What controls thickness of sediments and lithospheric deformation at a pull-apart basin?

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ABSTRACT

We present a simplified three-dimensional thermomechanical model of a pull-apart basin formed at an overstepping of an active continental transform fault. The modeling shows that for a given strike-slip displacement and friction on the faults, the major parameter that controls basin length, thickness of sediments, and deformation pattern beneath the basin is the thickness of the brittle layer. The unusually large length and sediment thickness of the Dead Sea basin, the classical pull-apart basin associated with the Dead Sea Transform, can be explained by 100 km of strike-slip motion and a thick (20–22 km, up to 27 km locally) brittle part of the cold lithosphere beneath the basin. The thinner sedimentary cover in the Gulf of Aqaba basin, located at the southernmost part of the Dead Sea Transform, close to the Red Sea Rift, is probably due to a thinner brittle part (<15 km) of the warmer lithosphere. The modeling also suggests no more than 3 km of Moho uplift beneath narrow (10–15 km) pull-apart basins formed in cold lithosphere, such as the Dead Sea basin. We also infer that a pull-apart basin may only form if a several-kilometer-thick ductile detachment zone exists between the brittle crust and upper mantle. Modeling shows that this would not be the case for the Dead Sea basin if the surface heat flow there were indeed as low as 40 mW/m^2 as previously reported. We consider this result as an indication that the surface heat flow at the Dead Sea might have been underestimated.

Keywords: pull-apart basin, numerical model, Dead Sea basin.

INTRODUCTION

Pull-apart basins belong to a special type of sedimentary basins associated with continental transform faults. They are depressions that are formed as a result of crustal extension in domains where the sense of fault overstepping or bending coincides with the fault motion sense (Crowell, 1974; Garfunkel, 1981) (see Fig. 1A). The outstanding classic example of a pull-apart basin is the 150-km-long Dead Sea basin (see Garfunkel and Ben-Avraham, 1996, and references therein), which is located at the Dead Sea Transform and where more than 8 km of sedimentary cover has accumulated since 15–17 Ma.

It remains unclear what determines the length of a pull-apart basin and the thickness of its sediments and how the associated extension strain is distributed at depth beneath the basin. Geological arguments (Garfunkel and Ben-Avraham, 1996) as well as gravity data (ten Brink, 1993) suggest that the deformation pattern beneath the Dead Sea basin may change significantly from upper crust to lower crust and to mantle lithosphere. For instance, in the case of the Dead Sea basin, strongly extended crust is apparently not accompanied by significantly uplifted Moho (ten Brink, 1993; Al-Zoubi and ten Brink, 2002), as is expected for a classical rift. Al-Zoubi and ten Brink (2002) explain this phenomenon by the presence of a strike-parallel ductile flow in the lower crust beneath a growing pull-apart basin, which compensates the extensional thinning of the upper crust. This process, however, requires rather low viscosity of the lower crust (Al-Zoubi and ten Brink, 2002) that is difficult to expect in the cold crust beneath the Dead Sea basin (ten Brink, 2002).

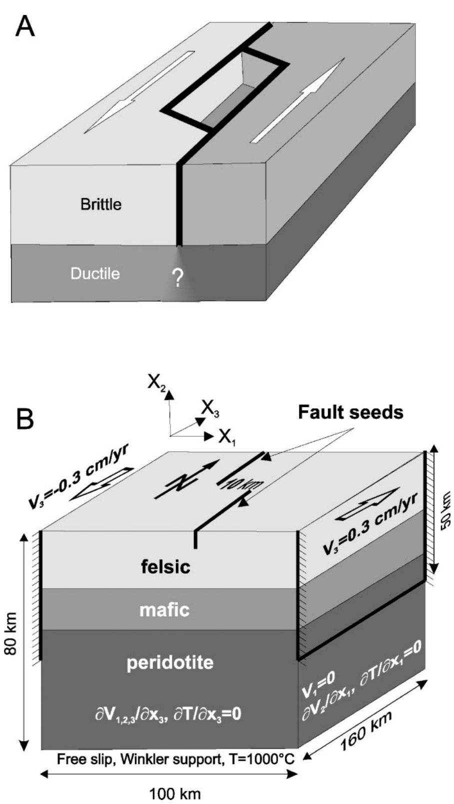

Figure 1. A: Conceptual model of a pull-apart basin formed at an overstepping of a transform fault. B: Model setup.

