelled thrust belts over viscous décollements. Compress-
sion rate influences the structural evolution of the wedge sequence (Smit et al., 2003)
due to the change of shear stress in the viscous layer (equation (4.11)).

Konstantinovskaya and Malavieille (2011) have documented how an additional inter-
mediate décollement promotes underplating and antiformal stacking of the bottom se-
quence. We numerically obtained a similar evolution with a second frictional décollement
within the cover sequence. Underplating is most likely leading to antiformal stacks if
the upper frictional décollement has a lower failure criterion (Figure 4.11; simulation
DF3). Pichot and Nalpas (2009) and Couzens-Schultz et al. (2003) also investigated the
structural effects of multiple décollement levels using a basal and upper, relatively thick
viscous silicone layer. Their results and our numerical simulations with two visco us
décollements are comparable (Figure 4.11b; simulation DV2).

The mechanics of fold-and-thrust belts has been numerical investigated for more than
two decades (Borja and Dreiss, 1989). Studies focussing on the structural evolution
of accretionary wedge-like systems overlying a single frictional décollement show in-
sequence faulting with isolated thrusts forming in a piggyback manner (Buiter et al.,
Selzer et al. (2007) tested the effect of weak shear zones on the structural evolution of thrust belts. They obtained underplating and stacking at the rear when they applied a horizontally continuous weak inclusion, comparable to additional inter-sequential décollements in this study. Stockmal et al. (2007) presented an extensive study investigating the influence of multiple weak frictional décollements within a compressed sequence. In contrast to our study, they only address inter-sequential décollement layers weaker than the basal horizon, focussing on the influence of surface processes and strain weakening. Stockmal et al. (2007) achieve similar results as ours in terms of thrust sheet wavelength, antiformal stacking and restriction of deformation to the upper structural level above a very weak inter-sequential layer. The latter authors also agree with the critical wedge theory with respect to the total wedge scale.

Simpson (2009, 2010b) applied a viscous versus frictional substratum as décollement. Decreasing the viscosity in the basal layer suppressed the formation of in-sequence, forward-vergent thrusts and stimulated the occurrence of backthrusts and single-standing folds. This was also obtained in the models presented here. Yamato et al. (2011) have shown that folding is promoted if the sedimentary sequence contains several viscous décollements. We obtained a similar response (Figure 4.11b, simulation DV3). Simpson (2011) argued that the critical wedge theory is not fully applicable because differential stresses out of localized strain zones are at a sub-critical state and principal stress directions can strongly vary within a wedge. However, in our understanding, the critical wedge theory does define an overall near-failure state of stress in a wedge; this depends on both the internal wedge properties and the bottom shear resistance and is independent from internal deformation processes. Our simulations underline the robustness of the critical wedge theory concerning critical taper angles (Figure 4.5) and principle stress orientations (Figure 4.6).

4.6 Conclusion

A 2D finite element numerical model with a visco-elasto-plastic rheology was used to investigate the difference between fold-and-thrust belts overlying either a (linear and non-linear viscous) salt or a (frictional) shale décollement and the influence of additional intermediate weak layers. Simulations with single frictional décollements lead to forward-
verging in-sequence thrusting forming ramps with dip angles depending on the wedge-
internal stress directions. Viscous décollements lead to backward- and forward-verging
thrusts equally, not forming a typical imbricate fan. This also leads to solitary box-folds
like those of the Jura Mountains. Nevertheless, simulations with either a viscous or
a frictional décollement are both consistent with the critical wedge theory, concerning
surface taper angles and internal principal stress directions.

Intermediate décollements strongly influence the structural evolution of fold-and-thrust
belts. In the case of frictional décollement style, the strength relation between the
basal and the inter-sequential décollement is the main factor influencing the structural
evolution of a fold-and-thrust belt. This strength relation likely is the cause of under-
plating and antiformal stacking as well as the installation of fault-propagation folds.
Inter-sequential viscous décollements support flexural slip within a layered rock pile,
which enhances folding for low Newtonian viscosities ($10^{18}$ Pa·s). For higher viscosities,
layer-parallel gliding leads to detachment folds and fault-propagation folds in the upper
part. Simulations with power-law salt rheology ($n = 5$) indicated viscosities between
$4 \cdot 10^{18} - 5 \cdot 10^{19}$ Pa·s for the basal, and approximately one order of magnitude higher
for intermediate décollements at 50°C.

Modelled structural styles reflect natural fold-and-thrust belts: underplating in the
Makran accretionary wedge indicates the existence of an intermediate décollement weaker
than the main sole thrust. We also conclude that the basal décollement was weaker than
the intermediate one during the structural evolution of the Sub-Andean Thrust Belt.
Solitary box-folds in the Jura and folding-dominated crustal deformation in the Zagros
both point to salt viscosities of $\sim 10^{18}$ Pa·s.

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High-resolution 3D numerical modelling of thrust wedges: Influence of décollement strength on transfer zones

Jonas Ruh, Taras Gerya & Jean-Pierre Burg

Abstract

The mechanics and dynamics of thin-skinned compressible thrust wedges with prescribed offsets in the backstop, i.e. transfer zones, are investigated using a three-dimensional finite difference numerical model with a visco-brittle/plastic rheology. The main questions addressed are: (i) What is the influence of the initial length of the backstop offset and (ii) what is the effect of the frictional strength of the main décollement on the structural evolution of the brittle wedges along such transfer zones? Results show that the shorter the backstop offset, the earlier these two thrust planes connect, forming a curved frontal thrust along the entire width of the model. Younger, in-sequence thrusts are formed parallel to this curved shape. Long backstop offsets produce strongly curved thrust faults around the indenting corner. Simulations with a weak basal friction evolve towards almost linear frontal thrusts orthogonal to the bulk shortening direction. Increased basal drag in models with a strong décollement favours propagation of the backstop offset into a transfer zone up to the frontal thrust. These simulations revealed that surface tapers of the wedge in front of the backstop promontory are larger than what the critical wedge theory predicts, whereas the tapers on the other side of the transfer zone are smaller than analytical values. This difference is amplified with increasing length of the backstop offset and/or strength of the décollement. Modelled surface elevation schemes reproduce well the topographic patterns of natural orogenic systems such as the topographic low along the Minab-Zendan transform/transfer fault between the Zagros and Makran.

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5.1 Introduction

Plastic thrust wedges such as submarine accretionary wedge systems or compressional thin-skinned fold-and-thrust belts have been intensely studied over several decades, in particular since the application of the critical taper theory (e.g., Dahlen, 1990; Platt, 1990; Stockmal, 1983; Westbrook and Smith, 1983). Accretionary wedges and fold-and-thrust belts are related to convergence subduction and/or continental collision tectonics and develop by scraping an upper crustal rock sequence off the subducting oceanic or converging continental plate along a weaker basal décollement. Such major décollements are often water-saturated, over-pressured shale layers that can be treated as frictional material (Kopf and Brown, 2003; Saffer et al., 2001; Takahashi et al., 2007; Wang et al., 1980). Shortening of the compressed sedimentary sequence develops in-sequence thrusting, ultimately forming an imbricate fan. Isolated out-of-sequence thrusts may appear at the rear of the wedge (Morley, 1988; Storti and McClay, 1995). Folds and thrusts of the wedges are rarely cylindrical. Instead, they display lateral variations and limits.

A wealth of scientific studies has been dedicated to the understanding of the dynamics and mechanics of thrust wedges. In the early to middle 1980's, a series of influential papers has analytically described the feedback of the base dip angle of a sedimentary wedge, its internal and its basal strength on the resulting surface taper (Dahlen, 1984, 1990; Dahlen et al., 1984; Davis and Engelder, 1985; Davis et al., 1983). The structural evolution of accretionary thrust wedges was then investigated by analogue models (Graveleau et al., 2012, and references therein), but the last decade has witnessed the usage of a growing number of numerical techniques (e.g., Buiter et al., 2006; Ruh et al., 2012; Simpson, 2009). Both experimental approaches bear advantages. Analogue models are employing “real” materials and the structures produced can be compared to those of natural examples. Different viscous and frictional materials can be utilised and actual failure of the material generates shear zones. Numerical simulations, on the other hand, bear the advantage that parameters such as time, viscosity, brittle strength and geometrical scale implemented into the system of equations can be rigorously scaled to large scale tectonic systems. Furthermore, numerical results provide direct and absolute values for strain and stress and boundary conditions can be controlled throughout the whole simulation time.

Numerical modelling of fold-and-thrust belts, where faulting occurs if stresses overcome the yield stress of brittle materials, needs accurate treatment of brittle/plastic rheology.
Numerical modelling also requires a high resolution to produce spontaneously localizing, high strain rate shear bands narrow to the point that they can be compared to “faults”. In the case of accretionary wedges and fold-and-thrust belts, this localization applies to the basal décollement and the various thrust flats and ramps, while respecting the high viscosity of the modelled, deforming sedimentary pile. Effective viscosity variations across narrow shear bands often range over five orders of magnitude. This poses a tangible numerical challenge.

Many geological problems, such as the influence of the strength of single and multiple décollements (Fillon et al., 2012; Ruh et al., 2012; Stockmal et al., 2007) or the effect of elasticity and viscosity of the décollement and overburden (Simpson, 2009) have been investigated using two-dimensional (2D) numerical models. A three-dimensional (3D) setup is needed to study the effects of lateral variations in any of the geometrical and/or mechanical parameters that control wedges. Although 3D numerical studies with plastic/brittle rheology exist (Allken et al., 2011; Braun et al., 2008; Gerya, 2010a; Popov and Sobolev, 2008; Zhu et al., 2009), few concern convergent thin-skinned thrust wedges. We present a 3D, high-resolution, fully staggered finite difference grid, marker in cell model for thin-skinned fold-and-thrust belts with a visco-brittle/plastic rheology. We use this model to understand the influence of pre-existing backstop offsets on the structural evolution of accretionary wedges and the role of basal frictional strength within these systems.

We compare our 3D numerical results to the transfer/transform zone between the Zagros and the Makran, where the ~ 300 km long Minab-Zendan right lateral fault zone separates the Zagros foreland fold-and-thrust belt from the Makran accretionary wedge (Peyret et al., 2009; Regard et al., 2005; Smith et al., 2005). Our aim is not to investigate the origin of this transfer zone, i.e. the cause of the recent offset in the actual backstop between the two wedge systems. We rather want to understand the effect of the existing transfer zone on the structural deformation and subsequent topography during ongoing shortening.
5.2 Numerical model

5.2.1 Governing equations

We developed a three-dimensional, high-resolution, fully staggered grid, finite difference, marker in cell model with a standard visco-brittle/plastic rheology and an efficient OpenMP-parallelized multigrid solver (Gerya, 2010b; Gerya and Yuen, 2007). The mechanical model is built on the equations for conservation of mass assuming incompressibility (i.e., sediment compaction is neglected)

\[
\frac{\partial u_i}{\partial x_i} = 0
\]  

(5.1)

and the conservation of momentum, the Stokes equation

\[
- \frac{\partial P}{\partial x_i} + \frac{\partial \tau_{ij}}{\partial x_j} = \rho g_i
\]  

(5.2)

where

\[
\tau_{ij} = 2\eta \dot{\varepsilon}_{ij}
\]  

(5.3)

and

\[
\dot{\varepsilon}_{ij} = \frac{1}{2} \left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right)
\]  

(5.4)

\(P\) is dynamic pressure, i.e. mean stress, \(u_i\) the velocity \((u_1 = u_x, u_2 = u_y, u_3 = u_z)\), \(x_i\) the spatial coordinates \((x_1 = x, x_2 = y, x_3 = z)\), \(\tau_{ij}\) the deviatoric stress tensor, \(\rho\) the density, \(g_i\) the gravitational acceleration \((g_1 = g_2 = 0, g_3 = 9.81 \text{ m/s}^2)\), \(\eta\) the viscosity and \(\dot{\varepsilon}_{ij}\) the strain rate tensor.

If differential stresses exceed the yield stress, plastic failure follows the Drucker-Prager yield criterion with the plastic yield function \(F\) (equation 5.5) depending on the second invariant of the stress tensor \(\tau^{II}\) and the yield stress \(\sigma_y\)

\[
F = \tau^{II} - \sigma_y
\]  

(5.5)

where

\[
\tau^{II} = \sqrt{\frac{1}{2} \tau_{ij}^2}
\]  

(5.6)
and
\[ \sigma_y = P \cdot \sin \varphi + C \cdot \cos \varphi \]  \hspace{1cm} (5.7)

\( C \) is cohesion and \( \varphi \) the friction angle of the material. Strain or strain-rate weakening (e.g., Buiter et al., 2006) is not used in our model.

### 5.2.2 Numerical implementation

The governing equations given in the previous section are solved numerically by discretizing equation (5.3) in an implicit manner, using an efficient OpenMP-parallelized multigrid solver, fully parallel on 16 threads. We adopted a standard geodynamic modelling approach which uses an effective viscosity formulation for the numerical treatment of visco-brittle/plastic deformation (e.g., Buiter et al., 2006, and reference therein). Similarly to non-Newtonian (e.g., power-law) viscous rheology, effective viscosity for visco-brittle/plastic flows characterizes a local ratio between the deviatoric stress and the strain rate (equation 5.3). In places, where the plastic yielding condition is not reached the deformation is purely linear viscous in accordance to the assumed background rock viscosity. The rheological behavior of the model is initially purely linear viscous. If stresses locally exceed the yield stress \( F(\tau_{ij}, P, \varphi, C) > 0 \), effective viscosities \( \eta \) are decreased depending on the second invariant of the strain rate tensor \( \dot{\varepsilon}^{II} \) and the yield stress \( \sigma_y \) until the maximum stresses are at the yield stress \( F = 0 \), according to

\[ \eta_{vp} = \frac{\sigma_y}{2 \dot{\varepsilon}^{II}} \]  \hspace{1cm} (5.8)

where

\[ \dot{\varepsilon}^{II} = \sqrt{\frac{1}{2} \dot{\varepsilon}_{ij}^2} \]  \hspace{1cm} (5.9)

and \( \eta_{vp} \) is the effective viscosity corrected for plasticity.

Direct plastic (Picard) iterations are simultaneous with time stepping. To ensure initial model convergence and proper initiation of shear bands, the first \( \sim 400 \) Picard iterations are computed with a very small time step (one year) implying a negligible amount of model displacement. The following computational time steps are also relatively short (1000 years), which guaranties small material displacement per time step (<10 m) and ensures accurate treatment of plasticity and good convergence of the multigrid solver.
5.2.3 Initial geometry and material parameters

We present model setups with two different sizes for the finite Eulerian grid. The Eulerian grids are dimensioned in a way that account for different requirements: vertical thickening of the wedge, possible lateral structural changes throughout the evolving wedge and increasing length of the wedge while the wedge front migrates. A high resolution grid is necessary to develop high strain rate shear bands sufficiently narrow to be comparable to natural fault systems. A first series of simulations had a perfectly cylindrical geometrical setup, i.e., a vertical, planar backstop formed by the Eulerian grid boundary. For these requirements, dimensions in $x$-, $y$-, $z$-directions are $150 \times 50 \times 15$ km with a nodal resolution of $309 \times 85 \times 149$, respectively. The resulting cell size in these models is $487 \times 595.2 \times 101.4$ m in $x$-, $y$-, $z$-directions. Models with an initial offset of the backstop boundary have a size of $150 \times 100 \times 15$ km in $x$-, $y$-, $z$-directions, respectively, including a 50 km wide ($y$-direction) orthorhombic rigid body reaching into the model domain of the wedge. In these simulations, the backstop is represented by the Eulerian grid boundary from $y = 0 - 50$ km. From $y = 50 - 100$ km, the actual backstop boundary is offset by the length ($x$-direction) of the attached body (Figure 5.1). The nodal Eulerian grid resolution of these simulations is $309 \times 165 \times 149$ in $x$-, $y$-, $z$-directions, respectively. The resulting cell size in these models is $487 \times 609.8 \times 101.4$ m in $x$-, $y$-, $z$-directions. Lagrangian marker resolution of all simulations is two markers per cell in every direction. This setting results in a resolution of 7.6 million nodes with about 60 million Lagrangian markers for models with highest resolution.

From bottom to top, the initial marker distribution defines a 500 m thick rigid plate at the bottom. Above this rigid plate, a 500 m thick décollement horizon with a frictional rheology soles a 5 km thick “sedimentary” sequence. The rest of the marker grid above the “sedimentary” sequence is defined as “sticky-air”, an approach that mimics a quasi-free surface between the model and the “sticky-air” (Crameri et al., 2012; Gerya and Yuen, 2003; Schmeling et al., 2008; Zaleski and Julien, 1992). The roofing “sticky-air”-layer has an initial thickness of 9 km (Figure 5.1). The low density ($1$ kg/m$^3$) and small viscosity ($10^{18}$ Pa·s) of the air layer ensures sufficiently small normal stresses at the topography surface. The analytical prediction of the quality of the free-surface approach, given by Crameri et al. (2012; equations 7 and 18), is not directly applicable to our model setup because it was developed to test free-surface behaviour influenced by a gravity driven viscous plume. Nevertheless, the main parameters of the equation are density, viscosity and thickness of the sticky-air layer. Applying the equation to our
5.2. NUMERICAL MODEL

Figure 5.1: Model setup. From bottom to top: 500 m thick rigid plate (black), 500 m thick frictional décollement (light gray), 5 km thick sedimentary sequence (gray). Lowermost rigid plate (black) enters and exits the Eulerian grid. Dashed lines: three tested geometries (lengths 20, 40, 80 km) for a 50 km wide promontory (dark gray). Black arrows: Applied velocity boundary conditions. Top right: Definition of domain A (no promontory) and domain B (additional rigid body).

setup yields very low quality factors $C_{\text{max}} (\ll 1)$, which fulfils the general condition for a traction free surface. Crameri et al. (2012) also tested the influence of numerical space and time resolution. A vertical grid resolution of $\sim 100$ m and time steps of 1000 a, as introduced in this study, are adequate to ensure a quasi-free surface (personal confirmation from Fabio Crameri, 2013).

To investigate the structural evolution in function of the offset of the backstop front, three initial geometries have been introduced, differing in length of the rigid body $L_B$, i.e. backstop offset, in $x$-direction: $L_B = 20, 40$ and $80$ km. These promontories lie above the décollement horizon. Rigid backstop bodies have an initial thickness of 6500 m, looming 1.5 km above the “sedimentary” wedge sequence, and are therefore roofed by 7500 m of “sticky-air” (Figure 5.1). To simplify the discussion throughout this paper, we divide the model setup into two main domains: Domain A is defined by $y$-coordinates from 0 to 50 km, where no additional body is introduced. Domain B is the volume between $y$-coordinates 50–100 km, where the additional rigid body represents the backstop offset and promontory (Figure 5.1; top right).

Material parameters of the different model elements are listed in Table 5.1. Every geometrical setup has been modelled with three décollement strengths, $\varphi_b = 5^\circ, 10^\circ, 15^\circ$. 


