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Epicenter distribution and magnitude of earthquakes in fold-thrust belts: insights from sandbox models

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Abstract. Scaled analogue models are used to illustrate the effect of basal friction and erosion on fault activity and hence on epicentre distribution and magnitude of earthquakes in the sedimentary cover of active fold-thrust belts. Model results suggest that in fold-thrust belts shortened above low-friction ductile decollements (rock salt or overpressured mudstone), low- to moderate-magnitude earthquakes ($M_w = 5.3-5.6$), distributed over a wide area, occur along several long-lived thrust faults. Conversely, in areas shortened above high-friction decollements large-magnitude earthquakes ($M_w = 6.6 - 6.8$), distributed over a narrow zone are likely to occur along few short- lived thrust ramps. Calculated magnitude of earthquakes from models and their distribution are in agreement with recorded earthquake pattern from the Zagros mountain belt, which is partially shortened above a ductile decollement of Hormuz salt formation. Model results also showed that erosion reactivates older inactive thrusts and promotes formation of out-of-sequence thrusts.

Introduction

In studies of the seismicity of fold-thrust mountain belts, the main focus of research has been on the seismic signature of deep basement faults ($> 15km$) under the sedimentary cover [Jackson, 1980; Fitch, 1981] because it is difficult to resolve the magnitude and epicenter of shallow seismic events from teleseismic data. However, structural and stratigraphical relationships in restored profiles reveal that the magnitude of shallow seismic events generated by thrust ramps in sedimentary cover rocks between $11 - 14$ km within the upper crust, can range between moderate to large ($5.0 < M_w < 7.5$), [Namson and Davis, 1990].

According to the critical-taper Coulomb wedge model [Davis et al., 1983], which was modified by David and Engelder [1985] and Dahlen [1990], thrust faults in a fold-thrust mountain belt or in a continental margin accretionary wedge propagate sequentially until a wedge taper is built such that the subsequent deformation involves transport of the whole wedge along the basal decollement. Wedges shortened above a ductile decollement develop a low-angle taper compared to the high-angle taper that characterises wedges shortened above a high friction decollement [Davis and Engelder, 1985]. In a fold- thrust belt, thrust faults usually form serially in front of a propagating basal thrust in a piggy-back mode, where the younger thrusts form in front of the older, less active ones. The conventional view on deformation kinematics in such a belt has been that displacement along older thrusts diminish as younger thrusts develop at the active front of the belt. However, in models of laminated visco-plastic materials shortened in a centrifuge, Dixon and Liu [1992] showed that folds and thrusts nucleate serially and propagate spatially from hinterland to foreland. They conclude that in their models, early-formed faults at the hinterland end of the fold-thrust belt continued to accumulate displacement as the deformation progressed. The life cycle of a thrust is strongly dependent on two main factors; basal friction and erosion/sedimentation. In this study, results of sandbox models are used to study the effect of erosion and basal friction on the life cycle of thrust faults within a fold-thrust belt.

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Model preparation and kinematics

Structural and stratigraphical relationships in geologic profiles produced from field and subsurface data can be used to monitor palaeoslip rates along thrust faults [Namson and Davis, 1990], but the precise timing of activity of individual thrusts is often poorly resolved. However, evolution of thrust faults can more easily be monitored in sandbox models, where initial stages are well documented. Two sets of analogue models are used here to study the effect of basal friction on distribution of active thrust faults and the slip rate along these thrusts within a fold-thrust belt. In both model sets, a package of passively layered loose sand, simulating sedimentary rocks in the upper ten kilometers of the brittle crust, was shortened from one end under normal gravity. Loose sand which shows Coulomb behaviour is an appropriate analogue material to simulate non-evaporitic sedimentary rocks of the brittle upper crust [Mulugeta and Koyi, 1987, 1992; Colletta et al., 1991; Liu et al., 1992; Weijermars et al. 1993; Koyi, 1995; Storti et al. 1995; Gutscher et al. 1996]. The cohessionless sand used in these experiments has an angle of internal friction of 30 degrees, similar to that of many sedimentary rocks. The models are scaled to nature by a length ratio of 10^{-5}, which implies that 1 cm in the model corresponds to 1 km in nature. The only parameter which was varied in the two sets of models was the basal friction. In the first set, models were shortened above a high-friction decollement (coefficient of basal friction $\mu_b = 0.47$, Figure 1a). In the second set, models were shortened above a ductile substrate introducing a low-friction decollement (Figure 1b). A Newtonian silicone polymer, commercially

