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1 Dynamic constraints on the crustal-scale rheology of the Zagros
2 fold belt, Iran

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11 **ABSTRACT**

12 Thin-skinned fold-and-thrust belts are generally considered as the result of contractional
13 deformation of a sedimentary succession over a weak décollement layer. The resulting surface
14 expression frequently consists of anti- and synclines, spaced in a fairly regular manner. It is thus
15 tempting to use this spacing along with other geological constraints to obtain insights in the
16 dynamics and rheology of the crust on geological time scales. Here we use the Zagros Mountains
17 of Iran as a case study as it is one of the most spectacular, well-studied thin-skinned fold-and-
18 thrust belts in the world. Both analytical and numerical models are employed to study what
19 controls fold-spacing and under which conditions folding dominates over thrusting. The models
20 show that if only a single basal décollement layer is present underneath a brittle sedimentary
21 cover, deformation is dominated by thrusting which is inconsistent with the data of Zagros Fold
22 Belt. If we instead take into account additional décollement layers that have been documented in

23 the field, a switch in deformation mode occurs and crustal-scale folding is obtained with the
24 correct spacing and timescales. We show that fold spacing can be used to constrain the friction
25 angle of the crust, which is ~ 5 degrees in Zagros Fold Belt. This implies that on geological
26 timescales, the upper crust is significantly weaker than previously thought, possibly due to the
27 effect of fluids.

28 INTRODUCTION

29 It is often assumed that fold belts can be explained by folding of a sedimentary layer
30 above a basal detachment formed by a weak layer. As the spacing between folds in such belts is
31 quite regular, we can consider them as a large-scale natural experiment of crustal deformation.
32 Ideally, it should be possible to combine fold spacing with other geological data and theory to
33 constrain parameters such as crustal rheology that are difficult or impossible to constrain from
34 field observations alone.

35 The classical explanation of folds in fold belts is that they are due to a folding instability,
36 which is well known for a homogeneous sedimentary sequence with either a power-law viscous
37 or an elastic rheology (Schmalholz et al., 2002; Burg et al., 2004; Schmalholz, 2006). The
38 dominant wavelength λ_{dom} , for a viscous power-law layer of viscosity η_{sed} and with exponent n
39 overlying a linear viscous layer of viscosity η_{salt} , is given by (Schmalholz et al., 2002):

$$40 \quad \lambda_{dom} = 3.1 \left(\frac{\eta_{sed}}{n\eta_{salt}} \right)^{\frac{1}{6}} \sqrt{\frac{H_{salt}}{H_{sed}}} H_{sed} \quad (1)$$

41 where H_{sed} and H_{salt} are the thicknesses of the sedimentary cover and of the salt, respectively.
42 The growth rate (q_{dom}) of this instability non-dimensionalized over the background strain rate
43 $\dot{\epsilon}$ is given by (Schmalholz et al., 2002):

$$44 \quad \frac{q_{dom}}{\dot{\epsilon}} = 2.5n \left(\frac{\eta}{n\eta_{salt}} \right)^{\frac{1}{3}} \frac{H_{salt}}{H_{sed}} = 2.5n \left(\frac{\lambda_{dom}}{3.1H_{sed}} \right)^2 \quad (2)$$

45 and a combination of numerical and analytical studies have shown that $q_{dom}/\dot{\epsilon}$ should be larger
46 than ~ 20 for folding to form observable folds, rather than homogeneous thickening (e.g., Kaus
47 et. al, 2008).

48 The Zagros Fold Belt of Iran constitutes a classical example of such a folded belt that is
49 geologically (e.g., Stocklin, 1968; Alavi, 2004; McQuarrie, 2004; Sherkati and Letouzey, 2004;
50 Mouthereau et al., 2007) and geophysically (e.g., Jahani et al., 2009; Hatzfeld and Molnar, 2010;
51 Nissen et al., 2010) well studied due to excellent exposure and extensive seismic and borehole
52 data from exploration. The main tectonic and stratigraphic units are summarized on Figure 1 and
53 show that a particular feature of the Zagros Fold Belt is a consistent spacing of folds with a
54 wavelength (λ_{dom}) of 14 ± 3 km. These folds are generally explained as detachment folding of
55 the post-Cambrian sedimentary sequence above a basal weak layer constituted by the Hormuz
56 salt.