### Table 5.1: Initial Material Parameters

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Description</th>
<th>Rigid Basal Plate</th>
<th>Décollement Layer</th>
<th>Rock Sequence</th>
<th>Rigid Backstop</th>
<th>Sticky-air</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\phi$</td>
<td>Friction angle (°)</td>
<td>-</td>
<td>5, 10, 15</td>
<td>30</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>$C_0$</td>
<td>Cohesion (MPa)</td>
<td>-</td>
<td>0.1</td>
<td>20</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>$\eta$</td>
<td>Initial viscosity (Pa·s)</td>
<td>$1e23$</td>
<td>$1e23$</td>
<td>$1e23$</td>
<td>$1e23$</td>
<td>$1e18$</td>
</tr>
<tr>
<td>$\eta_{\text{max}}$</td>
<td>Upper cutoff viscosity (Pa·s)</td>
<td>$1e23$</td>
<td>$1e23$</td>
<td>$1e23$</td>
<td>$1e23$</td>
<td>$1e18$</td>
</tr>
<tr>
<td>$\eta_{\text{min}}$</td>
<td>Lower cutoff viscosity (Pa·s)</td>
<td>$1e23$</td>
<td>$1e18$</td>
<td>$1e18$</td>
<td>$1e23$</td>
<td>$1e18$</td>
</tr>
<tr>
<td>$\rho$</td>
<td>Density (kg/m$^3$)</td>
<td>2400</td>
<td>2400</td>
<td>2400</td>
<td>2400</td>
<td>1</td>
</tr>
</tbody>
</table>
5.2.4 Boundary conditions

Instead of pushing the backstop over the weak décollement, boundary conditions are applied in a way similar to analogue sand box models where the bottom sheet is pulled out below a rigid backstop (e.g., Konstantinovskaya and Malavieille, 2011). For this, a velocity of 1 cm/a is defined at the lower ($z = 0$ km) and frontal boundaries ($x = 150$ km; Figure 5.1). New Lagrangian marker layers coming in through the front side allow large deformation. The back side of the setup ($x = 0$ km) acts as backstop with a no-slip boundary condition. At the bottom, the 500 m thick rigid plate of high constant viscosity is included, which allows for a better stability of the multigrid solver at the lower model boundary. Between the bottom rigid plate and the “backstop”, the 500 m thick décollement undergoes simple shear (1 cm/a at $z = 500$ m to 0 cm/a at $z = 1000$ m; Figure 5.1). Both lateral boundary conditions (at $y = 0$ km and $y = 100$ km) are defined as free slip along the boundary planes. At the upper boundary, an upward velocity is applied. This leads to an exit of “sticky-air” markers of the same volume as the material coming in at the front boundary to fulfill the conservation of volume.

5.2.5 Surface process

A two-dimensional diffusive surface process is implemented to address sedimentation in piggy-back basin style depressions. The topography surface is treated as a boundary surface between “rocks” and “sticky-air”. The position of the surface between “rock” and “sticky-air” is calculated for every single vertical Eulerian grid line by using the nodal density. Densities of the markers defined as “rock” and “sticky-air” are $\rho_{\text{rock}} = 2400$ kg/m$^3$ and $\rho_{\text{sticky-air}} = 1$ kg/m$^3$, respectively. The density of Lagrangian markers is averaged harmonically on the nodes. This averaging results in a density column that fades over four nodes from $\rho_{\text{rock}}$ to $\rho_{\text{sticky-air}}$ (Figure 5.2a; node 110 to 113). The elevation of the “sticky-air”/“rock” interface is placed above the highest node that has a density larger than the average ($\rho_{\text{surface}}$):

$$\rho_{\text{surface}} = \frac{(\rho_{\text{rock}} + \rho_{\text{sticky-air}})}{2} = 1200.5 \text{kg/m}^3 \quad (5.10)$$

In the example density column (Figure 5.2a), it is node number 111, at 11148.6 m elevation. A linear interpolation of density at node number 111 to density at the node number 112 (which is the first node with a density below $\rho_{\text{surface}}$, above node 111) crosses...
the average density $\rho_{\text{surface}}$. The elevation at which the density interpolation crosses $\rho_{\text{surface}}$ represents the elevation of the adjusted surface (Figure 5.2a). This procedure yields an “exact” topography independent of the nodal resolution. To simulate surface process, the topography surface is diffused at every timestep depending on its local 2D curvature:

$$\frac{\partial h_s}{\partial t} = \kappa \cdot \frac{\partial^2 h_s}{\partial x_i^2}$$

(5.11)

where $\kappa$ is the diffusion constant, $h_s$ the surface topography and $x_i$ the spatial coordinates ($x_1 = x$, $x_2 = y$). The diffusion constant is $\kappa = 10^{-9} \text{ m/s}^2$ in all simulations presented here. The topography diffusion equation was solved implicitly on one thread using the direct solver PARDISO (Schenk and Gartner, 2004, 2006).

To address a depositional process without erosion, markers with a “sticky-air” rheology below the updated surface transmute to “rock” rheology. Markers with “rock” rheology above the updated surface remain unchanged (Figure 5.2b). The present study does not aim investigating the influence of sedimentation and erosion on the evolution of thrust wedges. We refer to the wealth of publications investigating feedbacks between surface processes and the internal deformation of such systems (e.g., Beaumont et al., 1992; Hilley and Strecker, 2004; Koons, 1990). The intensity of topography diffusion process in our model is low and mainly aims to preclude overthrusting of low-viscosity “sticky-air” markers at the model surface, which may produce non-physical entering of the “sticky-air” into spontaneously forming thrust zones.

5.3 The critical wedge theory

In the late 1970's and early 1980's, several publications (Chapple, 1978; Dahlen, 1984, 1990; Dahlen et al., 1984; Davis and Engelder, 1985; Davis et al., 1983) compared the evolution of an accretionary wedge to a pile of sand pushed by a bulldozer. For a frictional décollement, as it is introduced in our model setup, the analytical solution predicts a minimum and a maximum critical total taper (defined by the surface $\alpha$ + base $\beta$ angle), depending on the internal strength of the wedge material, the strength of the décollement and the base dip angle $\beta$. In contrast, the analytical solutions of critical wedges overlying a viscous décollement depend also on wedge density, thickness and compression velocity. This dependence is due to the resistance to shear in a purely viscous décollement, which depends on the velocity difference between its bottom and top surfaces. The shear resistance in the linear viscous case is independent of the thickness
5.3. THE CRITICAL WEDGE THEORY

Figure 5.2: Diffusive surface process. a) Density change at the sticky-air/rock interface along a random vertical nodal grid line. Solid line: density along a node column. Dot-dashed line: average of sticky-air and rock density. Dashed line: elevation, where the interpolated density is equal to the sticky-air/rock density average. b) Application of the diffusive process. Solid line: interface before the diffusion. Dashed line: interface after the diffusion. Grey area: rock sequence after surface process application. White area: sticky-air after surface process application.

of the overlying sequence. In theory, whatever the type of décollement, the rear of the developing wedge thickens by thrusting until the critical minimum taper is reached, which sets the wedge in a stable mode. As long as the total taper remains above its critical minimum, the wedge is sliding along its décollement and is growing by frontal accretion of new material in an in-sequence style of thrusting. If the taper angle exceeds the maximum critical taper, it will be lowered by normal faults.

The critical wedge theory exists for non-cohesive as well as for cohesive cases. Taking into account the effect of cohesion, the wedge surface becomes concave if the décollement is planar (Dahlen et al., 1984). Simpson (2011; Figure 15) argued that cohesion strength actually has a negligible influence on the shape of brittle wedges. This argument is disputable, Nilfouroushan et al. (2012) having shown that the structural evolution of numerical wedges is actually sensitive to changes in cohesion.

The analytical critical taper theory assumes that the whole wedge is at a critical stress state throughout. Time is irrelevant for the analytical solution, which represents only the state when the taper increases or decreases while stresses are perfectly distributed. It is obvious that a mechanical brittle/plastic wedge, whether analogue or numerical, is not at failure throughout (Simpson, 2011). If the taper is below the critical minimum, the wedge fails and activates thrusts at its rear to increase its surface slope. If it is in a stable mode, above the critical minimum, the wedge does not deform internally.
and is accreting new material at the front, consequentially decreasing its critical taper. Analogue sandbox models (Davis et al., 1983) and two-dimensional numerical models (Ruh et al., 2012) demonstrated that tapers of compressional brittle/plastic wedges are levelling towards the minimum critical taper defined by the critical wedge theory.

5.3.1 Influence of basal friction

In this work, we are investigating whether the non-cohesive critical wedge theory can also be applied to non-cylindrical numerical wedge models. The analytically derived critical surface taper angles are plotted against the décollement angle for a wedge with an internal friction angle of $\varphi = 30^\circ$ and basal friction angles of $\varphi_b = 5^\circ$, $10^\circ$, $15^\circ$ (Figure 5.3a).

Basal friction affects the shape and the structural behaviour of a thrust wedge. The theory of non-cohesive critical wedges provides the expected surface slope ($\alpha$), the angle between the maximum principal stress $\sigma_1$ and the base $\psi_b$, and the angle between the maximum principal stress $\sigma_1$ and the top of the wedge $\psi_0$ (Dahlen, 1984; Davis et al., 1983):

$$\alpha + \beta = \tan \varphi_b + \frac{\beta}{1 + K} \quad (5.12)$$

where $K \approx \sin \varphi \frac{\sin \varphi_b + \cos \varphi_b (\sin \varphi - \sin^2 \varphi_b)^{1/2}}{\cos \varphi_b - \cos \varphi_b (\sin \varphi - \sin^2 \varphi_b)^{1/2}}$, $\varphi$ and $\varphi_b$ the internal and basal friction angles of the wedge and $\beta$ the slope of the wedge base.

$$\psi_b = \arcsin \left( \frac{\sin \varphi_b}{\sin \varphi} - \frac{1}{2} \varphi_b \right) \quad (5.13)$$

$$\psi_0 = \arcsin \left( \frac{\sin \alpha}{\sin \varphi} - \frac{1}{2} \alpha \right) \quad (5.14)$$

Two major points were raised from previous analogue and numerical studies: (i) The taper increases with increasing basal friction, which leads to a narrower wedge (Burbidge and Braun, 2002; Davis et al., 1983; Nilfouroushan et al., 2008) and (ii) with increasing basal friction, principal stresses plunge with higher angles towards the wedge toe, which leads to steeper back thrusts and flatter frontward-verging thrusts (Huiqi et al., 1992; Mulugeta, 1988; Ruh et al., 2012).
5.3. THE CRITICAL WEDGE THEORY

Figure 5.3: a) Flounder diagrams of the stability of frictional wedges with an internal friction angle $\varphi = 30^\circ$ and basal friction angles $\varphi_b$ of $5^\circ$, $10^\circ$, and $15^\circ$ (Dahlen, 1984) relating minimal and maximal critical tapers to dip angle $\beta$ of the basal décollement (points within the flounder-shaped area indicate a stable wedge, points outside are unstable wedges). Dashed lines: minimum critical surface taper angles for the three cases with a horizontal basal décollement ($\beta = 0^\circ$). b) Definition of angles $\alpha$, $\beta$, $\psi_0$, $\psi_b$ in a critical wedge and related principal stresses $\sigma_1$, $\sigma_3$ in two dimensions.
The dip angles of frontward- and backward-verging thrusts can be predicted from the angle $\psi_b$ between the wedge basis and the maximum principle stress. It is known that the angle between the inclination of shear bands and the maximum principal stress varies between the Roscoe angle $\theta_R$, the Coulomb angle $\theta_C$ and the Arthur angle $\theta_A$ (Arthur et al., 1977; Coulomb, 1773; Roscoe, 1970):

$$\theta_R = 45 - \frac{\Psi}{2}$$  \hspace{1cm} (5.15)

$$\theta_C = 45 - \frac{\varphi}{2}$$  \hspace{1cm} (5.16)

$$\theta_A = 45 - \frac{\varphi + \Psi}{4}$$  \hspace{1cm} (5.17)

where $\Psi$ is the dilation angle (Roscoe, 1970), which in our simulations is 0. Angles outside the Roscoe-Coulomb range are not expected to occur in numerical models (Kaus, 2010). The different angles of shear band plane dips ($\theta_R$, $\theta_C$, $\theta_A$) can be rotated by the predicted angle $\psi_b$ to calculate the expected dips of frontward and backward verging shear band.

Here, we want to compare our numerical results to the critical wedge theory in terms of surface slope and internal stress directions. Therefore, values for the expected taper, stress orientation and expected shear band plane dips after Roscoe, Arthur and Coulomb are listed in Table 5.2 and illustrated in Figure 5.3b.

### 5.4 Results

Twelve simulations are presented. Three simulations without backstop offset allow investigating the influence of basal friction ($\varphi_b = 5^\circ, 10^\circ, 15^\circ$). These three models have a width of $y = 50$ km and have been run for 7 Ma numerical time. Deformation initiated along the singularity line, where the rigid plate is pulled out below the backstop. Due to their simple geometry, these models developed cylindrical structures that compare well, in section, with two-dimensional simulations.

In addition to every cylindrical basal strength setup, simulations including three initial backstop offsets (transfer zones defined by adding rigid bodies; Figure 5.1) have been carried out. In the following, we refer to the rigid body as promontory because it is a strong backstop reaching into the wedge material. Simulations with a basal friction
### Table 5.2: Analytical Values Predicted and Derived from the Critical Wedge Theory for the three Cylindrical Model Setups (Figures 5.4 and 5.5)

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Description</th>
<th>Model 1</th>
<th>Model 2</th>
<th>Model 3</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\phi$</td>
<td>Internal friction (°)</td>
<td>30</td>
<td>30</td>
<td>30</td>
</tr>
<tr>
<td>$\phi_b$</td>
<td>Basal friction (°)</td>
<td>5</td>
<td>10</td>
<td>15</td>
</tr>
<tr>
<td>$\beta$</td>
<td>Base angle (°)</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>Surface slope (°)</td>
<td>1.7</td>
<td>3.4</td>
<td>5.3</td>
</tr>
<tr>
<td>$\psi_b$</td>
<td>Angle between $\beta$ and $\sigma_1$ (°)</td>
<td>2.5</td>
<td>5.2</td>
<td>8.1</td>
</tr>
<tr>
<td>$\psi_0$</td>
<td>Angle between $\alpha$ and $\sigma_1$ (°)</td>
<td>0.8</td>
<td>1.7</td>
<td>2.7</td>
</tr>
<tr>
<td>$\theta_{RF}$</td>
<td>Roscoe angle for frontward thrust (°)</td>
<td>42.5</td>
<td>39.8</td>
<td>36.9</td>
</tr>
<tr>
<td>$\theta_{RB}$</td>
<td>Roscoe angle for frontward thrust (°)</td>
<td>47.5</td>
<td>50.2</td>
<td>53.1</td>
</tr>
<tr>
<td>$\theta_{CF}$</td>
<td>Coulomb angle for frontward thrust (°)</td>
<td>27.5</td>
<td>24.8</td>
<td>21.9</td>
</tr>
<tr>
<td>$\theta_{CB}$</td>
<td>Coulomb angle for frontward thrust (°)</td>
<td>32.5</td>
<td>35.2</td>
<td>38.1</td>
</tr>
<tr>
<td>$\theta_{AF}$</td>
<td>Arthur angle for frontward thrust (°)</td>
<td>35.0</td>
<td>32.3</td>
<td>29.4</td>
</tr>
<tr>
<td>$\theta_{AB}$</td>
<td>Arthur angle for frontward thrust (°)</td>
<td>40.0</td>
<td>42.7</td>
<td>45.6</td>
</tr>
</tbody>
</table>

Angle of $\phi_b = 5^\circ$ were running for 5 Ma, models with larger basal friction ($\phi_b = 10^\circ$, $15^\circ$) for 6 Ma.

#### 5.4.1 Different basal décollement strength

Deformation of the compressed wedge, inferred from plotted Lagrangian markers, and the strain rate are illustrated for the three basal strengths after 7 Ma with the cylindrical model setup (Figure 5.4). Wedge tapers are narrower and have a steeper surface slope with increasing basal strength. The spatial distribution of markers transformed into new sediment indicates a piggyback style of transport with sedimentation taking place only at local troughs. Both, the deformed strata and strain rate plots (Figure 5.4) indicate an almost perfectly cylindrical deformation style of the wedge, independent of the basal strength (small irregularities along the $y$-axis are due to random marker distribution). This highlights the robustness of the numerical code and negates any lateral boundary effects. The second invariant of the strain rate tensor shows that the wedge material is deforming in a mainly brittle/plastic manner. Narrow, high strain rate shear bands delineate thrust faults. Strain rate values within the décollement indicate whether it is active or not.
Figure 5.4: Cylindrical simulations after 7 Ma run time, i.e., a total shortening of 70 km. Internal friction angle of \( \varphi = 30^\circ \), basal friction angles of a) \( \varphi_b = 5^\circ \), b) 10°, and c) 15°. Left: Lagrangian markers indicating wedge deformation. Right: Second invariant of the strain rate tensor.

If the décollement has a low frictional strength (\( \varphi_b = 5^\circ \)), a wide wedge with low maximum elevation develops (Figure 5.4a). The wedge front migrates toe-wards before a notable offset appears along shear bands. Therefore, it is difficult to identify thrusts with the plotted deformed strata. Frontal accretion takes place through conjugate, high strain rate shear bands whose symmetry is favoured by the almost horizontal main principal stress direction. The result is a wedge with a flat surface slope. Out-of-sequence thrusting takes place at the rear while failure is active at the toe of the wedge. This indicates that out-of-sequence thrusts act as a buffer to maintain the taper at its critical minimal value.
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Figure 5.5: Profiles of the second invariant of the strain rate tensor at 25 km width of the three simulations illustrated in Figure 5.4. Black dashed lines indicate frontward and backward verging Roscoe (R), Arthur (A), and Coulomb (C) angles. The rake of principal stress orientations, $\sigma_1$ and $\sigma_3$, and the dip of R, A, and C angles are listed in Table 5.2 for each model.

For higher basal décollement strengths ($\phi_b = 10^\circ$, $15^\circ$), the wedges are narrower and display steeper surface slopes (Figure 5.4b and c). Strain rate plots show two forward-verging thrusts very close to each other at the front of the wedges (Figure 5.4b and c). Accretion of new material mainly produces thrusts verging towards the wedge front, and minor shortening due to backthrusting. This is due to the increased plunge of stress directions within the wedge body for higher basal strength (Figure 5.3b, 5.5).

Analytically derived stress directions (equations (5.12 - 5.17) within a frictional wedge are compared to the orientation of shear band planes (Figure 5.5). 2D numerical models have shown that thrusts in compressional wedges tend to follow Coulomb orientations with respect to the main compressional stress direction (Ruh et al., 2012). In the present 3D modelling, high strain rate shear bands are close to the Arthur angle, sporadically tending towards the Roscoe angle (Figure 5.5a and c). In general, simulations coincide with the analytical theory and the principal compression direction is increasingly steep towards the wedge toe with increasing basal strength. A consequence is that developing thrust ramps dip shallower and back thrusts steeper than in wedges with low basal friction.

Snapshots of the profile at $y = 25$ km (Figure 5.6) of the simulation with a basal friction angle of $\phi_b = 15^\circ$ (Figure 5.4c) show that deformation initiates at the rear of the model. The deformation front migrates through time away from the rear by forming in-sequence, forward verging thrust sheets. Sedimentary basins are formed in a piggy-back manner.
on top of rotating thrust sheets. Material is also deposited in front of active thrusts. During horizontal growth of the wedge due to frontal accretion, active out-of-sequence thrusting deforms the rear part of the wedge. Through time, these formerly active, forward-verging and out-of-sequence thrusts are deactivated and rotate. This can be observed for the first thrust at the very back of the model (Figure 5.6). After 1 Ma it dips $\sim 45^\circ$ towards the backstop. After 7 Ma, it is almost vertical.

The surface slopes have been calculated by linear regression of all nodal surface elevations in a profile in $x$-direction at $y = 25$ km (Figure 5.7b). Another way of measuring the surface slope is to connect valleys (troughs in the surface line; Stockmal et al., 2007). For the linear regression, only surface elevations 100 m higher than their initial value have been considered to neglect the flat fore-wedge that has not yet been incorporated into the wedge. At the beginning ($\sim 0.25$ Ma), all slopes are negative. This is due to the activation of the first forward verging thrust. Then, they show very high tapers ($\sim 0.8$ Ma), which result from the fact that only one or two thrust sheets construct the surface slope of the still narrow taper (e.g., Figure 5.6; at 1 Ma). The total tapers of simulations with a frictional strength $\varphi_b < 15^\circ$ reach their critical minimum at around 3 Ma. Then, out-of-sequence thrusts keep the taper in the stable mode. At 7 Ma run time, all simulations produced stable wedges and therefore accrete new material at their front.