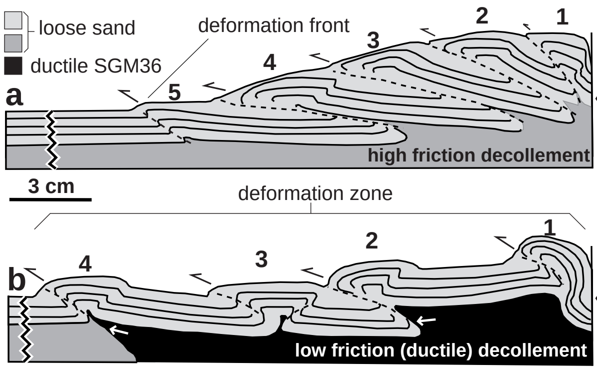

Figure 1. Line drawing of profiles of two models shortened above (a) frictional and (b) ductile decollements. Deformation zone is concentrated along the active thrust at the front of the wedge in (a), whereas in (b), the entire wedge is active. Half arrows indicate sense of movement along thrusts. The ductile layer in (b) smears the lower parts of the thrusts (white arrows). Numbers correspond to the structures, whose displacement is plotted in Figures 2 a and b, respectively.

labelled SGM36 [Weijermars, 1986], was used as a ductile substrate. The viscosity of this material ($\eta = 5 \cdot 10^4 Pa\,s$) is independent on strain rate. This material simulated rock salt or fluid overpressured mudstone and introduced a low friction decollement in the model. Herein, these two models are referred to as frictional and ductile decollement models respectively. During shortening of the models, the sand layers formed a stack of thrust imbricates simulating a fold-thrust belt or an accretionary prism. Displacement along individual thrusts in the models was calculated through continuous monitoring of the horizontal shortening distance between two marker points on either side of the trace of individual thrust faults on the surface of the model. Change in this horizontal distance (Δl) was converted to displacement (d) along each thrust fault by:

$$d = \Delta l / \cos\theta \qquad (1)$$

where θ is dip of the thrust fault.

Results

During shortening of the models above a frictional decollement, thrust faults formed serially in a piggy-back mode where younger thrusts developed in the footwall side of the older thrusts. In these models, with formation of new thrusts in front of them, older thrusts rotated to higher angle relative to the prevailing stress field and lock eventually. These thrust surfaces remained inactive as long as boundary conditions and the stress field were unchanged. With progressive shortening, only foreland verging thrusts formed and propagated spatially from the hinterland to foreland as the basal thrust propagated in a discontinuous mode towards the foreland. In these models, at any given time, shortening was accommodated by slip along one thrust which displayed a high average slip rate ($2.3 \cdot 10^{-2} - 4.4 \cdot 10^{-2} mm/s$), as compared with a bulk shortening rate of ($3.3 \cdot 10^{-2} mm/s$) (Figure 2a). With progressive contraction, the active thrust was locked and became inactive as a new thrust formed in the foreland to accommodate the bulk shortening [Mulugeta and Koyi, 1987; Koyi, 1995]. Previous models showed that a thrust fault developed in a model accretionary prism that was shortened above a frictional decollement had a limited lifetime, which started with its initiation and ended at the initiation of a new thrust fault in front of it [Koyi, 1995]. In models shortened above a ductile decollement, both foreland and hinterland verging thrusts formed (Figure 1b). During shortening of these models, several thrusts accommodated the contraction simultaneously since the basal decollement was active over a wide zone at a given time (Figure 2b). Here, older thrust faults remained active as new thrusts formed in front of them. Each thrust accommodated a portion of the bulk shortening. Therefore, at any given percentage bulk shortening, slip rate along individual thrust faults was relatively low ($2 \cdot 10^{-3} - 1.8 \cdot 10^{-2} mm/s$, Figure 1b). In these models, each thrust fault had a long lifetime and all remained active throughout model deformation.

Effect of erosion

Erosion is a dominant element in the tectonic evolution of mountain belts [DeCelles, 1994; Corrado et al. 1998; Mugnier et al. 1998]. Erosion thins and thereby decreases the load of sediments which a thrust needs to transport. Consequently, erosion prolongs the lifetime of an active thrust and may rejuvenate inactive thrusts. The effect of sedimentation and erosion on the kinematics of accretionary wedges has been studied by some workers [Storti and McClay, 1995; Corrado et al. 1998]. To study the effect of erosion on the evolution of individual thrusts and their seismic activity, in one of the models shortened above a frictional decollement, 30% of the wedge was eroded across the top of the model after 15% of shortening. By eroding the critical taper was "depleted". As a result, during progressive shortening, the wedge taper was rebuilt by tectonic thickening. This was