57 The centroid depths of waveform-modeled earthquakes indicate that faulting is restricted
58 to two structural levels located in the competent sediment cover units at 5–6 km depth and within
59 the Precambrian basement at depth larger than 11 km down to depths of 30 km (e.g., Talebian
60 and Jackson, 2004; Nissen et al., 2010). Seismic reflection profiles (Jahani et al., 2009) and field
61 observations in the Fars region (Mouthereau et al., 2007) show a lack of major thrust faults
62 cutting the folded cover up to the surface. This confirms that detachment folding rather than
63 thrusting is the dominant deformation mode in the Zagros Fold Belt. In this aspect, the Zagros
64 Mountains differ from other fold-and-thrust belts such as the Jura Mountains, where large-offset

65 faults are continuous across the stratigraphic sequence, well imaged through seismic studies
66 (Simpson, 2009).

67 Detachment folding theory should thus be perfectly applicable to the Zagros Fold Belt.
68 Equations (1) and (2) show that fold spacing depends strongly on the rheology of the overburden
69 and on the thickness of the basal salt layer. In the Zagros, a linear viscous overburden ($n = 1$) and
70 a viscosity contrast of 100 between salt and overburden, requires a salt layer thickness of ~ 7.8
71 km to fit the observed spacing of folds (Equation 1). Yet, seismic data indicates that the
72 thickness of the Hormuz salt is no more than 1 or 2 km (Jackson et al., 1990; Mouthereau et al.,
73 2006; Jahani et al., 2009). If the sedimentary cover has a power law rheology instead, its power
74 law exponent should be $n \sim 23$ (Equation 2) to explain the data, which is considerably larger than
75 estimates from rock creep experiments (Ranalli, 1995). Large power law exponents are often
76 taken as evidence for a brittle rheology. Currently, however, there is no theory that can reliably
77 predict the spacing of detachment folds in the case of a brittle overburden.

78 There is thus presently no satisfactory explanation for (1) why deformation in the Zagros
79 Fold Belt is dominated by folding and not by thrusting and for (2) what controls the spacing of
80 folds and how it is linked to crustal rheology. In order to address this, we performed thermo-
81 mechanical numerical simulations to study the dynamics of detachment folding in the presence
82 of a brittle sedimentary cover.

83 **NUMERICAL MODEL**

84 To study the effect of using visco-elasto-plastic rheologies on crustal dynamics, we have
85 performed a series of numerical experiments using the finite element code MILAMIN_VEP
86 (e.g., Kaus, 2010 and GSA Data Repository DR1). The viscosities of the weak layers are
87 assumed to be linear and constant, which is a reasonable approximation for the rheology of salt.

88 The brittle layers have a temperature-dependent rheology of limestone (see DR1), which
89 correspond to the majority of rocks within the sedimentary cover (Fig. 1C, Mouthereau et al.,
90 2007). A linear geotherm of $25\text{ }^{\circ}\text{C.km}^{-1}$ is initially applied (see DR1). For the low-temperature
91 conditions of the Zagros Fold Belt, stresses are such that the rocks effectively deform in the
92 brittle regime. Our model domain is initially 200×7.225 km in size (see DR1). The top boundary
93 is a free surface with no erosion (see DR1) and a constant background strain rate of 10^{-15} s^{-1} is
94 applied at the right of the model box, which results in 15% shortening after 5.5 Myrs consistent
95 with geological constraints (see DR1). All other sides of the model have free slip conditions.
96 Finally, to initiate the folding, the interface between the salt and the overburden rocks has
97 random noise with maximum amplitude of 100 m. Model simulations are performed for 5.5
98 Myrs, after which results are interpreted.

99 **RESULTS FROM NUMERICAL SIMULATIONS**

100 With a 1.5 km-thick single basal detachment layer underlying a homogeneous brittle
101 sedimentary cover, the models develop faults rather than folds (Fig. 2B). Such faults develop at
102 early stages with a spacing that is approximately twice the brittle layer thickness. Subsequent
103 deformation is locked around these folds that have a box-fold geometry. Compared to the Zagros
104 Fold Belt, we thus obtain a too large wavelength and an incorrect deformation style. Additional
105 simulations where we varied the frictional parameters of the crust, or the viscosity of the salt
106 layer gave similar results (see Fig. DR1). We thus infer that it is impossible to reproduce the
107 observed finite wavelength of Zagros Fold Belt folds (Fig. 1) by considering only one weak basal
108 décollement layer, unless this layer has an unrealistically large thickness.