Knowing the expected slope and the incoming material flux, one can predict the length ($x$-direction) of the evolving wedge. All incoming rock volume $V_I$ must be incorporated into the wedge and therefore be equal to the volume situated above the initially flat material (Figure 5.7a). $V_I$ can be calculated from the thickness of the incoming pile $H$, the velocity $v_x$ and time $t$ according to

$$V_I = H \cdot v_x \cdot t \quad (5.18)$$

The expected volume of the wedge above its initial level $V_W$ depends on the analytical taper angle $\alpha$ and the wedge length in $x$-direction $L_W$ (the horizontal distance from the backstop to the frontal thrust)

$$V_W = \frac{L_W^2 \cdot \tan \alpha}{2} \quad (5.19)$$
Figure 5.6: Temporal evolution of a brittle wedge (orange and brown, equal property layers) overlying a horizontal frictional décollement (green) at 25 km (y-direction) from the model side boundaries of the simulation with $\varphi = 30^\circ$ and $\varphi_b = 15^\circ$ (Figure 5.4c). Note in-sequence thrusting and wedge thickening due to out-of-sequence thrusting at the rear. Blue: new sediments deposited on the back limbs of fault anticlines. Solid line: analytically derived minimum critical taper. No vertical exaggeration.
Figure 5.7: Analytical test of the numerical results. a) Profile sketch of a wedge. Grey: incoming volume \((V_I)\) is equal to volume above initial level of the sequence \((V_W)\). b) Temporal evolution of surface tapers of perfectly cylindrical simulations according to their basal frictional strength (Figure 5.4). Full lines: numerical results. Dashed lines: analytically derived minimum critical tapers. c) Temporal evolution of the wedge lengths \((x\)-direction). Full lines: numerical results. Symbols: time-dependent analytically derived wedge lengths.

Since \(V_W\) and \(V_I\) must be equal, equations (5.18) and (5.19) can be simplified and solved for the expected length \(L_W\)

\[
L_W = \sqrt{\frac{2 \cdot H \cdot v_x \cdot t}{\tan \alpha}}
\]  

(5.20)

In Figure 5.7c, the analytically derived \(L_W\) is plotted against time for the three considered cases \((\varphi_b = 5^\circ, 10^\circ, 15^\circ)\). As previously seen, wedges with higher basal friction produce narrower tapers with steeper surface. The analytical wedge lengths match the numerically modelled wedges, especially for higher basal friction. For the low basal friction case \((\varphi_b = 5^\circ)\), the fit between the analytical and the numerical solution is not as accurate as for the stronger décollement simulations. A reason could be that deformation is reaching the box boundary \((x = 150 \text{ km})\) of this simulation at about 7 Ma.

5.4.2 Temporal evolution of models with a backstop offset

In the following part, the model with a basal friction angle of \(\varphi_b = 10^\circ\) and a backstop offset of \(L_B = 80 \text{ km}\) is described to illustrate the evolution of a thrust wedge influenced
by a predefined backstop promontory and lateral transfer zone. As already mentioned, we refer to the wedge part pushed by the indenting promontory as “domain B”, and to the wedge growing in front of the Eulerian grid boundary as the “domain A”. The promontory is clearly visible by its initially higher topography in contrast to the plastic sequence. The first thrust sheet develops along the backstop in domain A and at the front of the promontory in domain B (Figure 5.8a). The thrust sheet pushed by the promontory (domain B) is not linked to that formed in domain A. After 2 Ma, a second, in-sequence thrust develops in domain B, bending and fading out towards the transfer zone (Figure 5.8b). At the same time, domain A of the wedge already exhibits three thrusts, indicating that this part of the wedge is growing faster in \( x \)-direction (convergence direction) than domain B, i.e., strain is more distributed. In domain A, the thrusts do not develop perfectly cylindrical, i.e., orthogonal to the bulk shortening direction. Due to connection of the thrust sheets with the promontory side, shortening is not provided by large fault offsets along the thrusts (Figure 5.8c). After 4 Ma, a topographic low is detectable along the transfer, lateral boundary of the promontory in domain A, in front of the first thrust sheet (Figure 5.8d). After 5 Ma, the maximal elevation at the rear of the wedge is larger in domain A than in domain B (Figure 5.8e). Furthermore, the frontal thrust, though still curved, is connected over the whole width (\( y \)-direction) of the model. The offset of the wedge front at the lateral boundaries (\( y = 0 \) km and \( y = 100 \) km) is smaller (30 km) in \( x \)-direction than the initial backstop offset (80 km; Figure 5.8e). The topographic low, which started to develop at ca. 4 Ma, is significant at 6 Ma. This topographic low develops because the wedge in domain A is strongly non-cylindrical (Figure 8f). Whereas the first thrust fault at the rear develops throughout the whole domain A, the second and third ones are less wide (\( y \)-direction). This forms a triangular (map view) elevated zone bordering the relative topographic low (Figure 5.8f).

### 5.4.3 Influence of backstop offset \( L_B \)

The influence of the backstop offset \( L_B \) is illustrated by plotting the surface topography in map view for all models at their final stage (Figures 5.9, 5.11 and 5.13). To illustrate the mechanics of these models with an induced transfer zone, slices at \( y = 25 \) km and \( y = 75 \) km show material phases and the second invariant of the strain rate tensor for the different backstop offsets, according to their basal strength (Figure 5.10, 5.12 and 5.14).
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Figure 5.8: Map view of temporal evolution of a compressional wedge with a basal friction angle of $\phi_b = 10^\circ$ and a backstop offset of $L_B = 80$ km. a) Elevated quadrangle in the right top indicates position of the promontory. d) Black arrow points to topographic low next to the promontory. e) Dashed lines: position of the frontal thrust at the lateral boundaries ($y = 0$ km, $y = 100$ km). Solid line: orientation of the frontal thrust. 1 Ma = 10 km of shortening.

5.4.3.1 Low basal friction ($\varphi_b = 5^\circ$)

Models with different backstop offsets and a basal friction angle of $\varphi_b = 5^\circ$ are compared to their cylindrical equivalent at 5 Ma run time (Figure 5.9). Elevation of the cylindrical model is maximal at the rear, reaching $\sim 4.5$ km higher than the initial surface of the sedimentary sequence (Figure 5.9a). All models with a backstop offset show higher maximal elevation (Figure 5.9b-d). Maximal elevations of all modelled wedges with a low basal friction are similar in domains A and B, reaching 5 to 5.5 km higher than the
5.4. RESULTS

Figure 5.9: Topography of simulations with a basal friction angle of $\phi_b = 5^\circ$ at 5 Ma run time. a) No promontory. b) $L_B = 20$ km. c) $L_B = 40$ km. d) $L_B = 80$ km. White dashed lines: position of slices in Figure 5.10.

initial surface. In the model wedge with a backstop offset of $L_B = 20$ km, the 5 Ma frontal thrust is nearly linear and orthogonal to the bulk shortening direction (Figure 5.9b). The curvature of the frontal thrust and the thrust sheets around the promontory corner is tighter with longer backstop offsets ($L_B = 40$ and 80 km; Figure 5.9c and d). The models with longer backstop offsets also exhibit a larger-area topographic low along the transfer zone than models with a short backstop offset ($L_B = 20$ km).

Strain rate patterns illustrate that the wedges deform by plastic failure in narrow shear bands (Figure 5.10). Independent of the backstop offset, frontal thrusts form symmetric, conjugate shear bands. Active shear bands verging towards the wedge front exhibit higher strain rates than conjugate backthrusts (Figure 5.10).

Models with identical basal friction angles ($\phi_b = 5^\circ$) develop differently, depending on the backstop offset. Two major observations merit emphasis: (i) the domain B pushed by the promontory develops a narrower wedge with increasing backstop offset: Whereas there are eight thrusts in the model with a backstop offset of $L_B = 20$ km at 5 Ma (Figure 5.10a), there are seven for an intermediate backstop offset (Figure 5.10b) and
only six in the model with a backstop offset of $L_B = 80$ km (Figure 5.10c). (ii) Longer backstop offsets lead to wider wedges in domain A (Figure 5.10).

5.4.3.2 Medium basal friction ($\varphi_b = 10^\circ$)

Models with a basal friction angle of $\varphi_b = 10^\circ$ and different offsets in backstop at 6 Ma differ strongly in their topographic elevation depending on the backstop offset length (Figure 5.11). The cylindrical model (Figure 5.11a), matching the critical wedge theory (Figure 5.7), serves as a reference for the wedges influenced by transfer zones. The main difference between the models with different backstop offsets is the maximal elevation in domain B. For $L_B = 20$ km, the wedge domain B has a higher rear topography than domain A (Figure 5.11b). Furthermore, this model does not exhibit the topographic low described in previous experiments. In the model with a backstop offset $L_B = 40$ km, maximal elevation is similar in both wedge domains (Figure 5.11c). A topographic low also appears in domain A, next to the promontory, disconnecting the thrusts in domain
5.4. RESULTS

Figure 5.11: Topography of simulations with a basal friction angle of $\varphi_b = 10^\circ$ at 6 Ma run time. a) No promontory. b) $L_B = 20$ km. c) $L_B = 40$ km. d) $L_B = 80$ km. White dashed lines: position of slices in Figure 5.12.

B from those in domain A. For the backstop offset $L_B = 80$ km, the maximal elevation is lower in domain B than in domain A (Figure 5.11d). The thrusts in domain B are strongly curved around the promontory corner and topographic elevation decreases and fades out towards domain A.

Structurally, wedges with a basal friction angle of $\varphi_b = 10^\circ$ differ from models featuring a décollement friction of $\varphi_b = 5^\circ$. Thrusts in domain B exhibit larger offsets and are clearly verging towards the wedge front for a basal friction angle of $\varphi_b = 10^\circ$ (Figure 5.12). Like in the low basal friction models, wedges in domain A are increased in profile length with increasing backstop offset. Accordingly, single thrusts in the model with a backstop offset of $L_B = 20$ km (Figure 5.12a) exhibit larger offsets than thrusts in the model with a backstop offset of $L_B = 80$ km (Figure 5.12c).

5.4.3.3 High basal friction ($\varphi_b = 15^\circ$)

In the models with a backstop offset overlying a basal décollement with a friction angle of $\varphi_b = 10^\circ$, the maximal elevation in domain B strongly depends on the offset (Figure 5.11). In contrast, models with $\varphi_b = 15^\circ$ after a run time of 6 Ma exhibit different maximal elevations in domain A depending on the backstop offset (Figure 5.13). For
Figure 5.12: Slices showing wedge deformation inferred from Lagrangian marker grid (left) and second invariant of the strain rate tensor (right) for simulations with a basal friction of $\varphi_b = 10^\circ$ at 6 Ma run time. a) $L_B = 20$ km, b) $L_B = 40$ km c) $L_B = 80$ km. Location of the slices is indicated in Figure 5.11.

$L_B = 20$ km, the maximal rear elevations of both domains are similar at $\sim 8$km (Figure 5.13b). The frontal thrust and major thrust sheets are connected throughout the whole width ($y$-direction) of the model. With longer backstop offsets, the maximal elevation of the wedge is developed in domain A (Figure 5.13c). In contrast to models with lower friction angles ($\varphi_b = 5^\circ, 10^\circ$) and a backstop offset of $L_B = 40$ km, the topographic low is smaller. In the model with a backstop offset of $L_B = 80$ km, the maximal elevation in domain A is up to 2 km higher than the maximal elevation in domain B (Figure 5.13d). Thrust sheets are curved around the promontory corner and their surface elevation decreases and fades out towards domain A.

Plotted profiles through material phases show that wedges in domain B are wider ($x$-direction) for a shorter backstop offset (Figure 5.14a) and get narrower with less thrust sheets for an increasing backstop offset (Figure 5.14b and c). All models mainly exhibit thrusts verging towards the wedge front. For an offset $L_B = 80$ km, the wedges in domain A exhibit shear bands with little finite strain, i.e., a limited offset (Figure 5.14c). Strain rate patterns in these models show that frontal thrusts are not associated with conjugate
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Figure 5.13: Topography of simulations with a basal friction angle of $\phi_b = 15^\circ$ at 6 Ma run time. a) No promontory. b) $L_B = 20$ km. c) $L_B = 40$ km. d) $L_B = 80$ km. White dashed lines: position of slices in Figure 5.14.

faults as in low basal friction models (Figure 5.10). In addition, forward verging thrusts are flattened and backthrusts are steeper than in previously described models (Figure 5.14).

5.4.4 Indenter deformation

The indenter is compressed along the shortening direction in all experiments. The length of the indenters is decreasing linearly with time, where compression is larger for stronger basal décollements (Figure 5.15a). Pure shear within the indenter body results in $\sim 10\%$ thickening of close to the rear for all models and is negligible at its front. Most of the subsequent shortening is absorbed in an elongated buckle fold at the rear of the indenter, with a trend parallel to the model boundary (Figure 5.8, 5.9, 5.11, 5.13). The wavelength of the indenter fold depends on its thickness (Figure 5.10, 5.12, 5.14). The absolute amount of indenter shortening increases with increasing basal friction, as mentioned above. This effect is enhanced for longer backstop offsets (Figure 5.15b). This is illustrated by the relative shortening of the indenter to its initial length $L_B$ (Figure 5.15c).
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Figure 5.14: Slices showing wedge deformation inferred from Lagrangian marker grid (left) and second invariant of the strain rate tensor (right) for simulations with a basal friction of $\varphi_b = 15^\circ$ at 6 Ma run time. a) $L_B = 20$ km, b) $L_B = 40$ km c) $L_B = 80$ km. Location of the slices is indicated in Figure 5.13.

Even though shortening takes place within the indenter, it does not falsify the results in terms of surface slope. The minor deformation of the indenter lowers the actual shortening velocity within the wedge in front of the indenter. But because the analytical wedge theory is time independent, and therefore independent of shortening velocity, the expected tapers are equal.

5.5 Discussion

All modelled wedges grow in-sequence by developing frontal thrust ramps and accreting new material at their front (Figure 5.6). Structural and morphological variations in simulations without the backstop promontory are related to differences in basal friction. Results confirm numerically, after conclusions from analogue and 2D numerical models, that the rheological strength of the basal décollement is one of the most important factors for the structural and topographic evolution of a thin-skinned compressional thrust wedge. Low basal friction allows very wide (from rear to toe), low tapered fold-
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Figure 5.15: a) Length of the indenter over time for simulations with an initial indenter length of $L_B = 80$ km and basal friction angles of $\varphi_b = 5^\circ$, 10°, and 15°. b) Length of indenters after 5 Ma runtime depending on basal friction angle $\varphi_b$ and initial indenter length. c) Relative shortening of indenters after 5 Ma runtime depending on basal friction angle $\varphi_b$ and initial indenter length $L_B$.

...and-thrust belts (Figure 5.4a). The inclination of developing shear bands within the wedge also depends on the basal resistance to shear (Figure 5.5), as it has been shown in analogue (Davis et al., 1983) and 2D numerical models (Ruh et al., 2012). In a wedge with low basal friction, the main stress direction remains close to horizontal (Figure 5.5). In such a stress regime, frontal accretion tends to develop conjugate, symmetric shear bands. The principal, compressional stress directions are increasingly inclined with increased basal friction so that forward verging thrust ramps tend to dip more shallowly and back thrusts more steeply (Figure 5.5). Therefore, frontal accretion in wedges with high basal friction is defined mainly by frontward verging thrusts and less by back thrusts because compressional shortening due to a low angle thrust is energetically more efficient (less gravitational counter force and less finite strain in the shear zone for the same amount of shortening).

Tapers obtained in these new simulations match well the analytical solution of the critical wedge theory (Figure 5.7). Surface slopes for the models with basal friction of $\varphi_b = 5^\circ$ and 10° decrease to the minimum taper at around 3 Ma. Out-of-sequence thrusts close to the rear of the models balance tapers that reach the minimum critical taper (Figure 5.4) and surface slope increases. The model with $\varphi_b = 15^\circ$ does not reach the critical minimum after 7 Ma (Figure 5.7), but is in the stable field, accreting new material at its front. This indicates that the critical wedge theory is applicable to three dimensional numerical models with a linear backstop.

Simulations with offset backstops demonstrate the influence of predefined transfer zones on the evolution of critical wedges. In models with identical frictional strength but
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Figure 5.16: Vertically averaged horizontal shear stresses after 5Ma run time. Columns for same backstop offset, rows for same basal friction. Positive shear stress values report stress rotation with clockwise rotation in map view. Dashed lines: position of the rigid indenter.
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differing length of the promontory, shapes of wedges vary laterally in terms of topography and wedge length. For low basal friction (Figure 5.9), the wedge tapers in domain A decrease in maximal elevation with increasing backstop offsets. For intermediate basal friction (Figure 5.11), elevations in domain A remain similar for different backstop offsets, whereas the average width (from rear to toe) of the wedges in front of the promontory (domain B) decrease with increasing backstop offset. Models with a strong base (Figure 5.13) show higher maximal elevation in domain A for longer backstop offsets.

The fact that lateral structural variations exist for different basal strengths demonstrates that the structural evolution along transfer zones is influenced by both the length of the backstop offset and the frictional strength of the basal décollement. Competition between lateral (horizontal shear strain; $xy$-plane) and basal (shear strain in the $xz$-plane) drag controls the wedge tapers in both parts of the wedge. The wedges in domain B exhibit a slower growth in $x$-direction than those in the adjacent domain A (Figure 5.8). This produces narrower wedges and therefore steeper taper surfaces in front of the promontory.

The effects that the offset length of the backstop and the basal friction have together can be roughly quantified by the vertically averaged horizontal shear stresses $\tau_{xy}$ at 5 Ma run time. In all simulations, positive shear stresses (clockwise-rotated in the presented map view) are largest along the wall of the indenter at $y = 50$ km (Figure 5.16). The area of large positive shear stresses ($\tau_{xy} > 5 \cdot 10^7$ Pa) increases with indenter length; maximum shear stress magnitudes increase as well (e.g., Figure 5.16g-i). These magnitudes also increase with increasing basal friction (e.g., Figure 5.16c, f and i). Overall, the length of the backstop offset has a larger influence on horizontal shear stresses than basal strength and has therefore a more pregnant influence on the structural evolution along transfer zones.

The influence of the backstop offset on the slope of the wedge surface is illustrated by plotting taper angles $\alpha$ at different $y$-coordinates for the different basal strengths (Figure 5.17). In general, surface slopes decrease from $y = 0$ km towards the transfer zone at $y = 50$ km. Wedges in domain B ($y = 50 - 100$ km) are steeper than the predicted analytical solution. Furthermore, surface slopes in front of the promontory are increased with increasing backstop offset as well as with increasing basal friction (Figure 5.17). This matches the observations made from horizontal shear stresses. The stronger the basal décollement and the longer the backstop offset, the larger are horizontal shear stresses (Figure 5.16). Horizontal shear stresses along the indenter indicate lateral drag (horizontal shear strain; $xy$-plane). This lateral drag along the indenter supports wider and flatter wedges than the analytical solution in domain A, and narrower and steeper
wedges in domain B. This consists with taper measurements of simulations with indenters (Figure 5.17).