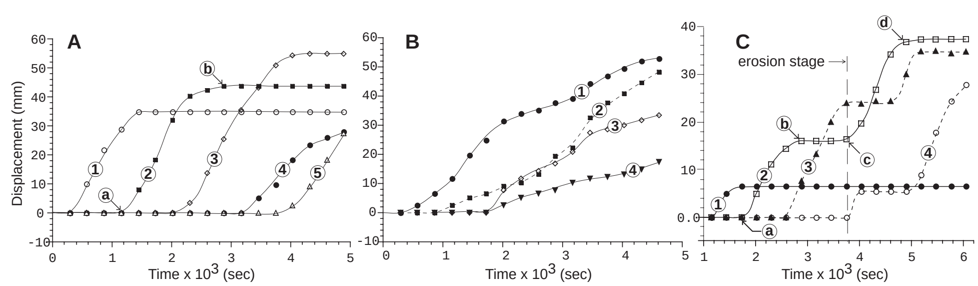

Figure 2. Plots of displacement along thrust faults versus time for models with (a) frictional and (b) ductile decollements. Circled numbers show the sequence of thrust formation. Circled letters (a and b in thrust 2) in (a) denote initiation and locking stages of the individual thrusts. (c) A similar plot for model with a frictional decollement that suffered erosion after 15% shortening. Note the renewed displacement along thrust number 2 immediately after erosion. Circled letters (a, b, c and d) denote initiation, locking, reactivation and re-locking stages of thrust number 2.

Table 1. Scaled magnitude (M_w) of earthquakes calculated from the models.

Type of	Model		Scaled quantity		M_w^a	M_w^b
decollement	A (cm^2)	D (cm)	A (cm^2)	D (cm)		
ductile	44	$2 \cdot 10^{-3}$	$4.4 \cdot 10^{11}$	20	5.38	5.58
ductile	49	$2 \cdot 10^{-3}$	$4.9 \cdot 10^{11}$	20	5.4	5.6
frictional	87	$4.2 \cdot 10^{-2}$	$8.7 \cdot 10^{11}$	420	6.45	6.66
frictional	150	$4.2 \cdot 10^{-2}$	$1.5 \cdot 10^{12}$	420	6.6	6.8

[a]Using rigidity constant of sandstone ($\mu = 1.5 \cdot 10^{11} dyn \cdot cm^{-2}$)
[b]Using rigidity constant of limestone ($\mu = 3 \cdot 10^{11} dyn \cdot cm^{-2}$)

achieved partly by folding of the wedge and partly by reactivation of the older thrusts located in the eroded section of the wedge. These reactivated older thrusts, which were inactive before the erosion, accommodated part of the bulk shortening (Figure 2c). Reactivation of these older inactive thrusts was necessary for rebuilding the wedge taper and further propagation of the basal thrust in front of the wedge. After rebuilding of the critical taper, displacement along the reactivated older thrusts ceased and it was instead accommodated along the youngest thrust. During rebuilding of the wedge taper, the youngest of the thrusts in front of the wedge was relatively inactive (Figure 2c). In fact, even after the wedge taper was rebuilt, erosion demobilised the wedge and, instead of the youngest thrust in front of the wedge resuming displacement, an out-of-sequence thrust formed behind the youngest thrust and accommodated the shortening. These results suggest that in eroding fold-thrust belts where denudation is effective, since erosion reactivates older inactive thrusts and cause formation of new out-of-sequence thrusts, the hinterland may remain seismically active during the entire shortening period. However, in the models shortened above a ductile decollement, out-of-sequence thrusts formed during the initial shortening stage without any erosion.

Seismicity

The moment magnitude (M_w) for an earthquake is [Aki, 1966; Kanamori, 1978]:

$$M_w = (\log(\mu AD)/1.5) - 10.7 \qquad (2)$$

where μ is rigidity constant of the ruptured material, A is the seismogenic rupture surface area and D is the average fault displacement during the earthquake. Since loose sand was used in all models μ was constant. The only variables between the two models were the seismogenic areas and average displacement along the thrusts. In the models, these parameters were measured assuming that only the ramp part of the thrust faults was producing earthquakes. The measured seismogenic ramp area ($A = 87 - 150cm^2$) along thrusts in models shortened above a frictional decollement was twice to almost four times larger the seismogenic ramp area ($A = 44 - 49cm^2$) along thrusts in models shortened above a ductile decollement (Table 1). Further, in models shortened above a ductile decollement, the thrusts incorporate the ductile layer, which smears the surface of the fault (Figure 2b). As a result, the smeared part of the thrust surface, which increases with time, becomes "aseismic". From the displacement rate of individual thrusts (Figures 2a and b), the average fault displacement (D) was calculated. Further, we assume that the sand layers simulate clastic and carbonate rocks with rigidity constant $\mu = 1.5 \cdot 10^{11}$ and $3 \cdot 10^{11} dyn \cdot cm^{-2}$, respectively. Calculated magnitude (M_w) of earthquakes scaled to nature shows that thrusts in fold-thrust belts shortened above a frictional decollement produce fewer, more localised earthquakes that are large (M_w up to 6.8). On the other hand, thrusts shortened above a ductile decollement produce more frequent and dispersed and smaller earthquakes (M_w as low as 5.4, Table 1).