109 A detailed look at the stratigraphic column, however, reveals that the sedimentary cover
110 is not rheologically homogeneous. Instead there are several layers that are composed of relatively

111 weak rocks such as evaporites or shales (Fig. 1B, C, see detailed descriptions in McQuarrie,
112 2004; Sherkati et al., 2006 and Mouthereau et al., 2007). A second set of simulations took this
113 fine-scale rheological structure into account (Fig. 3). The results are remarkably different from
114 the previous experiments: rather than being fault-dominated, deformation is now achieved by
115 folding, with a final wavelength similar to the one observed in the Zagros Fold Belt (Fig. 3). An
116 analysis of the simulation shows that the spacing of the folds is fixed at a very early stage, after
117 which the individual structures grow without clear geometric pattern, in accordance with field
118 constraints (Mouthereau et al., 2007). The initial fine-scale rheological stratification of the
119 sediment cover of the Zagros Fold Belt thus has a first-order effect on the development of upper
120 crustal-scale structures. These results are in full agreement with a recent study of active
121 seismicity in the Zagros Fold Belt which showed that both the Hormuz salt layer and the
122 intermediate layers are mechanically-weak zones that form barriers to rupture for active faults
123 (Nissen et al., 2010).

124 **CONSTRAINTS ON CRUSTAL RHEOLOGY**

125 The simulations presented in this study highlight the different modes of deformation that
126 might occur in fold-and-thrust belts. However, they give limited insights into the underlying
127 physics, as it remains unclear how sensitive the spacing of structures is to the rheology of the
128 crust. For this reason, we developed a semi-analytical methodology drastically reducing the
129 computational requirements that allows us to predict the outcome of numerical simulations in a
130 large parameter space (see details in DR2). The resulting wavelength versus growth rate
131 diagrams have a single maximum as a function of non-dimensional wavelength (Fig. 4A).
132 Rigorously, these semi-analytical results are only valid for very small deformations. Yet, a
133 comparison with numerical simulations reveals that they correctly predict the spacing of folds

134 even after 5.5 Myrs, which confirms that fold-spacing is selected at a very early stage in the
135 evolution of a fold and thrust belt (Fig. 4A).

136 Results for a homogeneous and brittle sedimentary cover reveal that the dominant growth
137 rate is smaller than 20, which essentially means that folding will not be able to overcome
138 background pure-shear thickening. Indeed, our numerical simulations indicate that this leads to
139 fault-dominated deformation rather than folding (box folds, Fig. 2). If, on the other hand, weak
140 layers are taken into account in the sedimentary sequence, the growth rate is significantly larger
141 and the dominant wavelength is reduced (Fig. 4B). The addition of a single weak layer is
142 sufficient to switch deformation styles from fault- to fold- dominated, and elasticity has a minor
143 effect only.

144 Using the same semi-analytical methodology, we performed a large number of
145 simulations and found that the two most important parameters are the viscosity of the salt/weak
146 layers and the friction angle of the crust, whereas rock density plays little to no role. Plots of
147 dominant wavelength and growth rate versus those two parameters show an approximate equal
148 dependence on the two parameters (Fig. 4). The results also show that weak layers in all cases
149 yields growth rates that are sufficiently large for the folding instability to dominate faulting.

150 In the case of Zagros Fold Belt, the effective viscosity of salt has been determined to be
151 close to 10^{18} Pa.s, a value consistent with scaled laboratory-derived values (Spiers et al., 1990)
152 and other modeling studies (Van Keken et al., 1993; Mouthereau et al., 2006). If we combine this
153 with our method, we estimate that the effective friction angle for the crust in the Zagros Fold
154 Belt on geological timescales is around $5^{\circ} \pm 5^{\circ}$ (Fig. 4B).

155 **DISCUSSIONS AND CONCLUSIONS**

156 Contrary to the common view of fold belts that often consider a single major basal
157 décollement only, we demonstrate through the example of the Zagros Mountains that the whole
158 stratigraphic sequence might influence the dynamics of the belt. Heterogeneities within the
159 sedimentary cover, and weak layers in particular, control whether deformation is dominated by
160 crustal-scale folds or by thrusts. The stratigraphy of a fold belt plays a much larger role than
161 previously appreciated and should thus be taken into account if one wishes to reconcile field
162 observations with physically consistent models of geological processes.