5.5.1 Comparison to previous modelling studies

Simulations of three dimensional cylindrical setups (Figure 5.4) can be compared to previous two dimensional numerical models. Several numerical approaches have simulated thin-skinned compressional thrust wedges with brittle/plastic rheology over the last decade. In terms of sequential formation of thrust sheets, orientation of high strain rate shear bands and the authentication of the critical wedge theory, profiles of our simulations compare well to two dimensional numerical models (Buiter, 2012; Buiter et al., 2006; Ruh et al., 2012; Selzer et al., 2007; Simpson, 2011).

Transfer zones in compressional thrust wedges have mainly been investigated by analogue models (Calassou et al., 1993; Experiment 2) with the backstop offset constructed parallel to the bulk shortening direction. In these analogue models, both sides of the transfer zone developed similar tapers. This indicates that the basal strength had a larger influence than horizontal drag along the edge of the indenter. This is verified by the rather high basal friction angle (18°) compared to the internal friction angle (30°) they applied in this experiment.

Our results, like analogue models (Macedo and Marshak, 1999; Figure 12d), have shown that the initial offset in the deformation front tends to be absorbed into a linear frontal thrust orthogonal to the bulk shortening direction.

Analogue models of Reiter et al. (2011) applied lateral variations in backstop velocity instead of a rigid backstop reaching into the model domain. They employed very high basal friction angles (22.3° < ϕ_b < 29.2°) for an initial internal friction angle of ϕ = 35.4°, comparable to our model setup with a basal friction angle of ϕ_b = 15° (Figure 5.13). The main difference in topography evolution is that, in their work, the wedge in front of the faster backstop segment is larger than in front of the slow backstop (Reiter et al., 2011; Figure 8). In our simulations, the wedge parts in domain A exhibit higher elevations and are longer from rear to toe in contrast to wedges in domain B. These observations vary with different backstop offsets (Figure 5.13). The development of strike-slip shear zones in analogue simulations, but not in the 3D numerical simulations, explains this discrepancy. High strain rate shear zones along the transfer zone would release shear
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Figure 5.17: Numerically derived surface tapers over width (y-direction) for simulations with different initial backstop offsets. Circles: $L_B = 20$ km. Squares: $L_B = 40$ km. Triangles: $L_B = 80$ km. Dashed lines: analytically derived minimum critical taper. a) Basal friction angle $\phi_b = 5^\circ$ at 5 Ma runtime. b) Basal friction angle $\phi_b = 10^\circ$ at 6 Ma runtime. c) Basal friction angle $\phi_b = 15^\circ$ at 6 Ma runtime.
stresses. We contend that vertically dipping strike-slip faults would spontaneously occur in numerical simulations if we had implemented strain or strain-rate weakening.

A three dimensional numerical model with a similar geometrical setup was presented by Braun and Yamato (2010). A major difference, though, is that they induced a vertical strike-slip zone along the transfer zone by their formulation of the velocity boundary conditions. In their paper, they mainly investigated the orogenic (triangular pop-up structure) response to the surface evolution. They document a curvature of the evolving thrust sheet and of the frontal fault as reported in our results.

5.5.2 Comparison to natural examples

Results indicate that the structural and topographic evolutions of thin-skinned wedges are strongly influenced by both the backstop indentation and the strength of the basal décollement. We compare our model results to the Zagros-Makran transition zone, which separates the Makran accretionary wedge in the east from the Zagros foreland fold-and-thrust belt in the west (Peyret et al., 2009; Regard et al., 2004, 2005; Smith et al., 2005). The ∼ 300 km long dextral Minab-Zendan fault system strikes close to N-S, nearly parallel to the convergence direction between Arabia and Eurasia (Vernant et al., 2004) (Figure 5.18a). The Makran represents a submarine accretionary wedge, evolved by scraping crustal material off the subducting Arabian oceanic plate (McCall, 2002; Platt, 1990). On the other hand, the Zagros Folded belt results from the continental collision between Arabia and Eurasia (Agard et al., 2011; Stocklin, 1968). In the Makran, the major décollement layer is formed by over-pressed shale (Platt, 1990). Opposed to that, the Zagros Folded belt is detached from the basement along an initially 1 km, locally up to 2 km thick salt horizon, the Hormuz salt (Kent, 1958; Talbot, 1998). Nevertheless, we here want to show that topographic and structural features along the Zagros-Makran transition zone can be independent of the appearing transitions in décollement and sequence rheology.

At the southeastern end of the Zagros the fold belt is narrow to almost not existent, fading out towards the Zagros-Makran transition zone (Figure 5.18a). This narrowing of the fold belt coincides with a topographic low to the west of the Minab-Zendan fault. In the numerical simulation (φb = 10°, LB = 80 km at 6 Ma), we pointed out the topographic low that appears in the wedges along the promontory side. The wedge taper in domain A is not cylindrical and high elevations show a triangular zone, fading out towards the transfer zone (Figure 5.18b). Furthermore, numerically derived horizontal
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Figure 5.18: Comparison of numerical results with the Zagros-Makran transition zone. a) Elevation map of SE Iran, indicating the most important plate boundaries with the right lateral Minab-Zendan fault system in the center. White arrows: horizontal GPS velocities for a rigid Central Iranian Block (Vernant et al., 2004). Black box: location of subfigure (c). b) Elevation plot of numerical simulation with a basal friction of $\varphi = 10^\circ$ and an initial backstop length of $L_B = 80$ km after 6 Ma runtime. Black arrows: vertically averaged stresses. Dashed line: location of the slices in Figure 5.19b–d. c) Satellite picture indicating major structural strike directions. d) Horizontal slice at an elevation of 6300 m of the numerical simulation with a basal friction of $\varphi = 10^\circ$ and an initial backstop length of $L_B = 80$ km after 6 Ma run time. Plotted Lagrangian markers indicate strong bending of structure axes at the promontory corner.

Main stress directions (black arrow in Figure 5.18b) at one kilometer depth compare well with reported stress directions calculated from inversion of striation measurements (Regard et al., 2004).
CHAPTER 5. TRANSFER ZONES IN FOLD-AND-THRUST BELTS

Folds in western Makran, close to the Minab-Zendan fault zone, are strongly influenced by the dextral Minab-Zendan fault system (Figure 5.18c). They trend W-E, orthogonal to the bulk shortening direction for most of the length of Makran. In the west, folds turn to trend parallel to the N-S Minab-Zendan strike slip fault system. A similar change in structural trend occurs in our simulations. A horizontal slice ($xy$-plane), cutting at 8600 m ($z$-direction; where zero is the bottom of the Eulerian grid), shows material phases (brown and orange colors represent initially layered sequence as in Figure 5.6; Figure 5.18d). Folds are curved around the promontory corner and strike parallel to the bulk shortening direction along the transfer zone.

A roughly W-E geological profile through the northern part of the dextral Minab-Zendan fault system displays several reverse faults defining a positive flower structure (Figure 5.18a). Vertical slices through the numerical simulation presented in Figure 5.18b ($\phi_b = 10^\circ$, $L_B = 80$ km at 6 Ma) are cutting the Eulerian grid orthogonal to both the bulk...

Figure 5.19: Profile cutting the Minab-Zendan right lateral fault system (after Regard et al., 2004; Smith et al., 2005). Location is shown in Figure 5.18a. Subfigures b) to d) slices of numerical simulations with a basal friction of $\phi_b = 10^\circ$ and an initial backstop length of $L_B = 80$ km after 6 Ma runtime. From $y = 50 - 100$ km slices cut the rigid promontory. Location of profiles b) to d) shown in Figure 5.18b). b) Second invariant of the strain rate tensor. c) Vertical velocity, upward movement. d) Horizontal velocity in $x$-direction, motion toward the backstop negative.
shortening direction and the transfer zone. The second invariant of the strain rate tensor (Figure 5.19b) exhibits a flat shear zone that roots into the basal décollement right below the predefined transfer zone ($y = 50$ km). A topographic high appears atop the fault root. The topography in the numerical model and the flat thrust geometrically resemble the geological profile. Vertical velocities (Figure 5.19c) indicate that the hanging wall is uplifted. Plotted horizontal velocities (Figure 5.19d) parallel to the bulk shortening ($x$-direction) reveal a right lateral movement along this fault. According to the plotted velocities, the low-angle shear zone acts as an oblique thrust. This is also reported along the Minab fault (Regard et al., 2004) (Figure 5.19a).

The models presented here simulate general features of the Zagros-Makran transition zone, especially in terms of surface topography and thrust orientation. We acknowledge that differences in décollement type (salt in Zagros, shale in Makran) may also influence the overall mechanics along this transfer zone. However, this study shows that first-order features such as surface topography and thrust orientation may be more independent of décollement type and/or variations in rock sequence than advocated by some authors.

5.6 Conclusion

A high-resolution 3D finite difference numerical model with a visco-plastic rheology was created and employed to investigate how the frictional strength of a décollement affects the development of thin-skinned compressional fold-and-thrust belts along transfer zones. Compression of sedimentary rocks above a frictional décollement leads to forward, in-sequence thrusting, forming ramps with dip angles depending on the within-wedge stress directions, which can be analytically derived. Perfectly cylindrical simulations correlate well with the critical wedge theory.

An offset backstop was simulated with a rigid, rectangular promontory added to the cylindrical setup. This promontory strongly influences the evolution of the thin-skinned brittle wedge. Wedge tapers in front of the promontory tend to exceed the analytically derived tapers. Surface slopes in domain A, where no promontory exists, are lower than the analytical prediction. In this domain, single thrust zones exhibit less offset, which leads to more scattered structural patterns. Wedges in front of the promontory structurally differ from the wedges that form where no promontory is present. The wedges pushed by the promontory usually form steeper surface slopes and build thrust sheets with large fault offsets. The impact of the transition zone on the surface tapers of
both simulation domains, the one pushed by the promontory and the other by the grid boundary, increases with increasing basal décollement strength.

These new 3D numerical simulations explain the occurrence of folds parallel to the shortening direction as observed in the Zagros-Makran transfer/transform zone. The low angle faulting with a lateral component similar to the Minab fault was reproduced. Importantly, our simple geometrical setup has shown how topographic lows like that observed north of the Strait of Hormuz, in the Zagros, can be produced in compressional settings. Such lows (intramontane basins) are indicative for transfer zones in particular.

Acknowledgement

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Towards 4D modelling of orogenic belts: example from the transpressive Zagros Fold Belt

Jonas Ruh, Taras Gerya & Jean-Pierre Burg

Abstract

Latest developments in computer power and computational solutions open new ways to envision complex natural systems. Advances in high-resolution 3D geodynamic modelling, in particular, allow investigating tectonic processes like transpression, which involve “normal” and “tangential” kinematics with respect to the bulk orogenic trends. We demonstrate that implementing the 4 dimensions (space and time) provides and constrains new answers to long lasting discussions. As case study, we have chosen the Zagros Simply Folded Belt in Iran, which exhibits spatial variations of structural patterns from the homogeneously shortened, wide fold belt in the Fars to the narrower, higher elevated area in the Izeh. High-resolution, 4D numerical modelling shows how a low-viscosity décollement becoming frictional towards the Dezful embayment influences the exposed fold patterns. Results also emphasize the importance of an oblique backstop to produce en-échelon arranged folds where convergence-related and backstop-controlled folds are mingled in the transpressional orogen.
6.1 Introduction

Mechanical modelling is an important tool to understand the dynamics and investigate the characteristics of compressional orogenic wedges. Analogue and laboratory models implemented for more than a century (Cadell, 1890) are three dimensional and scaled by using low-strength materials, such as sand, mica and silicone putty to simulate tectonic processes (Graveleau et al., 2012). In the last decades, fully mechanical numerical codes have been developed to mimic viscous, elastic and brittle/plastic deformation (Buiter et al., 2006; Gerya and Yuen, 2007; Kaus, 2010). As an advantage on analogue models, numerical models are able to introduce complex initial geometries and a wide range of material properties. Additionally, model parameters such as velocity, stress and strain can be directly measured from the numerical solution (Buiter, 2012). Due to the relatively high resolution needed to accurately simulate plastic deformation, most models of brittle thrust wedges have been two dimensional. Nevertheless, they gave important new insight into the dynamics and mechanics of evolving compressional wedges and refined tectonic interpretations. For example, the mode of deformation in a compressional regime reflects the competition of two fundamental instabilities, folding and faulting, depending on parameters like elastic moduli, viscosity and thickness ratio of the overburden and décollement (Simpson, 2009). Additional weak layers within the brittle/plastic wedge promote folding as the dominant mode of deformation (Yamato et al., 2011). Furthermore, the structural evolution of compressional wedges depends on the décollement rheology (Ruh et al., 2012). A low-viscous basal décollement holds high strain rates, even at low stresses, and supports folding, whereas a frictional décollement leads to piggy-back imbricate fans.

Even though these results help understanding geological and geophysical processes, they reach their limit when it comes to three dimensional geometrical problems such as lateral changes in material properties or non-cylindrical backstop geometries. Although three dimensional numerical models including brittle/plastic rheology exist (Allken et al., 2011; Braun and Yamato, 2010; Gerya, 2010a; Popov and Sobolev, 2008), it was not yet possible to simulate compressional thrust wedges on geological scales. To produce narrow shear bands, indicating thrusts, high spatial and temporal resolution is needed. We present the first high-resolution three-dimensional visco-plastic finite difference model suitable to simulate thrust wedges.
6.1. INTRODUCTION

We chose a setup with three-dimensional geometrical and rheological complexity to illustrate the efficiency of the new numerical code. The Zagros fold-and-thrust belt, SW Iran (Figure 6.1), is ideal for this purpose. It results from post-collision shortening between the Arabian plate and the Iranian continental blocks and is therefore related to bulk compressional tectonics (Stocklin, 1968). The recent shortening direction is oblique to the Main Zagros Thrust, which represents the suture between the Arabian and Central Iranian continental plates (Figure 6.1). This obliquity impels lateral backstop variations that can only be introduced in a three dimensional setup. The Zagros fold-and-thrust belt is divided from the NE to SW into the Thrust Zone, the Imbricate Zone and the Simply Folded Zone (Falcon, 1974). Our model setup concentrates on the Simply Folded Zone (SFZ), which is characterized by along-strike structural variations. In the Fars, southeast of the Kazerun and Borazjan strike slip faults (Figure 6.1), the SFZ is up to 250 km wide with mostly symmetrical open folds with about
16 km wavelength (Mouthereau et al., 2007b). The famous en-échelon, “whale-back” shaped anticlines, prominent on satellite images, characterize this area. In the Fars, the deformation front of the Zagros fold-and-thrust belt is the south-western limit of the SFZ (Figure 6.1). To the northwest of the Kazerun and Borazjan faults, in the Izeh domain (Bahroudi and Koyi, 2003), the SFZ is much narrower and exhibits a steeper surface taper than in the Fars (McQuarrie, 2004). Furthermore, the deformation front of the SFZ in the Izeh domain is the Mountain Front Fault, in front of the up to 200 km wide (in profile), low topography Zagros Foredeep (Figure 6.1). The topographic and structural differences between the Fars and Izeh regions have been attributed to differences in basal décollement. In the Fars, the major décollement is predominantly formed by Hormuz salt (Letouzey et al., 1995). In the Izeh, the décollement changes from the SFZ to the Foredeep Zone. In the northeast, the spatial distribution of salt diapirs suggests that the narrow SFZ detaches along the Hormuz salt (Bahroudi and Koyi, 2003; Kent, 1979). In the southwest, the Foredeep Zone is geographically overlapping with the Dezful embayment (Allen and Talebian, 2011). Folds in this area either detach on shale equivalent to the Hormuz series in the Fars (McQuarrie, 2004) or along evaporitic formations higher up in the stratigraphy (Allen and Talebian, 2011; Sherkati et al., 2005).

We decided to test these hypotheses. High resolution, 4D modelling directly shows how the structural style primarily results from lateral variations of the major décollement and, in addition, demonstrates the up-to-now overlooked importance of oblique convergence (transpression). We illustrate numerically (i) the effect of a viscous décollement within the Hormuz salt pinched-out towards the foredeep of the Izeh domain and (ii) the influence of the backstop orientation on the structural pattern of the fold-and-thrust belt.

6.2 Numerical model

To investigate the influence of an oblique backstop and the lateral variation of décollement rheology in a thin-skinned fold-and-thrust belt, we developed a three-dimensional, high-resolution, fully staggered grid, finite difference, marker in cell model with a visco-brittle/plastic rheology and an efficient OpenMP-parallelized multigrid (Gerya, 2010b; Gerya and Yuen, 2007). The mechanical model is built on the equations for conservation of mass for an incompressible case and the conservation of momentum, the Stokes equa-
Plasticity is calculated on the markers and is activated if the second invariant of the stress tensor exceeds its yield stress $\sigma_y$. Marker viscosities are decreased in order to bring the second invariant of the stress tensors down to the yield stresses of the markers. Simulations run for 6 Ma with a time step of 1 ka.

The model size of all simulations in this study is $150 \cdot 150 \cdot 15$ km ($x$, $y$, $z$-direction) with a resolution of $309 \cdot 101 \cdot 101$ nodal points (Figure 6.2a). This results in a resolution of 3050701 nodes with around 24.5 million markers. The numerical model has been set up to mimic the geometric and dynamic properties of the Zagros. Initially, the upper 7.5 km of the model are defined by “sticky-air” (Crameri et al., 2012), with a low viscosity of 1018 Pa·s and a density of 1 kg/m$^3$. The lower 7.5 km are built up by a lowermost, 500 m thick strong layer, which moves through the Eulerian grid like a rigid basal plate. On top lays a 7 km thick cover sequence underlain by a 1 km thick purely viscous décollement, comparable to stratigraphic thicknesses in the Zagros (Colmanskopf, 1978; Kent, 1970; Koop and Stoneley, 1982; Stocklin, 1968). Similar to Yamato et al. (2011), we introduces several weak horizons within an otherwise visco-plastic rock pile (Figure 6.2), which should mimic evaporites and shale horizons within the otherwise strong cover sequence as reported from the Zagros (Mouthereau et al., 2007b; Sherkati et al., 2005).

The modeled brittle rocks have an initial viscosity of $10^{23}$ Pa·s with a friction angle of 30° and a cohesion of 20 MPa and a density of 2700 kg/m$^3$. A low viscosity of $10^{19}$ Pa·s and a slightly lower density of 2200 kg/m$^3$ define the viscous basal “salt”. The weak “evaporite” layers within the cover sequence have a viscosity of $10^{20}$ Pa·s. Both basal and intermediate viscosities are consistent with published values for salt (Mukherjee et al., 2010). Due to a lack of Hormuz salt below the Dezful embayment, a frictional major décollement is introduced between 0 – 50 km in $y$-direction, with 50 km distance to the backstop (Figure 6.2a). The strength of the frictional décollement is given by its friction angle (10°) and its cohesion (20 MPa). The main basal décollement has an initial thickness of 1 km, intermediate weak layers are either 300 m or 500 m thick (Figure 6.2b).

To investigate the effect of a shortening velocity oblique to a rigid backstop, we introduced a triangular rigid backstop, which has a viscosity of $10^{23}$ Pa·s and no plasticity. The backstop has a height of 10 km and linearly increases in length ($x$-direction), with increasing width ($y$-direction), from 1.5 to 90 km (Figure 6.2a), imposing the same angle between shortening direction and backstop boundary as in Zagros (Figure 6.1). At the right side as well as at the bottom, a velocity of 1 cm/a in $x$-direction is applied, consistent with horizontal GPS velocities measured in this area (Vernant et al., 2004).
These boundary conditions are comparable to analogue model setups, where a plastic sheet lying below sand layers is pulled out below a fixed and rigid backstop (Konstantinovskaya and Malavieille, 2011). At the left side, all velocities are set to zero, except for the basal 500 m thick rigid “plate”, which is pulled out with 1 cm/a and a gradational velocity decrease throughout the lowermost weak horizon (Figure 6.2). This setup also matches the boundary conditions used in the analytical critical wedge theory (Dahlen, 1984; Davis et al., 1983).