Zagros fold-thrust belt

A suitable area to test our model results is the Zagros fold-thrust belt (Iran), where the focal mechanism solutions of most earthquakes imply that the seismicity is predominantly associated with thrust faulting [Nowroozi, 1972, Jackson & McKenzie, 1984, Jackson et al., 1995]. The (8-14 km) focal depths have been attributed to large earthquakes occurring in the uppermost part of the Arabian basement, beneath the Hormuz salt formation [Jackson & Fitch, 1981; Ni & Barazangi, 1986; Baker et al., 1993]. These observations have led many workers to ascribe NE dipping thrust faulting in the basement [Jackson & Fitch, 1981; Jackson & McKenzie, 1984; Ni & Barazangi, 1986; Berberian, 1995]. It is also suggested that unlike the Himalayas where there have been several major earthquake events ($M_w > 8$) during the last century, the largest events in the Zagros seem to be smaller ($M_w < 8$), [Molnar and Chen, 1982]. However, the sedimentary cover in the Zagros fold-thrust belt ranges between 7 to 13 kilometres [O'Brien, 1957; Colman-Sadd, 1978]. It is therefore argued here that some of the large earthquakes with focal depths of 8 to 14 km can be attributed to thrust ramps within the sedimentary cover. It is shown in the Cost Ranges of California that the magni-

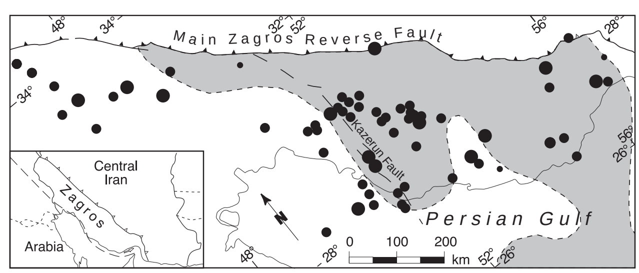

Figure 3. Map of southeastern Zagros fold-thrust belt showing epicenters of shallow ($< 15km$) earthquakes (U. S. Geological Survey, Preliminary Determination of Epicenters (1973-1999)). Note that the area shortened above ductile Hormuz salt (grey shade) show a broader seismic zone with more earthquakes than the area shortened above a frictional decollement (outside the grey area). The boundary of Hormuz salt is from Kent [1979].

tude of shallow seismic events generated by thrust ramps in sedimentary cover rocks located between 11 to 14 km within the upper crust, can range between moderate to large ($5.0 < M_w < 7.5$), [Namson and Davis, 1990].

In the southeastern part of Zagros fold-thrust belt, the sedimentary cover is shortened above a layer of salt (Hormuz formation) acting as a ductile decollement. In this area, microearthquakes occur within the sedimentary cover [Ni & Barazangi, 1986]. Whereas, medium to large magnitude earthquakes that occur in this area are attributed to fault displacement in the basement faults beneath the fold-thrust belt [Ni & Barazangi, 1986; Jackson & McKenzie, 1984]. However, our models suggest that the wide spatial distribution of shallow earthquakes recorded in the southeastern part of the Zagros fold-thrust belt indicate that the majority of the thrusts in the sedimentary cover are still active (Figure 3). In contrast, in the northwestern part of Zagros, where the salt layer is thin or missing, the sedimentary cover is shortened above a frictional decollement. In this areas, the epicentres of the earthquakes are restricted to a narrower zone implying that they are generated along one or two of the frontal thrust ramps (Figure 3).

Conclusions

Our models predict that small- to moderate-magnitude earthquakes along several long-lived thrust faults occur within a wide area of fold-thrust belts shortened above weak decollements (rock salt or mudstone). On the contrary, large- magnitude earthquakes along a few short-lived thrust ramps concentrated in a narrow area are characteristic of the outer edge of fold-thrust belts shortened above frictional decollements. However, since erosion plays a role in the tectonic evolution of natural fold-thrust belts, this narrow area may be located anywhere within the belt, not necessarily at the outer edge.

In model shortened above a frictional decollement, out-of-sequence thrusts form when the critical taper is depleted by erosion. In models shortened above a ductile decollement, out-of-sequence thrusts form without erosion.

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