163 Balancing geological cross-sections in fold-thrust belts is a difficult exercise that aims at
164 providing a consistent structural and kinematic interpretation of usually independent structural
165 data. Our method paves the way for developing future generations of 2D and 3D dynamic
166 reconstruction models for fold and thrust belts (e.g., Lechmann et al. 2010).

167 Moreover, we show that the regular spacing of folds puts constraints on the rheology of
168 the crust on geological timescales. In the case of Zagros Fold Belt, the value for the friction
169 angle we obtained in this manner is small ($<10^\circ$), which indicates that the crust was rather weak,
170 potentially due to large fluid pressures (e.g., Huismans et al. 2005).

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242 **FIGURE CAPTIONS**

243 Figure 1. Field constraints for the Zagros folded belt. A: Topography illustrates the regular
244 spacing of folds with amplitude ~500–1000 m over on an area of ~80 000 km². Fold crest length

245 are of ~50 km in average. Inset shows the distribution of fold wavelengths measured for 88
246 anticlines from the Zagros Folded Belt. B: Cross-section (aa') across the Zagros Fold Belt based
247 on field measurement (Mouthereau et al., 2007). λ corresponds to the average wavelength of the
248 folds. This value is slightly smaller than the 15.8 +/- 5.3 km from Mouthereau et al. (2007) that
249 took into account the folds from the whole Fars area. MFF and SF correspond to the seismogenic
250 Mountain Front Fault and the Surmeh Fault, respectively, associated with basement faulting.

251 Vertical fold velocity is 0.3–0.6 mm.yr⁻¹. C: Synthetic stratigraphic log where WL1, WL2 and
252 WL3 correspond to the weak layers in the sedimentary sequence (Fm: Formation).

253 Figure 2. Simulation with a basal décollement layer only. A: Initial setup with a sedimentary
254 thickness of 7.225 km. All rocks above the basal salt layer are homogeneous and have a friction
255 angle of 5° and a cohesion of 20 MPa. A background strain rate of 10⁻¹⁵ s⁻¹ is imposed at the
256 right model boundary. B: Geometry, strain rate, and vertical velocities after 1.5 Myrs and 5.5
257 Myrs respectively. Deformation is localized along crustal-scale plastic shear zones and
258 deformation structures are fault-dominated.

259 Figure 3. Simulation with intermediate crustal detachment layers. A: Initial setup as in figure 2,
260 but with three additional weak crustal detachment layers with 10¹⁸ Pa.s. B: Snapshots of
261 geometry, strain rate and vertical velocities at different times, which illustrate that crustal-scale
262 folds rather than faults dominate the deformation pattern. Note that folds do not grow
263 continuously with time, but rather grow to certain amplitude after which activity switches to a
264 different fold.

265 Figure 4. Influence of multiple weak layers and elasticity on folding. λ , H, q and $\dot{\epsilon}$ corresponds
266 to the wavelength, the total thickness, the growth rate and the background strain rate,
267 respectively. VP and VEP correspond to visco-plastic and visco-elasto-plastic simulations,

268 respectively. This diagram was produced using the semi-analytical approach described in DR2.
269 A: Growth rate values obtained for given values of λ/H for 0, 1, 2 and 3 weak layers. For each
270 case, the characteristic wavelength value corresponds to the highest value of growth rate (e.g.,
271 white star for the case with 3 weak layers). Insets show results of numerical simulations after 5.5
272 Myrs, which develop folds with a spacing that is in excellent agreement with the predicted
273 characteristic wavelength. A single basalt décollement layer results in small folding growth rates
274 and in thrust-dominated deformation. Addition of one or more weak layers to the brittle
275 sedimentary cover results increases the growth rate significantly and leads to folding-dominated
276 deformation. The ZFB brown area corresponds to the λ/H ratio of the Zagros Fold Belt. B:
277 diagrams of characteristic wavelength (left) and corresponding growth rate (right) versus
278 viscosity of the weak layers and friction angle of the crust. Thick white lines show the
279 constraints for Zagros Fold Belt (average ± 1 standard deviation). As in the Zagros Fold Belt
280 salt viscosity is constrained independently, the best-fit friction angle for the crust is $5\pm 5^\circ$. The
281 white star corresponds to the simulation of figure 3.
282 ¹GSA Data Repository item 2011xxx, xxxxxxxx, is available online at
283 www.geosociety.org/pubs/ft2011.htm, or on request from editing@geosociety.org or Documents
284 Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.