A simple two-dimensional diffusive surface process is implemented to address sedimentation in developing depressions. The surface between rocks and “sticky-air” is interpolated from the marker density contrast for every column in the three-dimensional grid, to have an exact topography, independent of the nodal resolution. Diffusion of this resulting surface is curvature dependent:

\[
\frac{\partial h_s}{\partial t} = \kappa \cdot \frac{\partial^2 h_s}{\partial x_i^2}
\]

where \(\kappa\) is the diffusion constant, \(h_s\) the surface topography and \(x_i\) the spatial coordinates \((x_1 = x, x_2 = y)\). To address a depositional process without erosion, markers with a “sticky-air” rheology below the updated surface are transmuted to markers with a rock rheology. Rock markers above the updated surface are not transmuted and therefore, no erosion is applied. This simulates a submarine depositional environment.
6.3 Application of the model to the Zagros Folded Belt

Illustrations of the topography evolving through time show that, like in analogue and 2D numerical models, deformation starts along and in front of the backstop. Ongoing shortening expands the wedge “in sequence”, i.e., the deformation front progressively migrates away from the backstop so that the fold-and-thrust belt grows in profile width (Figure 6.3). Two types of fold orientation imprint the topography plots: (i) folds tending parallel to the oblique backstop and (ii) folds initially trending orthogonal to the bulk shortening direction and occasionally rotating towards the backstop orientation (Figure 6.3c). En-échelon fold arrays result from interaction between these two fold types and therefore express the competition of backstop- and convergence-induced folding. The topographic morphology of the model varies markedly according to the basal décollement. In the left third of the model, where the viscous décollement pinches out into a frictional material, a narrower but higher altitude zone develops (Figure 6.3c). In the right part of model, a wider and lower wedge is formed.

The evolution of the surface tapers of two profiles through these two parts of the model can be related to the critical wedge theory, an analytical solution determining a total taper in function of the internal and basal strength of a wedge (Dahlen, 1984; Davis and Engelder, 1985; Davis et al., 1983). Theoretically, a wedge with a surface taper below its critical minimum deforms at its rear to steepen. If the surface angle is above the critical minimum, the wedge can be considered stable. In this case, it slides along its décollement and accretes new material at its front. Due to our lateral variation in décollement rheology, we have to differ between two cases. Rock salt is ductile (Carter and Hansen, 1983) and its resistance to shear depends on its viscosity (Turcotte and Schubert, 2002), whereas the resistance of the frictional base is defined by its plastic strength (Kopf and Brown, 2003; Takahashi et al., 2007; Wang et al., 1980). The analytically derived minimum critical taper angle for a wedge with an internal friction angle of $\varphi = 30^\circ$ overlying a viscous “salt” décollement with a viscosity of $\eta = 10^{19}$ Pa·s is $0.51^\circ$. The resulting taper related to the frictional “shale-type” décollement with a basal friction angle of $\varphi_b = 10^\circ$ is $3.4^\circ$. These analytical values match the observed surface taper variation between the Fars and Izeh domains (McQuarrie, 2004; Mounthreau et al., 2006).

The surface taper in the left of the simulation depends on the strength of the frictional layer and not on the viscous material, as in the right part (Figure 6.4). This lateral
Figure 6.3: Evolution of the topographic elevation through time. a) Left third: dominant folding along the pinch-out viscous décollement. Lower right corner: folding orthogonal to shortening. b) Right part: widening of the wedge in contrast to the left part. c) Left third: high elevations in a narrow folding zone. Fold axes are both parallel to the backstop and orthogonal to the shortening direction. Dashed lines: profiles A and B in Figure 6.4. Black box: Figure 6.6b.
change in décollement dependency has an important effect on the stress state within the folded sequence. Stresses diffuse very far through a wedge above a viscous décollement, because low viscosity layers undergo high strain rates even for very low shear stresses (Ruh et al., 2012). Rocks above a frictional décollement do not deform as long as basal stresses do not overcome the yield stress of the décollements rock, which is naturally much larger than stresses in low viscous layers.

Several observations on plots of modeled elevation are verified by slices cutting the corresponding fold-and-thrust wedge (Figure 6.5). If the folded sequence lies over a viscous décollement ($y$-direction: 50 – 150 km), shortening is dominantly accommodated by symmetrical folding with a wavelength proportional to the stratigraphic thickness. Some folds become asymmetric at depth, where axial planes dip towards the backstop (Figure 6.5a; slices 3 – 5). Where the basal viscous layer disappears ($y$-direction: 0 – 50 km), foreland-verging faults sole the folded sequence. Thrust sheets of the lowermost part of the sequence are stacked above each other illustrating underthrusting of the bottom levels below tight folds with a similar wavelength as the folds in slices 3 – 5. The axial plane of the anticline at the rear of the model dips toward the wedge front (Figure 6.5a; slices 1 – 2). The second invariant of the stress tensor reveals that the frontal anticline evolves into a thrust rooting in the viscous pinch-out zone (Figure 6.5b; slices 1 – 2). This thrust is comparable to the Mountain Front Fault of the Izeh domain, between the tightly folded SFZ and the low topography Foredeep Zone (Figure 6.1). In
Figure 6.5: Profiles 1, 2, 3, 4, and 5 in x-direction at y = 5, 40, 75, 110 and 145 km after a simulation time of 3.4 Ma. a) Illustration of the rock type defined on markers. Profiles 1 and 2: appearance of the pinch-out zone. Underthrusting leads to high topography. Profiles 3 - 5: salt exclusively forms the décollement. Folding is the dominant mode of deformation. b) Second invariant of strain rate tensor. Profiles 1 and 2: major thrust rooting in the basal décollement. Frictional décollement is partly inactive. Activation of intermediate weak layer as décollement shortens upper parts of the cover sequence. Profiles 3 - 5: basal low viscosity décollement is active.

slices 1 and 2, the model deformation front is not this thrust since the cover sequence further away from the backstop is also folded.

High strain rates within the viscous décollement horizons denote its activity (Figure 6.5; slices 3 – 5). Low strain rates indicate reduced activity where the décollement becomes frictional (Figure 6.5b; slices 1 – 2). This fact has important effects on the
6.4 Style of folding induced by oblique convergence

Besides the lateral décollement variation, our simulation reveals the importance of the backstop obliquity on the evolution of the fold pattern in the SFZ. Unlike cylindrical folding in front of a backstop orthogonal to the shortening direction, as implied in 2D modelling, 3D models are able to produce variations in fold axis lengths and orientation. Although models concerning the evolution of 3D folding exist (Schmalholz, 2008; Schmid et al., 2008), they do not address the importance of oblique convergence. The growth and linkage of elongated folds with an offset as observed in the Zagros has been examined in a previous model (Grasemann and Schmalholz, 2012). Nevertheless, this previous study could not explain the spatial initiation of folds because it placed seeds to prescribe the fold orientation.

The comparison between elevation plots of the numerical model (Figure 6.3) and the elevation map of the SFZ (Figure 6.1) uncovers the strong correlation of shapes and orientation of folds. As the fold wavelength in brittle/plastic layers depends upon the thickness of the folded sequence (Ruh et al., 2012), the models demonstrate that a 7 km thick sequence (including the décollement) is appropriate for the SFZ. The style of fold linkage in form of an en-échelon arrangement of anticlines in the Fars domain (Figure 6.6a) is spontaneously produced in our 3D numerical simulations (Figure 6.6b). The lengths of the fold axes scale well with nature. The axial length of en-échelon folds depends on both the direction of shortening and the backstop orientation. Folds with an axial orientation parallel to the backstop occur close to the backstop and are cylindrical. Folds with shorter axial lengths are always oblique to the backstop (Figure 2a-c). A straightforward scaling law inferred from model observation for the relation of the axial length LA to the short wavelength LW of folds can be described with respect to the

model evolution: (i) the inactivation of the frictional décollement inhibits shortening of the bottom part of the overlying sequence and leads to strong compression at the rear, where the décollements is viscous. (ii) Furthermore, where the basal décollement is practically inactive, the next upper intermediate weak layer is activated as main décollement for the upper layers. This shallower décollement causes shorter wavelength of folds, according to the sequence thickness, and relatively low topography. This low topography is comparable to the morphology in the Foredeep Zone.
Figure 6.6: Comparison model/natural observations.

a) Elevation map of en-échelon anticlines in the Zagros (black box in Figure 6.1). b) En-échelon structures from elevation plots of 3D model (black box in Figure 6.3c). Both images show similar axial length of folds (see scale bars). White line: backstop orientation. White arrow: shortening direction. c) Sketch of length of fold axes and en-échelon arrangement of folds depending on the angle $\theta$ between their long axis (straight lines) and the backstop orientation (dashed lines).
angle \( \theta \) between the fold axis and the backstop orientation:

\[
\frac{L_A}{L_W} = \frac{1}{\sin \theta}
\]  

(6.2)

According to this law, folds parallel to the backstop (\( \theta = 0^\circ \)) show an infinite axial length \( L_A \) and are therefore cylindrical. Obliquity of fold axis implies shorter fold length and the development of en-échelon type fold arrangements (Figure 6.6c). For \( \theta = 30^\circ \), the fold axis is orthogonal to the shortening direction in Zagros. The \( \frac{L_A}{L_W} \) ratio is 2. Folds in the SFZ and in the numerical simulation trend between \( \theta = 0 \) and \( 30^\circ \). Nevertheless, the type of folds initially directed orthogonal to the shortening direction (Figure 6.3a-c; lower right corner) are not applicable to the scaling law, as they are not primarily influenced by the backstop orientation but instead by the shortening direction at the velocity boundary.

### 6.5 Conclusion

The observed correlations between our simulations and the SFZ of the Zagros orogen in terms of structural variations between the Izeh and the Fars domains underline the rheological difference of their respective décollements. Applied values of \( \eta = 10^{19} \text{ Pa·s} \) for the viscous and \( \varphi_b = 10^\circ \) for the frictional décollement very well constrain analytical as well as natural taper angles reported for the Zagros. Above the Dezful embayment, intermediate weak layers host the active décollement due to the relative weakness of these evaporites to the lower frictional base. We also showed that en-échelon folds in the Fars are determined by the obliquity of the backstop. 3D modelling shows that this is due to the application of a bottom velocity boundary condition that drags the cover sequence towards the backstop. This strongly supports that the basement below the Fars is not shortened by the same amount as the SFZ and that it is pulled out northeastward, below the SFZ. This conclusion could not have been predicted without quantitative, 3D modelling of structures abiding by physical rules such as the rock rheology.
3D effects of strain vs. velocity weakening on deformation patterns in accretionary wedges

Jonas Ruh, Taras Gerya & Jean-Pierre Burg

Abstract

Shear zones are weaker than surrounding rocks. Whether weakening depends on accumulated strain or strain-rate is largely debated. We used a three-dimensional numerical code with a visco-plastic/brittle rheology to investigate the influence of strain and strain-rate (velocity) weakening on the evolution of thin-skinned fold-and-thrust belts. Two model setups are studied: (i) A uniform setup to recognize the effects of each weakening mode on the structural evolution and dynamics of fold-and-thrust belts and (ii) A dual setup with two adjacent domains of décollement strength to investigate the structural response of laterally varying systems. Results show that strain weakening and velocity weakening lead to remarkably different structural patterns. Especially, strain weakening favours out-of-sequence shear zones whilst some amount of finite plastic strain has to be achieved before the structural development initiates weakening. In contrast, velocity weakening starts with the onset of high strain-rate bands and does not favour out-of-sequence deformation. We also tested separately the influences of shortening velocity, cover sequence thickness and weakening style and amount on the structural evolution of the dual systems. The shortening velocity has a minor effect in strain-weakened models. Velocity weakening and a thinner cover sequence enhance the occurrence of strike-slip shear zones. We conclude that the weakening mode strongly influences the fault patterns and the dynamics of thrust wedges. Inversely, natural fault patterns are keys to discriminate weakening processes.

This chapter is under review in Tectonophysics
CHAPTER 7. 3D EFFECTS OF STRAIN VS. VELOCITY WEAKENING

7.1 Introduction

Numerical modelling of accretionary wedges and thin-skinned fold-and-thrust belts has been an intense field of investigation using different techniques in two (e.g., Buiter, 2012; Buiter et al., 2006; Burbidge and Braun, 2002; Fillon et al., 2012; Miyakawa et al., 2010; Ruh et al., 2012; Simpson, 2011; Stockmal et al., 2007) and, more recently, in three dimensions (Braun and Yamato, 2010; Ruh et al., 2013). The simplest and most-often applied setup, like sandbox models (e.g., Konstantinovskaya and Malavieille, 2011), consists of a rectangular box that contains a thick layer of frictional material overlying a relatively weak and thin décollement layer. The bottom boundary is pulled out below the rigid walls, one of which representing the fixed backstop. This setup simulates a sequence of upper crustal rocks scraped off a subducting plate by the rigid backstop of a converging plate boundary. The advantages of this simple setup is (i) the comparability between models obtained with different numerical methods and/or codes, and (ii) its relevance to the analytical critical wedge theory (Dahlen, 1984; Dahlen et al., 1984; Davis and Engelder, 1985; Davis et al., 1983).

Mechanical studies of compressional thin-skinned fold-and-thrust belts often focus on the influence of décollement strength (Burbidge and Braun, 2002; Ruh et al., 2012), surface processes (Fillon et al., 2012; Selzer et al., 2007; Simpson, 2006), rheology of cover sequence and/or décollement (Ruh et al., 2012; Simpson, 2009), appearance of internal weak layers/inclusions (Ruh et al., 2012; Selzer et al., 2007; Stockmal et al., 2007; Yamato et al., 2011) and material cohesion (Nilfouroushan et al., 2012). Many of these studies applied a routine to weaken the modelled upper crustal layer, usually by decreasing the friction angle and/or cohesion of the corresponding plastic/brittle layer as a function of plastic finite strain (strain weakening). Few studies tested the effects of mechanical weakening on numerical wedges (e.g., Buiter et al., 2006; Ellis et al., 2004; Nilfouroushan et al., 2012; Selzer et al., 2007). According to Ellis et al. (2004), the amount of strain weakening and the strain threshold values above which weakening appears control the width, the abundance and the temporal development of shear zones. Selzer et al. (2007) supported this statement and showed that single shear zones due to strain weakening accommodate more displacement relative to shear zones in models without strain weakening.

In nature, several processes contribute to dynamic weakening of the lithosphere and to localization of high strain-rate into narrow shear bands. These processes have been
studied analytically as well as with analogue and numerical techniques. Examples for weakening processes are dynamic recrystallization (e.g., Rutter, 1999), shear heating (e.g., Regenauer-Lieb et al., 2008; Thielmann and Kaus, 2012) and partial melting (e.g., Tumarkina et al., 2011), hydration metamorphism (e.g., Escartin et al., 1997), and grain-damage and -pinning (Bercovici and Ricard, 2012). These processes are essentially pertinent for deep crustal and mantle dynamics. They are implemented with non-physical routines, which are depending on finite strain and/or strain-rate (e.g., Alkken et al., 2012; Lavier et al., 2000; Montesi and Zuber, 2003; Tackley, 2000). However, accretionary wedges and fold-and-thrust belts are for the most parts within the brittle regime of the crust, where modes of viscous weakening might not apply.

Two weakening processes are considered in upper crustal rocks: (i) lubrication of faults and shear zones and (ii) seismic dynamics. Evidence for fluid involvement in faulting and the consequent weakening effects has been argued in a wealth of studies (Hickman et al., 1995, and references therein; Stewart et al., 2000). It is well known that an increase in fluid pressure reduces the yield strength of brittle rocks (Fischer and Paterson, 1989). This is put forward for major décollements of accretionary wedges, where water-saturated shales form relatively weak horizons. High pore pressure may have different origins. On the one hand, layers containing pore fluids may be roofed and buried below impermeable sediments. With increasing overload, the pore space reduces as the dynamic pressure increases. Because the content in pore fluid remains constant, fluid pressure increases and weakens the rock. On the other hand, fluids can be released by diagenetic and metamorphic reactions such as the smectite to illite transition, which appears $\sim 2.5$ – $5$ km depth in clay-rich accretionary complexes and can account for up to 80 percent of the pore fluid (Bekins et al., 1994; Lanson et al., 2009). This input of additional fluid into an already water-saturated rock leads to fluid over-pressure and a decrease in rock strength. This would explain the low strength of deep décollement layers. The overlying rocks, though, have a lower pore-fluid pressure and are initially stronger than the basal décollement. During shortening of the sedimentary complex, brittle deformation develops fractures along which fluids can migrate through the formerly impermeable seal (Grando and McClay, 2007). The fracture is weakened by the locally increased fluid-pressure as soon as fluids infiltrate it. Mud volcanoes are exit ways and geological evidence of such over-pressured fluids escaping the accretionary system (e.g., Schluter et al., 2002). Since brittle failure occurs beyond a plastic threshold, fluids in faults can be considered to be strain-weakening components.
One of the main components, among others, to model long-term seismic processes is rate-dependent friction (e.g., Wang, 2007). In earthquake research, the dependence of rate (velocity) on friction has been discussed for several decades (Burridge and Knopoff, 1967; Dieterich, 1979; Ruina, 1983). Co-seismic slip-rate dependency of friction has been documented by rotary experiments (e.g., Di Toro et al., 2011) as well as in continuum visco-elasto-plastic numerical models (vanDinther et al., 2013).

Lubrication and seismic weakening are both physically complicated and influenced by several processes. They can be differentiated by the manner frictional strength is weakened. In the case of fault lubrication, the rock strength is reduced by accumulated plastic strain. Fluids infiltrate a fault as soon as it appears (plastic strain increases), reduce the dynamic pressure of the infiltrated rock and thus lowers its plastic yield strength. At variance, the seismic yield strength is rate dependent.

We present different series of simulations to investigate the effects of strain vs. velocity weakening. Then, high resolution, three-dimensional models allow testing how structural coherence is maintained between two adjacent plastic/brittle wedges depending on different weakening properties. Models with a change in décollement strength were performed to understand three-dimensional effects of the different weakening styles. Especially, the evolution of transfer components, i.e. frontal fault bending or strike-slip faults, are investigated and compared to analogue modelling studies, where the development of transfer zones is also under intense debate (e.g., Bahroudi and Koyi, 2003; Calassou et al., 1993; Macedo and Marshak, 1999; Reiter et al., 2011). We evaluate our results in the context of the analytical critical wedge theory and shortly present their relevance to geological examples of fold-and-thrust belts.

7.2 Model setup

The numerical model is similar to that documented and used in an earlier study (Ruh et al., 2013). It is based on a three-dimensional, high-resolution, fully staggered grid, finite difference, marker in cell model with a standard visco-brittle/plastic rheology and an efficient OpenMP-parallelized multigrid solver (Gerya, 2010b; Gerya and Yuen, 2007). Governing equations and their numerical implementation are specified in appendix 7.A.
7.2. MODEL SETUP

7.2.1 Geometrical setup and material properties

Model setups of all simulations are constructed on Eulerian grids dimensioned to account for different requirements: vertical thickening of the wedge, possible lateral structural changes throughout the evolving wedge and increasing length of the wedge while the wedge front migrates. The planar backstop is defined by a vertical grid boundary with zero horizontal velocity. High resolution allows the development of high strain-rate shear bands sufficiently narrow to be comparable to natural fault systems. We present two series of simulations. The first series uses a geometrically uniform setup (series U) with geometry and rheology being equal over the whole model (Figure 7.1a). Small but negligible disparities in results are due to the initially random distribution of Lagrangian markers within the Eulerian grid. These models were performed to compare results with (i) the two dimensional critical wedge theory (Dahlen, 1984; Dahlen et al., 1984; Davis et al., 1983) and (ii) earlier, two-dimensional numerical studies (e.g., Ellis et al., 2004; Selzer et al., 2007). Series U shows the structural and dynamic repercussions of strain vs. velocity weakening. The second series incorporates two halves with different décollement strength (series D). The sharp, vertical boundary at \( L_y = 50 \) km (Figure 7.1b) is contained in the décollement layer, parallel to the shortening direction (x-axis).

Eulerian grid dimensions of series U models are 150-50-20 km in \( x-, y-, z- \) directions with a nodal resolution of 309-85-149, respectively; the resulting cell size is 487/595.2/135.1 m in \( x-, y-, z- \) directions (Figure 7.1a). Series D models are expected to structurally vary laterally (\( y \)-direction) and therefore have a wider setup. The dimension of the Eulerian

![Figure 7.1: Model setup. Black: 500 m thick rigid plate entering and exiting the Eulerian grid. Light grey: 500 m thick frictional décollement layer. Grey: brittle/plastic cover sequence with thickness \( H \). Bottom left: treatment of velocity at the backstop boundary. a) Uniform models (series U). \( H = 5 \) km, \( V_x = 1 \) cm/a. b) Laterally varying simulations (series D). \( H = 2.5 \) or 5 km, \( V_x = 1, 2 \) or 5 cm/a. Décollement strength varies over width. \( y = 0 - 50 \) km: \( \phi_b = 15^\circ \). \( y = 50 - 100 \) km: \( \phi_b = 5^\circ \).](image)
Table 7.1: Parameters Definition

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Meaning</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\phi_0$</td>
<td>initial friction angle</td>
</tr>
<tr>
<td>$\phi_w$</td>
<td>weakened friction angle</td>
</tr>
<tr>
<td>$C_0$</td>
<td>initial cohesion</td>
</tr>
<tr>
<td>$C_w$</td>
<td>weakened cohesion</td>
</tr>
<tr>
<td>$\varepsilon_w$</td>
<td>weakening strain</td>
</tr>
<tr>
<td>$\dot{\varepsilon}_w$</td>
<td>weakening strain-rate</td>
</tr>
<tr>
<td>$\phi_b$</td>
<td>décollement friction angle</td>
</tr>
<tr>
<td>$\kappa$</td>
<td>diffusion constant</td>
</tr>
<tr>
<td>$H$</td>
<td>rock sequence thickness</td>
</tr>
<tr>
<td>$V_x$</td>
<td>shortening velocity</td>
</tr>
</tbody>
</table>

The grid is 100·100·20 km with a resolution of 245·245·149 in $x$-, $y$-, $z$-directions, respectively. The resulting cell size is 409.8/409.8/135.1 m in $x$-, $y$- and $z$-directions (Figure 7.1b). Lagrangian marker resolution for both series is two markers per cell in every direction. This setting results in a resolution of 8.9 million nodes with about 70 million Lagrangian markers for models with highest resolution.

Material properties are initially defined on the Lagrangian markers, which are advecting through the Eulerian grid. From bottom to top of the Eulerian grid, marker types are distributed as follows: A 500 m thick rigid plate is the bottom. Above this rigid plate, the 500 m thick décollement horizon with a frictional rheology soles a “sedimentary” layer of thickness $H$ (Figure 7.1). Density is 2700 kg/m$^3$ for the lower rigid plate, the décollement and the sediment layer in all models of this study. Lower and upper cutoff viscosities are $10^{18}$ Pa·s and $10^{23}$ Pa·s, respectively. Other material parameters are listed in Table 7.1. Simulation parameters of series U are given in Table 7.2, parameters of series D in Table 7.3.

All Lagrangian markers above the plastic “sediments” are defined as “sticky-air”, an approach that mimics a quasi-free surface (Crameri et al., 2012; Gerya and Yuen, 2003; Schmeling et al., 2008; Zaleski and Julien, 1992). The low density (1 kg/m$^3$) and small viscosity ($10^{18}$ Pa·s) of the sticky-air layer ensures sufficiently small normal stresses at the sediment surface. Analytical predictions have been made for a similar setup to test
the quality of this free-surface approach (Ruh et al., 2013). They showed that it fulfills the general condition for a traction-free surface.

### 7.2.2 Boundary conditions

A velocity $V_x$ is defined at the lower ($z = 0$ km) and frontal boundaries ($x = L_x$; Figure 7.1), which is equivalent to pulling out the bottom sheet of a sand box model below a rigid backstop (e.g., Konstantinovskaya and Malavieille, 2011). New Lagrangian marker layers coming in through the front side allow large deformation. The back side of the setup ($x = 0$ km) acts as backstop with a no-slip boundary condition. At the bottom, the 500 m thick rigid plate of high constant viscosity is included; it allows a better stability of the multigrid solver at the lower model boundary. Between the bottom rigid plate and the “backstop”, the 500 m thick décollement undergoes simple shear ($V_x$ at $z = 500$ m to 0 cm/a at $z = 1000$ m; Figure 7.1). Both lateral boundary conditions (at $y = 0$ km and $y = L_y$) are defined as free slip along the boundary planes. An upward velocity is applied at the upper boundary. This balances the volume of exiting “sticky-air” markers and the volume of incoming material at the front boundary to fulfill the conservation of volume.

### 7.2.3 Strain weakening

Strain induced by plastic deformation is accumulated every time step and stored on every marker. Strain weakening is modelled as a linear decrease of frictional angle and/or cohesion between accumulated plastic strains of $\varepsilon_{w0}^0 = 0.5$ and $\varepsilon_{w1}^1 = 1.5$ (Figure 7.2a). These lower and upper thresholds of strain for weakening activation and completion are similar to those in previous numerical investigations of strain weakening of crustal rocks (e.g., Allken et al., 2012; Huismans et al., 2005; Selzer et al., 2007; Stockmal et al., 2007). How the amount of strain weakening ($\varphi_{w}$) influences the structural and the overall wedge shape is investigated in U series, uniform simulations. Series D models with strain weakening focussed on the relationships between the amount of strain weakening ($\varphi_{w}$), the thickness of “sediment” ($H$) and the shortening velocity ($V_x$). Our code has no healing term as in other codes (e.g., Gerya, 2013; Tackley, 2000), so that weakened material does not strengthen again.
<table>
<thead>
<tr>
<th>Simulation</th>
<th>$\varphi_0$ (°)</th>
<th>$\varphi_w$ (°)</th>
<th>$C_0$ (MPa)</th>
<th>$C_w$ (MPa)</th>
<th>$\varepsilon_w^{0.1}$ (-)</th>
<th>$\varphi_b$ (°)</th>
<th>$\kappa$ (m/s$^2$)</th>
<th>$H$ (m)</th>
<th>$V_x$ (cm/a)</th>
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<td>30</td>
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<td>20</td>
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<td>10</td>
<td>$10^{-12}$</td>
<td>5000</td>
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<td>3</td>
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<td>20</td>
<td>0.2</td>
<td>0.5 - 1.5</td>
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<td>5000</td>
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<td>3</td>
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<td>5000</td>
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<table>
<thead>
<tr>
<th>Simulation</th>
<th>$\varphi_0$ (°)</th>
<th>$\varphi_w$ (°)</th>
<th>$C_0$ (MPa)</th>
<th>$C_w$ (MPa)</th>
<th>$\dot{\varepsilon}_w$ (s$^{-1}$)</th>
<th>$\varphi_b$ (°)</th>
<th>$\kappa$ (m/s$^2$)</th>
<th>$H$ (m)</th>
<th>$V_x$ (cm/a)</th>
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<td>10</td>
<td>$10^{-12}$</td>
<td>5000</td>
<td>1</td>
<td>6</td>
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<td>$10^{-12}$</td>
<td>5000</td>
<td>1</td>
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<tr>
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<td>20</td>
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<td>$3.3 \cdot 10^{-15}$</td>
<td>10</td>
<td>$10^{-12}$</td>
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<td>1</td>
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### Table 7.3: Parameters of Mixed Simulations

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<th>$C_w$ (MPa)</th>
<th>$\dot{\varepsilon}_w^{0.1}$ (-)</th>
<th>$\varphi_b$ (°)</th>
<th>$\kappa$ (m/s²)</th>
<th>$H$ (m)</th>
<th>$V_x$ (cm/a)</th>
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<td>30</td>
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<td>0.5 - 1.5</td>
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<td>10</td>
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<td>5/15</td>
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<td>0.5 - 1.5</td>
<td>5/15</td>
<td>$10^{-20}$</td>
<td>5000</td>
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<td>DV1</td>
<td>30</td>
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<td>20</td>
<td>0.2</td>
<td>3.3 · $10^{-14}$</td>
<td>5/15</td>
<td>$10^{-12}$</td>
<td>5000</td>
<td>1</td>
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<td>DV3</td>
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<td>$10^{-12}$</td>
<td>2500</td>
<td>5</td>
<td>15,17</td>
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CHAPTER 7. 3D EFFECTS OF STRAIN VS. VELOCITY WEAKENING

Figure 7.2: Plastic weakening behaviour of material. a) Strain weakening is initiated after reaching an accumulated plastic strain $\varepsilon_{w}^{0}$. Then, initial friction angle and cohesion ($\phi_{0}$, $C_{0}$) both linearly decrease between $\varepsilon_{w}^{0}$ and $\varepsilon_{w}^{1}$ to the weakened values ($\phi_{w}$, $C_{w}$), which they keep. b) Velocity weakening instantaneously lowers the frictional strength from $\phi_{0}$, $C_{0}$ to $\phi_{w}$, $C_{w}$ when reaching the threshold strain-rate $\dot{\varepsilon}_{w}$. Unlike strain, strain-rates can decrease and material is strengthened again to its initial values if strain-rates fall below the critical value $\dot{\varepsilon}_{w}$.

7.2.4 Strain-rate weakening

Velocity weakening is computed like strain weakening. The friction angle and cohesion are lowered to a weakened value ($\phi_{w}$, $C_{w}$) when strain-rate is exceeding a critical value. In contrast to the strain-weakening routine, weakening does not increase linearly between two strain-rates, but depends only on one threshold strain-rate, $\dot{\varepsilon}_{w}$, above which weakening is activated (Figure 7.2b). The effect of strain-rate- or velocity-dependent weakening has been tested in several studies and has been compared to a linked strain-weakening/healing system (e.g., Bercovici, 1993; Tackley, 1998). It is argued that strain-rate dependent weakening has a similar effect as when weakened zones are healed due to healing strain-rate, i.e. subtracting the healing strain-rate multiplied with the time step from the accumulated strain. Series U simulations with velocity weakening are all weakened to equal values ($\phi_{w}$, $C_{w}$) but controlled by different threshold strain-rates $\dot{\varepsilon}_{w}$ (Table 7.2; UV models). Series D simulations are conducted with different weakening ($\phi_{w}$), sediment thickness ($H$) and shortening velocity ($V_{x}$) (Table 7.3; DV models).

Both, strain- and velocity-weakening limits are grid step-size dependent and are scaled to the respective amount of displacement and rate of displacement along the spontaneously forming faults (e.g., Gerya, 2013; Lavier et al., 2000; vanDinther et al., 2013).
7.2.5 Surface processes

A routine is implemented to mimic sedimentation in piggy-back, surface depressions. The surface between “rock” and “sticky-air” markers defines the apparent topography. The way the exact topography is obtained, independent of the nodal resolution, is described in Ruh et al. (2013). To simulate surface process, the topography surface is diffused at every time step depending on its local 2D curvature:

\[
\frac{\partial h_s}{\partial t} = \kappa \cdot \frac{\partial^2 h_s}{\partial x_i^2}
\]  

(7.1)

where \(\kappa\) is the diffusion constant, \(h_s\) the surface topography and \(x_i\) the spatial coordinates \((x_1 = x, x_2 = y)\). The diffusion constants applied in the presented simulations are listed in Tables 7.1 and 7.2. The topography diffusion equation was solved implicitly on one thread using the direct solver PARDISO (Schenk and Gartner, 2004, 2006).

To address deposition without erosion, markers with a “sticky-air” rheology below the updated surface transmute to “rock” rheology. Markers with “rock” rheology above the updated surface remain unchanged. The intensity of topography diffusion is low in most simulations and mainly aims to preclude overthrusting of low-viscosity “sticky-air” markers at the model surface, which may produce non-physical intrusion of “sticky-air” into spontaneously forming thrust zones. One set of simulations was conducted to investigate the influence of sedimentation on the structural evolution of the wedge by applying different diffusion constants (US3,7,8; Table 7.2).

Surface slopes of developing wedges have been calculated by linear regression. Only surface elevations 100 m higher then their initial value have been considered to discount the flat fore-wedge that has not yet been incorporated into the wedge.

7.3 Results

Results from simulations with uniform and dual setups and with either strain- or velocity-dependent weakening of the numerical wedges are presented separately. Deformation was initiated along the backstop in all simulations.
7.3.1 Uniform models (series U)

7.3.1.1 Strain-weakening uniform models (models US)

Results of uniform models with weakening depending on finite plastic strain with a wide range of internal friction angles and cohesions are presented in Figure 7.3. Two points deserve emphasis: (i) The larger the amount of weakening, (i.e. the lower fully weakened friction angle $\varphi_w$), the steeper the surface taper, and (ii) shear zones have larger offsets in more weakened models (Figure 7.3a). The surface tapers of our model match the theoretical minimum calculated with the critical wedge theory, taking the fully weakened values ($\varphi_w, C_w$) for internal strength (Figure 7.3a). The effect of internal wedge strength on critical tapers is illustrated with a flounder diagram; minimum and maximum critical tapers delimit the area of wedge stability (Figure 7.4a). For identical décollements, the minimum critical tapers are lower for stronger wedges. The stable wedge area is smaller for internally weaker wedges. If the internal strength of the wedge is equal to that of the décollement, the wedge has no stable area and minimum and maximum critical tapers are identical (Figure 7.4a). Surface slopes of simulations US1-5 have been plotted after a run time equivalent to 3 and 6 Ma, depending on their post-weakening strength (Figure 7.4b). After 3 Ma all simulations reached their minimum critical taper. After 6 Ma, most of the wedge tapers plot within the stable zone, except model US1, which theoretically has no stable area (Figure 7.3a).

Accumulated plastic strain after 11.5 – 11.6 Ma of models US1-6 show that deformation is more diffuse in wedges with less weakening (Figure 7.3b). Conversely, strain in strongly weakened models (e.g., US1-3) is localized into shear zones, including short-lived, backward-verging shear zones that delimit pop-up structures during frontal accretion of a new thrust sheet (Figure 7.3b). Such structures are not seen in less weakened models like US4-6, in which strain distribution is diffuse. Comparing models US5 and 6 discloses the influence of cohesion on shear zone development. Although the angle of the taper is similar, the development of localized shear zones with increased strain is clearly enhanced when internal cohesion is weakened (Figure 7.3b).
Figure 7.3: Profiles at $y = 25$ km across uniform simulations after 11.5 - 11.6 Ma with decreasing amounts of strain weakening from top to bottom. a) Deformed strata inferred from Lagrangian markers. Green: basal décollement. Brown and beige: sediment; different colours are visualisation passive markers with no rheological differences. Blue: sediments added in surface depression. Solid lines: analytical surface slopes calculated with weakened internal strength. b) Accumulated plastic strain (same colour scale, on top, for all profiles).
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Figure 7.4: a) Flounder diagrams defining minimum and maximum critical tapers for wedges with a basal friction angle $\phi_b = 10^\circ$ and internal friction angles $\phi = 10 - 30^\circ$. b) Analytical test of strain-weakened uniform wedges. Solid line: theoretically derived minimum critical taper. Circles: Surface slopes calculated by linear regression at $y = 25$ km after 3 Ma run time. Triangles: same profile after 6 Ma run time.

Viscosity and plastic strain are stored on every marker throughout the modelling process. Every “sediment” marker with a viscosity lower than its initial value of $10^{23}$ Pa·s can be considered to be plastically failed, according to the viscous formulation of the plasticity-law used in this study. In the following, we consider a marker to be strongly failed if its viscosity is $< 10^{22}$ Pa·s, the value that brings stresses down to yield stresses. Dynamics of the strain-weakened simulations are further investigated by looking at the strain-rate of failed and strongly failed markers after 10 Ma run time in each model (Figure 7.5).

For simulation US1, about 25% of all sediment markers are at failure ($\eta < 10^{23}$ Pa·s) with a dominant strain-rate of $\sim 10 - 15$ s$^{-1}$ (Figure 7.5a). Around 10% of these markers classify as strongly failed ($\eta < 10^{22}$ Pa·s) with a dominant strain-rate of $\sim 2 \cdot 10 - 14$ s$^{-1}$. With large fully weakened friction angles $\phi_w$, i.e. reduced weakening, several observations stand out: (i) The percentage of failed markers within the wedge increases from about 25% (model US1; Figure 7.5a) to about 45% (model CS6; Figure 7.5f); (ii) The percentage of strongly failed markers decreases from $\sim 10\%$ in US1, to $\sim 6\%$ in US6; (iii) The spectrum of strain-rate in failed markers becomes narrower for models with less weakening. A small amount of failed markers of model US1 have strain-rates $< 10^{-16}$ s$^{-1}$ and $> 10^{-13}$ s$^{-1}$ (Figure 7.5a). Strain-rate values for failed markers in model US6 range all between $10^{-16}$ s$^{-1}$ and $10^{-13}$ s$^{-1}$ (Figure 7.5f).
Figure 7.5: Strain-rate distribution of plastically yielded Lagrangian markers within the wedge of simulations US1-6 with different amounts of strain weakening after ∼10 Ma run time. Grey: All failed markers (η < 10^{23} Pa·s). Black: Strongly failed markers (η < 10^{22} Pa·s). Percentage numbers of all sedimentary markers. Insets: Temporal evolution of the percentage of Lagrangian markers that exhibit a plastic strain between ε^0_w = 0.5 and ε^1_w = 1.5.
Figure 7.6: Profiles at $y = 25$ km across velocity-weakened uniform simulations after 11.5 - 11.6 Ma with thresholds decreasing from top to bottom. a) Deformed sediments inferred from Lagrangian markers. Same colour code as Figure 7.3. b) Accumulated plastic strain. Large strain areas in piggy-back style depressions portray gravity-driven mass flows.
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The temporal development of accumulated plastic strain within strain-weakened models reveals that weakening is especially efficient if the friction angle is sufficiently lowered (Figure 7.5a-c). The small percentage of markers between plastic strains of $\varepsilon^0_w = 0.5$ and $\varepsilon^1_w = 1.5$ indicates that weakened markers ($\varepsilon_p > 0.5$) accumulate strain very fast to become fully weakened ($\varepsilon_p > 1.5$). Strain-rate, and therefore strain, is not localized in models where weakening does not reduce the friction angle (Figure 7.5e and f).

### 7.3.1.2 Velocity-weakening uniform models (models UV)

Simulations including velocity weakening are presented in Figure 7.6. Wedge deformation after a run time of $\sim 10$ Ma indicates that there are less shear zones in models where weakening is more difficult, i.e. the thresholds of $\varphi_w$ and $C_w$ are higher (Figure 7.6a). In general, spacing between shear zones is larger and shear zones dip with lower angles towards the backstop than in strain-weakened models (Figure 7.6a). For weakening thresholds of $10^{-13}$ s$^{-1}$ (model UV1) and $3.3 \cdot 10^{-14}$ s$^{-1}$ (model UV2), thrust sheets construct a classical imbricate fan type wedge. Backstop-verging shear zones occur for low strain-rate thresholds (model UV3; Figure 7.6a). In contrast to backstop-verging or out-of-sequence shear zones in strain-weakened simulations, a strongly strained rear, close to the backstop boundary, is another remarkable feature of models with low weakening thresholds. “Sediments” have thickened and rotated, dipping towards the trench (e.g., model UV4; Figure 7.6a). In fact, these strained rears result from the vertical shear zone along the backstop boundary, which contributes to local uplift.
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Figure 7.8: Strain-rate distribution of plastically yielded Lagrangian markers within the wedge of simulations UV1-4 with different thresholds for velocity weakening after ~ 10 Ma run time. Grey: All failed markers ($\eta < 10^{23}$ Pa·s). Black: Strongly failed markers ($\eta < 10^{22}$ Pa·s). Percentage numbers of sedimentary markers.

All velocity-weakened simulations have in common that ramp anticlines growing at the wedge toe collapse. The collapsed and deposited material is illustrated by increased plastic strain in depressions between thrust sheets (Figure 7.6b). Ramp shear bands predominantly proceed below such trench fillings. Surface slopes of velocity-weakened simulations in function of strain threshold reveal two trends: (i) The larger the weakening threshold, the flatter the tapers and (ii) surface slopes decrease with increasing run time (Figure 7.7). All surface slopes lay above the theoretical critical minimum taper calculated with an internal friction angle of $\varphi = 15^\circ$. 

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Strain-rate distribution of failed and strongly failed markers after \( \sim 10 \) Ma run time vary strongly with weakening thresholds (Figure 7.8). For model UV1 (\( \dot{\varepsilon}_w = 10^{-13} \) s\(^{-1}\)) 22% of cover Lagrangian markers are at failure (Figure 7.8a). The failed markers dominantly deform with a strain-rate slightly below \( 10^{-15} \) s\(^{-1}\), whereas strongly failed markers reach very high strain-rates up to \( 10^{-12} \) s\(^{-1}\). Strain-rates are less focussed with smaller velocity-weakening thresholds (Figure 7.8b and c). For \( \dot{\varepsilon}_w = 3.3 \cdot 10^{-15} \) s\(^{-1}\), only about 6% of “sediment” markers are at failure, half of them being strongly failed (Figure 7.8d). Strain-rates of strongly failed markers show a slight peak at the very largest strain-rates for all models of series UV (Figure 7.8).

7.3.1.3 Effect of sedimentation

Simulations differing only in their surface diffusion constant \( \kappa \) were conducted to test the effect of sedimentation on the structural and mechanical evolution of wedges (Figure 7.9). As most of known accretionary wedges are sub-marine (Morley et al., 2011), erosion was not implemented. In general, the volume of sediments filling piggy-back style basins increases with increasing surface diffusion constant. If surface processes are kept low

Figure 7.9: Snapshots of wedges with identical amount of strain weakening (\( \varphi_w = 20^\circ \)) and different surface diffusion constants \( \kappa \) after \( \sim 11 \) Ma run time. Same colour code as in Figures 7.3.
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($\kappa = 10^{-20} \text{ m/s}^2$), thrust sheets are stacked with little intervening sedimentary material (Figure 7.9a). Vertical offset components of shear zones are roughly equal to the sediment thickness. For intermediate sedimentation rates ($\kappa = 10^{-12} \text{ m/s}^2$), shear zone offsets are similar to offsets in low sedimentation simulations, whereas shear zone spacing increases (Figure 7.9b). For high sedimentation rates ($\kappa = 10^{-6} \text{ m/s}^2$), shear zones exhibit larger offsets than in lower sedimentation simulations and thrust spacing is further increased (Figure 7.9c). This widening of thrust spacing is impelled by the additional load from more sedimentation in front of the wedge, which locally increases the wedge strength. The consequence is to shift the deformation front further away from the rear. Tapers of all simulations are roughly equal and consistent with the critical wedge theory (Figure 7.9).

7.3.2 Dual models (series D)

Ten simulations with a linear discontinuity in décollement strength parallel to the shortening direction are presented (Table 7.3). The difference in décollement strength is the same in all models: the basal décollement is strong (friction angle $\varphi_b = 15^\circ$) from 0 to 50 km in y-direction and weaker (friction angle $\varphi_b = 5^\circ$) from 50 to 100 km (Figure 7.1b). Models differ in the amount of strain weakening, the shortening velocity and the sediment thickness to investigate how these parameters influence the development of a transfer zone.

7.3.2.1 Strain-weakening models (models DS)

Sediment thickness $H = 5 \text{ km}$

Three simulations with different weakening values of the wedge after $\sim 4 \text{ Ma run time}$ illustrate the influence of strain weakening along the separation between the décollement halves (Figure 7.10). If the wedge internal friction angle is reduced from $\varphi = 30^\circ$ to $\varphi = 10^\circ$ (model DS1), the surface trace of the frontal thrust is everywhere sub-parallel to the backstop (Figure 7.10a). Topography of the rear of the model varies, higher surface elevations standing above the stronger half of the décollement. The deformed wedge at $y = 0$ and $y = 100 \text{ km}$ displays different deformation styles (Figure 7.10a; black and white in the topography plots). Above the weak décollement ($y = 100 \text{ km}$), an imbricate thrust fan looks like that obtained in the uniform models (Figure 7.3a). Above the strong décollement ($y = 0 \text{ km}$), thrust sheets are more disturbed, especially
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Figure 7.10: Topography of dual simulations with change in décollement strength \( \phi_b \) after \( \sim 4 \) Ma run time for different values for weakening \( \phi_w \). Black/white: deformed strata at \( y = 0 \) and \( y = 100 \) km. Right column: Temporal evolution of surface slopes at \( y = 25 \) and \( y = 75 \) km width for each model (solid lines). Dashed lines: theoretical critical minimum tapers depending on internal and décollement strength.

by a normal shear zone dipping shallowly towards the wedge toe (Figure 7.10a). The evolution of this normal shear zone strongly influences the surface slope of the wedge. The fact that the highly strained “sediment” is weaker than the underlying décollement
leads to a repose angle equal to the minimum critical taper (no stable wedge area; Figure 7.4a).

Surface slopes are investigated at 25 and 75 km of the $y$-axis, the mid-widths of each wedge part. The temporal evolution of the surface slope at $y = 25$ km shows strong oscillations above and below the minimum critical taper. Surface slope is there slightly larger while the slope atop the weak base ($y = 75$ km) is slightly lower than the analytical value (Figure 7.10a).

For a smaller weakened friction angle ($\phi_w = 20^\circ$; model DS2), topographic peaks are lower than in model DS1 (Figure 7.10b). This is consistent with the critical wedge theory (Figure 7.4). The trace of the frontal shear zone is curved above the boundary between weak and strong décollement. Structurally, the wedge above the relatively weak base develops thrust imbrication dominantly dipping towards the backstop. Atop the strong base, the leading imbricate fan has less shear zones that each absorbs larger displacements than shear zones in the other half of the model (Figure 7.10b). As seen in the uniform models, wedges are generally flatter for smaller strain weakening (Figure 7.3). Surface slopes along profiles at $y = 25$ and 75 km indicate that tapers of the wedge initially exceed and subsequently decrease towards the theoretical minimum critical taper (Figure 7.10b). Tapers are relatively steep after $\sim 1$ Ma, because only one thrust sheet is contributing to the calculated surface slope.

The wedge exhibits a relatively low surface elevation if only internal wedge cohesion is decreased while the internal friction angle remains equal (Figure 7.10c; model DS6). Migration of the wedge toe above the weak décollement is not defined by shear zones since deformation remains diffuse. Atop the strong base, out-of-sequence shear zones cut through the earlier leading imbricate fan (Figure 7.10c). Low elevations go hand in hand with low surface slopes. Like in model DS2, surface slopes of both parts initially exceed the theoretical minimum taper and then decrease towards the critical value (Figure 7.10c).

The evolution of wedges with the same amount of weakening ($\phi_w = 20^\circ$, $C_w = 0.2$ MPa) but with different shortening velocity $V_x$ is illustrated with snapshots of the second invariant of the strain-rate tensor and accumulated plastic strain (Figure 7.11) after an equal volume of incoming material, at 5.6 Ma for DS2 ($V_x = 1$ cm/a) and 1.13 Ma for DS3 ($V_x = 5$ cm/a). In general, both wedge halves evolved by in-sequence development of shear zones. Strain-rates of the slow DS2 simulation indicate dominant shear band activity at the wedge front. In the wedge, slightly increased strain-rates indicate front-
ward verging out-of-sequence shearing along previously active weakened zones (Figure 7.11a). Accumulated plastic strain values indicate the weakening of frontward verging shear zones above the weak base. Plastic strain in conjugate backthrusts is below the critical value for initiation of their weakening (Figure 7.11b). Imbricate thrust zones are flatter above the strong décollement than above the weak décollement part. Backthrusts cut earlier shear zones.

In model DS3, a larger number of high strain-rate shear zones than in model DS1 contributes to the wedge shortening (Figure 7.11c). Frontward and backward verging shear zones are conjugate. Above the relatively weak base, shear zones in model DS3 are steeper and more closely spaced than in model DS2 (Figure 7.11d). Shear zones above the strong décollement do not evolve in sequence. Out-of-sequence shear zones lead to imbrication of tectonic lenses between anastomosed shear zones (Figure 7.11d).

Vertically averaged horizontal shear stresses are quantified within the wedge to test the effect of shortening velocity on the potential development of strike-slip shear zones above the décollement discontinuity (Figure 7.12a and b). As the applied numerical code is
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Figure 7.12: Vertically averaged horizontal shear stresses within the wedge of simulations with a) \( V_z = 1 \text{ cm/}a \) and b) \( V_z = 5 \text{ cm/}a \). c) Solid lines: numerically derived surface slopes over width (y-axis) for slow and fast compressed wedges. Grey dashed line: analytical critical minimum taper.

Based on a viscous Stokes flow, it is expected that stresses increase where the shortening velocity is boosted. However, the plastic rheological law sets an upper threshold to stresses. Although strike-slip shear zones cannot be identified within strain-rate plots (Figure 7.11), vertically averaged horizontal shear stresses indicate their potential occurrence along the décollement strength change (Figure 7.12a and b). Maximal values for averaged horizontal shear stresses in map view are \( \sim 70 \text{ MPa} \) for both models, DS2 and DS3.

Surface slopes of the DS2 and DS3 simulations are analytically identical because the critical wedge theory is time independent. Plots of these surface angles at 5.6 Ma (DS2) and 1.13 Ma (DS3) run time show that wedge tapers are steeper for faster shortening (Figure 7.12c). Model DS2 exhibits slopes \( \sim 1^\circ \) steeper than the analytical prediction, with a smooth transition between the two model parts. Tapers of model DS3 are up to \( 5^\circ \) steeper than the prediction above the strong décollement and around \( 3^\circ \) steeper than the critical minimum in the area with a weak base (Figure 7.12c). In general, all numerical wedge tapers are found to be within the stable area of the plot (Figure 7.4a).

Sediment sequence thickness \( H = 2.5 \text{ km} \)

Rock with a brittle/plastic rheology gets stronger with depth, i.e. larger deviatoric stresses are needed to fail the material. The maximal strength of a pile of equal material is therefore at its bottom and depends on the dynamic pressure. Simulations with an initial “sediment” thickness of 2.5 km test this hypothesis (Figure 7.13).
Accumulated plastic strain of models with $\varphi_w = 20^\circ$ and $V_x = 2$ cm/a show shear zones connected over the whole model width and curved where the switch in décollement strength is placed (Figure 7.13a). The deformed wedge displays forward-verging imbrication above the weak and strong bases. Thrust faults have wider spacing above the weak décollement (Figure 7.13a). A horizontal slice of the second invariant of the strain-rate tensor at $z = 1.5$ km shows that the most active frontal shear zones are interrupted and offset at the boundary between the different décollements (Figure 7.13b). Along this offset, a relatively high strain-rate zone occurs.

Highly strained narrow shear bands reaching the wedge surface are more numerous than in the slower-compressed model (Figure 7.13c). The deformation front above the weak base is shortened by several coeval shear zones. No frontal shear zone extends over the whole width of the model. Horizontal and vertical slices of strain-rate show a large deformation zone without strain localization along the wedge front (Figure 7.13d).
Towards the rear of the wedge, active, i.e. high strain-rate shear zones are offset along the basal strength boundary, marked by a slight increase of strain-rate.

7.3.2.2 Velocity-weakening models (models DV)

Sediment thickness $H = 5$ km

Velocity-weakened simulations ($\dot{\varepsilon}_w = 3.3 \cdot 10^{-14}$ s$^{-1}$) with an initial sediment thickness $H = 5$ km show very different wedge evolutions according to bulk shortening velocity (Figure 7.14). This contrasts with strain-weakened models, for which bulk shortening velocity has no major impact (Figure 7.11). The wedge above the weak décollement of a slowly shortened model ($V_x = 1$ cm/a) is characterized by shallow-dipping ramps cut by steep backthrusts (Figure 7.14a). A horizontal slice showing the second invariant of the strain-rate tensor at $z = 1.5$ km reveals a narrow shear zone along the boundary of décollement strength, separating a narrow, steep taper ($y = 0 – 50$ km) from a wide, flat
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Figure 7.15: Velocity-weakened simulations with lateral décollement strength change and a starting sediment thickness $H = 2.5$ km shortened at 2 cm/a (model DV3) and 5 cm/a (model DV4). Left column: accumulated strain with profiles of deformed strata (black/white) at $y = 0$ and $y = 100$ km. Right column: Slices of the second invariant of the strain-rate tensor at $z = 1.5$ km (horizontal slice) and $y = 0$ (vertical slice).

one (Figure 7.14b). Across the wedge, narrow, high strain-rate shear bands only occur at the wedge front; out-of-sequence deformation is absent.

The simulation with a shortening velocity of $V_x = 5$ cm/a exhibits frontward-verging shear zones with minor backthrusts (Figure 7.14c). Both wedge parts exhibit steeper tapers than their counter parts of the $V_x = 1$ cm/a model. A consequence of the steep tapers is large finite strain representing gravity avalanches at the surface (Figure 7.14c). The wedge is roughly equally narrow over the whole model width with no significant offset of the deformation front across the basal strength boundary (Figure 7.14c). Nevertheless, a horizontal slice of the second invariant of the strain-rate tensor at $z = 1.5$ km shows a narrow high strain-rate shear zone along this boundary (Figure 7.14d). Strain-rates also illustrate that several active shear zones, not only the frontal ramp as in simulation DV1 (Figure 7.14b), accommodate shortening of the wedge.
Sediment thickness $H = 2.5$ km

In contrast to simulations with a cover thickness $H = 5$ km (Figure 7.14), models with $H = 2.5$ km exhibit similar wedge lengths ($x$-axis) for different shortening velocities but at equal incoming volume (Figure 7.15). The wedge of model DV3 shows mainly forward verging shear zones above both, weak and strong bases (Figure 7.15a). Sporadic backthrusts above the weak décollement are steep. Slices of the second invariant of the strain-rate tensor at $z = 1.5$ km and $y = 0$ indicate that the two frontal shear zones are active without noticeable out-of-sequence deformation (Figure 7.15b). A narrow shear zone along the basal strength boundary offsets the wedge front at $y = 50$ km.

Increasing the shortening velocity to $V_x = 5$ cm/a leads to increased surface movement in form of gravitationally driven avalanches as described for models with a thicker sediment layer (Figure 7.15c). The wedge above the weaker décollement exhibits more backthrusts than in experiment at $V_x = 1$ cm/a. Shortening takes place along several shear zones in the wedge part with the relatively strong base (Figure 7.15d).

7.4 Discussion

Modelling results demonstrate (i) that there are major differences in structural and dynamic evolution depending on the amount and style of material weakening and (ii) that strain and strain-rate dependent weakening of the friction angle has a strong influence on the structural evolution along strike-slip transfer zones.

7.4.1 Strain vs. velocity weakening

There are obvious mechanical differences between strain-weakening and velocity-weakening simulations. Strain-weakening routines decrease cohesion and/or friction angle when accumulated plastic strain reaches a previously defined threshold. Once the material weakens, it remains weak and deformation remains localized within the weak zones. On the other hand, velocity weakening occurs when a threshold strain-rate is exceeded. This is transient since the material recovers its initial strength as soon as the strain-rate falls below the weakening threshold. This can be compared to healing of previously active shear zones. Healing and strength recovery of active fault zones are natural processes (Bos et al., 2000; Bos and Spiers, 2000) already investigated in numerical modelling (Gerya, 2013; Tackley, 2000). Nevertheless, these previous numerical studies
applied a strain-weakening routine to initiate the loss of strength but a healing strain- and/or temperature-rate. An important difference of this strain-weakening/healing approach and velocity weakening of this study is that initial weakening is instantaneous in velocity-weakened systems, but needs some amount of plastic deformation if weakening is strain-dependent.

The differences in dynamical and structural evolution of strain- and velocity-weakened systems is demonstrated by illustrating the weakened zones during the development of a strain-weakened and a velocity-weakened simulation (Figure 7.16). The strain-weakened simulation exhibits several weak bands, which all define an active, previously active or reactivated shear zone (Figure 7.16a). Shear bands initially occurred in conjugate sets. With ongoing shortening, forward-verging shear zones were preferred and backthrusts more or less abandoned (see also Figure 7.3a). Strain weakening favours the out-of-sequence reactivation of previously weakened shear zones. The stronger the weakening effect, i.e. the lower $\varphi_w$, the narrower are shear zones (Figure 7.5). With further growth of the wedge, shear zones close to the rear passively rotate towards steeper attitudes, although the critical wedge theory predicts shallower dipping shear planes according to main stress orientations (Figure 7.16a). Single shear zones within strain-weakened models accommodate more displacement if weakening is increased. The spatial and temporal development of shear zones is also strongly dependent on the amount of weakening ($\varphi_w, C_w$). These findings support earlier studies investigating the effect of strain weakening on thrust wedges (Ellis et al., 2004; Selzer et al., 2007).

In velocity-weakened simulations, only the instantaneously active shear zone is weak (Figure 7.16b). Single shear zones are constantly active for $\sim 4$ Ma. Low-inclination shear zones evolve into ramp to flat systems towards the wedge toe (Figure 7.16b; 7.46 Ma). The décollement is active in front of the active shear zone (increased strain-rates at 7.46 Ma). This indicates pure shear shortening and an increase of viscous stresses within the underthrusting sediments (7.46 Ma). Further shortening in front of the active shear zone ends with a jump of the active shear zone (Figure 7.16b; 9.37 Ma). The new ramp develops at the wedge toe, which is defined by a surface flow deposit. Spacing between shear zones is therefore mainly defined by the extension of gravity-driven flows. Down-slope, gravity-driven mass flows lower the taper at the wedge toe (9.37 Ma). Generally, higher local strain-rates are reached within shear zones than in strain-weakened models (Figure 7.5 and 7.8). Nevertheless, the main part of failed material is characterized by rather low strain-rates, around $10^{-15}$ $s^{-1}$ (Figure 7.8a-c).
Figure 7.16: Temporal evolution of weakened zones in strain- and velocity-weakened simulations. a) Accumulated plastic strain between $\varepsilon_0^w$ and $\varepsilon_1^w$ show weakened areas in a simulation with $\phi_w = 15^\circ$. b) Second invariant of the strain-rate tensor between $3.29 \cdot 10^{-14}$ and $3.31 \cdot 10^{-14}$ s$^{-1}$ show weakened zones of a velocity weakened simulation with a weakening threshold of $\dot{\varepsilon}_w = 3.3 \cdot 10^{-14}$ s$^{-1}$.

For rate-dependent weakening models, recurrence of the above described events (pure shear and stress increase, ramp development, gravity flow) lead to an in-sequence imbrication typical of fold-and-thrust belts. Mass flows at the wedge toe occur only in velocity-weakened setups (Figure 7.16) because the response of the material strength is instantaneous, whereas it needs building up some minimal amount of plastic strain in strain-weakened material.

7.4.2 Comparison with the critical wedge theory

The critical wedge theory predicts a minimum and a maximum critical total taper (surface $\alpha$ + base $\beta$ angles), depending on the internal strength of the wedge material, the strength of the décollement and the base angle $\beta$ (Dahlen, 1984; Dahlen et al., 1984; Davis et al., 1983). Furthermore, inclination of the main stress can be analytically predicted and be compared to numerical wedge models (Ruh et al., 2013, 2012).
Strain weakening leads to a non-recoverable reduction of the internal friction angle and cohesion. Surface slopes of uniform, strain-weakened simulations (CS models) indicate that the strength of the wedge is defined by its weakened strength \((\varphi_w, C_w)\). The more the wedge is weakened, the steeper the surface slopes (Figure 7.3a and 7.4b).

Surface slopes of velocity-weakened simulations \((\varphi_w = 15^\circ)\) are steeper than the analytical prediction made with an internal friction angle of \(\varphi = 15^\circ\) (Figure 7.7). In contrast to strain-weakened models, the internal strength is not defined by the weak friction angle and cohesion because weakening takes place only within one shear zone that heals to initial strength when the strain-rate decreases. Steeper tapers for lower threshold is understandable because weakening occurs at lower strain-rates, i.e. earlier, and weakens more intensely shear zones within the wedge. A weaker internal strength results in steeper tapers. However, due to the time-independency of the critical wedge theory, comparison with velocity-weakened simulations is inadequate.

Surface slopes in both parts of dual models are, like in uniform models, steeper where weakening is more efficient (Figure 7.10). According to the critical wedge theory, normal faulting takes place when surface slopes exceed the maximum critical taper. In the strong décollement part of model DS1, the minimum and maximum critical tapers are identical, because the internal strength \(\varphi_w = 10^\circ\) of the wedge is weaker than the underlying décollement \(\varphi_b = 15^\circ\). This explains alternating thrusting and normal faulting (Figure 7.10a). Surface slope oscillations result from these normal faults, which instantly lower the taper. Surface slopes of simulations with wedges weakened to \(\varphi_w = 20^\circ\) and \(30^\circ\) do not develop normal faulting and are consequently slightly above their critical minimum in both wedge parts (Figure 7.10b and c).

Although the analytical theory is time-independent, the bulk shortening velocity has a strong effect on average surface slopes (Figure 7.12c). Model DS3 with \(V_x = 5\) cm/a has a surface slope \(2 - 4^\circ\) steeper than model DS2 with \(V_x = 1\) cm/a. The reason lies in the numerical treatment of plasticity. If plastic failure occurs, viscosity is decreased so that maximal stresses are at the local yield stress (see appendix 7.A). Practically, this means that the active décollement is a low viscous layer. This low viscous basis develops shear stresses depending on the velocity difference through the layer \(\Delta V\), its viscosity \(\eta\) and its thickness \(H_b\):

\[
\tau_b = \eta \frac{\Delta V}{H_b} \tag{7.2}
\]

It is obvious that the faster the bulk shortening velocity, i.e. the potential shear velocity within the décollement, the larger the basal shear stresses. Viscous shear stresses within
the basal décollement drag the “sediments” towards the rear and increase the taper. The surface slope gains $\sim 1^\circ$ for a 500 m thick décollement with a viscosity of $10^{19}$ Pa·s and a shortening velocity of $V_\alpha = 1$ cm/a (Davis and Engelder, 1985; Ruh et al., 2012). This angle value matches the excess of surface slope in model DS2 with respect to its critical minimum taper (Figure 7.12c). An increase of bulk shortening velocity to $V_\alpha = 5$ cm/a increases basal shear stresses and additional surface slope by an factor of 5. The excess of taper is lower than $5^\circ$, because the décollement has a lower effective viscosity than in the slow model. Surface slopes of model DS3 fit the values predicted by the critical wedge theory combined with basal décollement drag (Figure 7.12c).

7.4.3 Effects of plastic/brittle weakening on dual models

Different amounts and types of weakening influence the evolution of uniform numerical wedges. Therefore, input parameters such as bulk shortening velocity, amount and type of weakening and sediment initial thickness affect the structural and dynamic developments of wedges where the décollement strength changes.

The influence of shortening velocity on dual, strain-weakened models is minor. Although fast wedges exhibit larger strain-rates than slow ones (Figure 7.11a and c), the horizontal shear stresses that act along the boundary of décollement strength are similar for simulations that started with a 5 km thick sedimentary layer (Figure 7.12a and b). Results of models with an initial 2.5 km thick sedimentary layer reveal that an increased amount of weakening can lead to localization of a high strain-rate, vertical shear band along the décollement strength-boundary (Figure 7.13b and d). Plots of plastic strain show that the transfer zone between a wide and a narrow wedge is wider for less weakened simulations (Figure 7.13e and g). Strike-slip movements in a thinner sediment layer are due to lower brittle/plastic yield strength than in thicker models. Therefore, smaller horizontal shear stresses are needed to yield the material.

Velocity-weakened numerical wedges exhibit major differences in deformatinal patterns and dynamic evolution. High strain-rate shear zones in the brittle/plastic wedge parallel to the décollement strength boundary develop independently of sediment thickness and bulk shortening velocity (Figure 7.13, 7.14 and 7.15). Horizontal slices of velocities in $x$-direction and vertical profiles of strain-rates of model DS4, DV1 and DV4 show that these shear bands are not cutting through the whole sediment package in strain weakened simulations (Figure 7.17a). If sediments are velocity-weakened, sub-vertical strike-slip shear zones develop (Figure 7.17b and c).
Figure 7.17: Horizontal slices at $z = 1.5$ km illustrating velocities in $x$-direction and related vertical profiles of the second invariant of the strain-rate tensor at $x = 60$ km for model DS1 (a), DV1 (b) and DV4 (c).

Like in uniform, velocity-weakened models, shallow-dipping normal shear zones act as detachments for surface mass flows. In dual simulations, mass flows occur at the toe of a wedge and all over the wedge surface above the relatively strong décollements (Figure 7.18a). Such gravity-driven mass flows can be compared to mass waste events described in accretionary wedges, such as the Makran or the Gibraltar arc (Burg et al., 2008; Medialdea et al., 2004). These surface flows are constantly lowering the wedge taper although the wedge is theoretically within the stable area, according to the critical wedge theory (Figure 7.18b).

7.4.4 Comparison to analogue models and natural systems

The formation and evolution of strike-slip systems has been extensively investigated with analogue modelling techniques (Dooley and Schreurs, 2012, and references therein). In compressional wedge models, shortening-parallel strike-slip faults can, for example, develop along transfer zones induced by lateral changes in indenter velocity (e.g., Reiter et al., 2011) or lateral variations in sediment thickness (e.g., Marques and Cobbold, 2006). In analogue models with lateral variation of décollement resistance, the formation of strike-slip faults depends on the strength contrast between the different décollements (Graveleau et al., 2012). Cotton and Koyi (2000) investigated structures that develop along a sudden change from ductile to frictional décollement. They produced strike-slip faults parallel to the shortening direction but the contrast of basal strength in their setup is large. Calassou et al. (1993) set up a milder change in basal strength, with both parts
of the base defined by a frictional rheology. The resulting transfer zone developed as bending of wedge frontal structures without strike-slip faults.

Like in experiment 3 of Calassou et al. (1993), strain-weakened numerical wedges of this study form curved frontal structures instead of strike-slip faults (Figure 7.11, 7.13 and 7.17a). This agrees with the fact that strength of layered sandbox models is also strain-dependent (Graveleau et al., 2012, and references therein).

In nature, large-scale strike-slip fault systems parallel to the shortening direction in fold-and-thrust belts are often induced by basement structures, such as the Minab-Zendan transfer fault defining the boundary between the Zagros fold belt and the Makran accretionary wedge (e.g., Regard et al., 2005), lateral strike-slip faults of the central Zagros (Baker et al., 1993) or the Oregon accretionary prism (Tobin et al., 1993). Major sinistral strike-slip/tear faults that are limited to the sedimentary cover occur in the Jura foreland fold-and-thrust belt (e.g., Mosar, 1999; Sommaruga, 1999). These faults are almost transverse to the fold trend and do not show any significant offset in the basement. According to simulations presented in this study, subvertical strike-slip zones uncoupled from the basement can only develop if wedges are velocity-weakened. Changes in thickness of the viscous salt décollement in the Jura mountains, would lead variations in basal strength that trigger the lateral fault zone development.
7.5 Conclusion

New numerical models were performed to investigate the effects of strain vs. velocity weakening on the dynamics and structural evolution of uniform wedges and wedges with a lateral change in décollement strength. Uniform settings show that:

1. Strain weakening generates shear zones that remain active throughout shortening. Increased weakening causes stronger localization into less shear zones with larger offsets. Out-of-sequence shear zones tend to become steeper with increasing distance to the wedge toe. 2. Numerical surface slopes compared to those from the analytical critical wedge theory show that the internal strength of strain-weakened wedges depends on weakened strength $C_w$ and $\dot{\gamma}_w$. A wedge is as strong as its weakest shear zone. 3. Velocity-weakened wedges develop low-angle ramping shear zones and gravity-driven mass flows at the toes, which are triggered by the instantaneous response of internal strength to locally increased strain-rate. 4. Lower strain-rate thresholds for velocity weakening produce steeper surface slopes, like in strain-weakened simulations.

The most important results from dual simulations are: 1. The bulk shortening velocity is not a critical factor to build up lateral shear stresses in strain-weakened models but strongly influences the evolution of velocity-weakened models. 2. Rather steep normal shear zones occur in strain-weakened wedges if the internal friction angle is reduced to a value lower than the décollement friction angle. In velocity-weakened simulations, normal shear zones trigger large mass flows and constantly lower the surface taper. 3. Strain-weakened dual simulations develop a curved frontal thrust transitioning from the weak to the strong décollement part. 4. Basement unrelated tear faults in the Jura mountains are not reproducible with strain-weakened wedge models. Velocity weakening is necessary to develop subvertical narrow shear zones with a strike-slip component.
7. A Numerical formulation

7. A.1 Governing equations

The mechanical model is built on the equations for conservation of mass assuming incompressibility (i.e., sediment compaction is neglected)

\[
\frac{\partial u_i}{\partial x_i} = 0 \quad (7.3)
\]

and the conservation of momentum, the Stokes equation

\[
-\frac{\partial P}{\partial x_i} + \frac{\partial \tau_{ij}}{\partial x_j} = \rho g_i \quad (7.4)
\]

where

\[
\tau_{ij} = 2\eta \dot{\varepsilon}_{ij} \quad (7.5)
\]

and

\[
\dot{\varepsilon}_{ij} = \frac{1}{2} \left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right) \quad (7.6)
\]

\(P\) is dynamic pressure, i.e. mean stress, \(u_i\) the velocity \((u_1 = u_x, u_2 = u_y, u_3 = u_z)\), \(x_i\) the spatial coordinates \((x_1 = x, x_2 = y, x_3 = z)\), \(\tau_{ij}\) the deviatoric stress tensor, \(\rho\) the density, \(g_i\) the gravitational acceleration \((g_1 = g_2 = 0, g_3 = 9.81 \text{ m/s}^2)\), \(\eta\) the viscosity and \(\dot{\varepsilon}_{ij}\) the strain-rate tensor.

If differential stresses exceed the yield stress, plastic failure follows the Drucker-Prager yield criterion with the plastic yield function \(F\) (equation 7.7) depending on the second invariant of the stress tensor \(\tau^{II}\) and the yield stress \(\sigma_y\)

\[
F = \tau^{II} - \sigma_y \quad (7.7)
\]

where

\[
\tau^{II} = \sqrt{\frac{1}{2} \tau_{ij}^2} \quad (7.8)
\]

and

\[
\sigma_y = P \cdot \sin \varphi + C \cdot \cos \varphi \quad (7.9)
\]

\(C\) is cohesion and \(\varphi\) the friction angle of the material.
7.A.2 Numerical implementation

The governing equations given in appendix 7.A.1 are solved numerically by discretizing equation (A7.5) in an implicit manner, using an efficient OpenMP-parallelized multigrid solver, fully parallel on 16 (series C) or 32 (series L) threads. We adopted a standard geodynamic modelling approach, using an effective viscosity formulation for the numerical treatment of visco-brittle/plastic deformation (e.g., Buiter et al., 2006, and reference therein). Like for non-Newtonian (e.g., power-law) viscous rheology, effective viscosity for visco-brittle/plastic flows characterizes a local ratio between the deviatoric stress and the strain-rate (equation A7.5). At places where the plastic yielding condition is not reached, deformation is linear viscous in accordance to the assumed background rock viscosity. The rheological behavior of the model is initially linear viscous. If stresses locally exceed the yield stress \( F(\tau_{ij}, P, \varphi, C) > 0 \), effective viscosities \( \eta \) are decreased following the second invariant of the strain-rate tensor \( \varepsilon^{II} \) and the yield stress \( \sigma_y \) until the maximum stresses are at the yield stress \( F = 0 \), according to

\[
\eta_{vp} = \frac{\sigma_y}{2\varepsilon^{II}} \tag{7.10}
\]

where

\[
\varepsilon^{II} = \sqrt{\frac{1}{2} \varepsilon_{ij}^2} \tag{7.11}
\]

and \( \eta_{vp} \) is the effective viscosity corrected for plasticity.

Direct plastic (Picard) iterations are simultaneous with time stepping. To ensure initial model convergence and proper initiation of shear bands, the first \( \sim 400 \) Picard iterations are computed with a very small time step (one year) implying a negligible amount of model displacement. The following computational time steps are also relatively short (max. 1000 years, depending on marker velocities), which warrants small material displacement per time step (< 10 m) and ensures accurate treatment of plasticity and good convergence of the multigrid solver.

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8 | Conclusions

8.1 Dynamic and mechanical implications on the Makran

The Makran is one of the largest accretionary wedge systems on earth. The fact that a large part of the wedge is exposed above sea level makes the Makran an excellent example to study the dynamics and mechanics of thin-skinned tectonics in general, and the evolution of accretionary wedges in particular. Growth strata are a tectono-sedimentary feature that links deformation and surface processes locally. In Makran, growth strata have been found in a large-wavelength low-amplitude syncline within Upper Miocene shallow marine marls (Dolati, 2010). The magnetostratigraphic investigation did not yield useful data to calculate accumulation or limb rotation rates for this particular fold. Nevertheless, several conclusions can be drawn from the geometry of the growth strata and the stratigraphic ages: (i) In the Late Miocene, the northern Coastal Makran represented a shallow marine, tectonically active foreshore; (ii) The offlap geometry (Riba, 1976) of the unconformity implies that the rate of fold amplification exceeded sedimentation rate; (iii) The shape of the syn-sedimentary synclines (large-wavelength, low-amplitude) points to a rheologically weak décollement, potentially represented by the up to 600 m thick underlying olistostrome emplaced in the Tortonian (Burg et al., 2008).

Mud volcanoes near the shore line of the Makran indicate a similar age (Middle to Late Miocene) for intermediate décollement layers. The effect of multiple décollements on the evolution of the Makran is evident: (i) The presence of several décollements in an incoming and compressed sedimentary pile leads to underthrusting of stratigraphy and subsequent tectonic underplating and (ii) Oligocene sedimentary sequences in the Inner Makran are more deformed than younger and more southern sequences, presumably detached along the intermediate décollement. Large-scale tectonic underplating leads to antiformal stacking and rear-uplift.
8.2 Temporal evolution of the Zagros Simply Folded Belt

Growth strata within the folded foreland of the Zagros Fold Belt have been investigated to determine the age of the onset of deformation in the Central Fars. The magnetostratigraphic study revealed a transition from a mainly marine to a fluvial depositional centre at \( \sim 6 \) Ma; the braided river deposits commenced at \( 3.2 - 3.1 \) Ma. Between 8 and 2 Ma, sediments were accumulated at an average rate of \( \sim 27 \) cm/ka (after compaction). A progressive unconformity was identified in the upper part of deltaic sediments, before the onset of alluvial river systems. This growth strata is dated to \( 3.7 - 3.5 \) Ma. This indicates a difference of \( \sim 10 \) Ma between folding in the north-eastern and the central part of the Fars arc. The geographic distance of \( \sim 100 \) km discloses a migration of the deformation front of \( \sim 1 \) cm/a in the Upper Miocene Fars. The in-sequence propagation of folding in the Zagros is furthermore supported by numerical models, which generate, once scaled to nature, a similar rate for deformation migration.

8.3 Mechanics of thin-skinned fold-and-thrust belts

The mechanics of thin-skinned fold-and-thrust belts were investigated with numerical models. A two-dimensional visco-elasto-plastic finite element code was applied to test the effect of rheological differences in the décollements layers on the evolution of brittle wedges. In addition, the effect of multiple weak layers within the numerical stratigraphy was tested. A three-dimensional visco-plastic finite difference code was developed to investigate the effect of rheological and geometric variations along strike. Internal wedge strength was varied by strain- and velocity-dependent weakening.

8.3.1 Influence of décollement rheology

There are main differences in fold-and-thrust belts overlying salt (viscous) or shale (frictional) décollements: (i) a viscous rock exhibits increased strain-rates even at low shear stresses, unlike a frictional rock that has to overcome its yield stress. Therefore, stresses in the overlying stratigraphy are dispersed much further if the décollement is viscous, resulting in wider wedges with smaller surface slope. (ii) Wedges above a shale décollement form a typical imbricate fan with thrust ramps verging forward. Viscous décollements
equally lead to backward and forward verging thrusts comparable to thrusts observed in solitary box folds as in the Jura Mountains.

The occurrence of multiple décollement levels within the deformed stratigraphy plays a major role in the structural evolution of thin-skinned fold-and-thrust belts. The strength relation between the basal and the within-sequence décollement is an important factor determining the deformation style in wedges with frictional décollement rheology. A relatively weak intermediate layer leads to tectonic underplating and antiformal stacking in the rear of the wedge. Multiple viscous décollements enhance bedding parallel gliding and enhance folding as deformation mode.

8.3.2 Influence of wedge strength

The internal strength of a wedge is defined by its weakest part. This is shown by modelling results and application of the critical wedge theory. Any form of rheological weakening strongly influences the structural and mechanical evolution of a thin-skinned fold-and-thrust belt. Strain-dependent weakening needs a certain amount of plastic strain to weaken the rock, whereas velocity weakening responds in immediate weakening, where strain-rates are locally increased. Several conclusions can be drawn: (i) Strain weakening supports the activity of out-of-sequence thrusts. (ii) Velocity-weakened wedges develop low-angle ramping shear zones and gravity-driven mass flows at the toes. Laterally varying simulations revealed that (i) subvertical narrow strike-slip shear zones reaching down to the décollement can only be reproduced if weakening is velocity-dependent. (ii) Strain-weakened simulations develop a curved frontal thrust transitioning from the weak to the strong décollement part.

8.4 Integration of geological and numerical results

It is important to combine field geological observations with long-term mechanics to understand the evolution of thin-skinned fold-and-thrust belts such as the Makran and the Zagros. Field observations can be incorporated into numerical models to assess the importance of peculiar features. Mud volcanoes and multiple décollement levels are Makran examples.

In the Zagros, dating of growth strata constrained the in-sequence development of the Simply Folded Belt, and new data were implemented in numerical models that tested the
role of the strength of the basal décollement. The spectacular, elongated fold chain of Zagros were reproduced in 3D numerical simulations. Folding as the dominant process is strongly depends on the occurrence of within-sequence, occasionally multiple weak layers in the folded sequence. En-échelon arrangement of folds results from the locally transpressive character of the Zagros orogen.
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Bibliography


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# Soundtrack

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