The way to study factors controlling the three-dimensional strain distribution at depth beneath a pull-apart basin is analog or numerical simulation of deformation. However, existing analog laboratory models are limited either to purely brittle rheology or to a brittle layer above a homogeneously viscous layer (Rahe et al., 1998; Corti et al., 2003; Smit, 2005). Previous numerical models of pull-apart basins either employ a two-dimensional thin-plate approximation, which does not resolve depth distribution of stress and strain (e.g., Segall and Pollard, 1980), or are limited to elastic rheology and small strains (Katzman et al., 1995; ten Brink et al., 1996).

In this study, we perform a simplified three-dimensional thermomechanical modeling of lithospheric deformation at a pull-apart basin with a structural and tectonic setting similar to the Dead Sea basin. Our model operates with realistic temperature- and stress-dependent visco-elasto-plastic rheology and allows for large strains. The modeling is focused on the analyses of factors controlling the length of a pull-apart basin and thickness of its sedimentary cover, as well as the magnitude and spatial distribution of the associated deformation in the lithosphere.

MODEL SETUP

We consider a model box of 100 × 160 × 80 km simulating a domain of continental lithosphere (Fig. 1B). The lithosphere is lithologically layered and thermally heterogeneous, including a two-layer crust and a mantle lithosphere with visco-elasto-plastic rheology. Brittle failure is simulated by the Mohr-Coulomb friction rheology with strain softening as in Sobolev et al. (2005; see also GSA Data Repository Table DR1[1]). In most of the models, the initial temperature distribution corresponds to a steady-state geotherm with a temperature of 1000 °C at 80 km depth and

[1]GSA Data Repository item 2006077, material-parameter and model-description tables, and additional model-results figures, is available online at www.geosociety.org/pubs/ft2006.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

60 mW/m^2 surface heat flow. We also computed several models with lower and higher temperatures, corresponding to surface heat flows of 50, 70, and 80 mW/m^2. Additionally, we introduce seeds of two parallel vertical faults with friction coefficient of 0.10 in the upper crust with an offset of 10–15 km, simulating overstepping of a transform fault (Fig. 1B). The model domain is subjected to left-lateral transform motion with a velocity of 0.6 cm/yr, leading to a total displacement of 105 km in 17.5 m.y., similar to the Dead Sea Transform tectonic setting (Garfunkel and Ben-Avraham, 1996). In all models, we assume that a surface depression deeper than 0.5 km is filled with sediments.

We model the deformation process by finite-element numerical integration of the fully coupled system of three-dimensional conservation equations for momentum, mass, and energy using realistic visco-elasto-plastic rheological models (Sobolev et al., 2005). We seek the solution at 16 parallel two-dimensional cross sections equally spaced along the X_3 axis (Fig. 1B). At each cross section, we employ an extended two-dimensional finite-element integration technique (Sobolev et al., 2005), and between cross sections we use a second-order finite-difference integration technique.

MODELING RESULTS

We have run a number of numerical experiments to study the sensitivity of the depth of a pull-apart basin and the lithospheric strain distribution to temperature and rheology of the crust, distance between the faults, and density of basin sediments. All models are indexed according to the values of the parameters (Table DR2; see footnote 1). The models with the ''strong'' crust correspond to a surface heat flow of 60 mW/m^2 and have crustal rheology according to the experimental data (Gleason and Tullis, 1995; Rybacki and Dresen, 2000). The models with ''weaker'' crust have reduced viscosities in the lower crust either due to the modified rheological parameters or due to the higher temperature. The ''weak'' model also has reduced viscosity in the upper crust. See Tables DR1 and DR2 for details of rheological models.

Some features of basin evolution are similar in all the models. During the first 1–3 m.y. model time, seeds of the faults are growing down and laterally, forming a few tens of kilometers long initial depression (Fig. 2A). The size of this depression does not depend much on the length of the fault seeds. Later on, the initial basin deepens and grows parallel to the strike-slip direction (Figs. 2A–2E). It is worth noting that the basin length grows faster than the strike-slip displacement (compare dashed line and other curves in Fig. 2D). After 100 km of strike-slip displacement, the basin length reaches 150–160 km in the models with the strongest crust. The weaker the crust, the closer is the final basin length to the total strike-slip displacement.

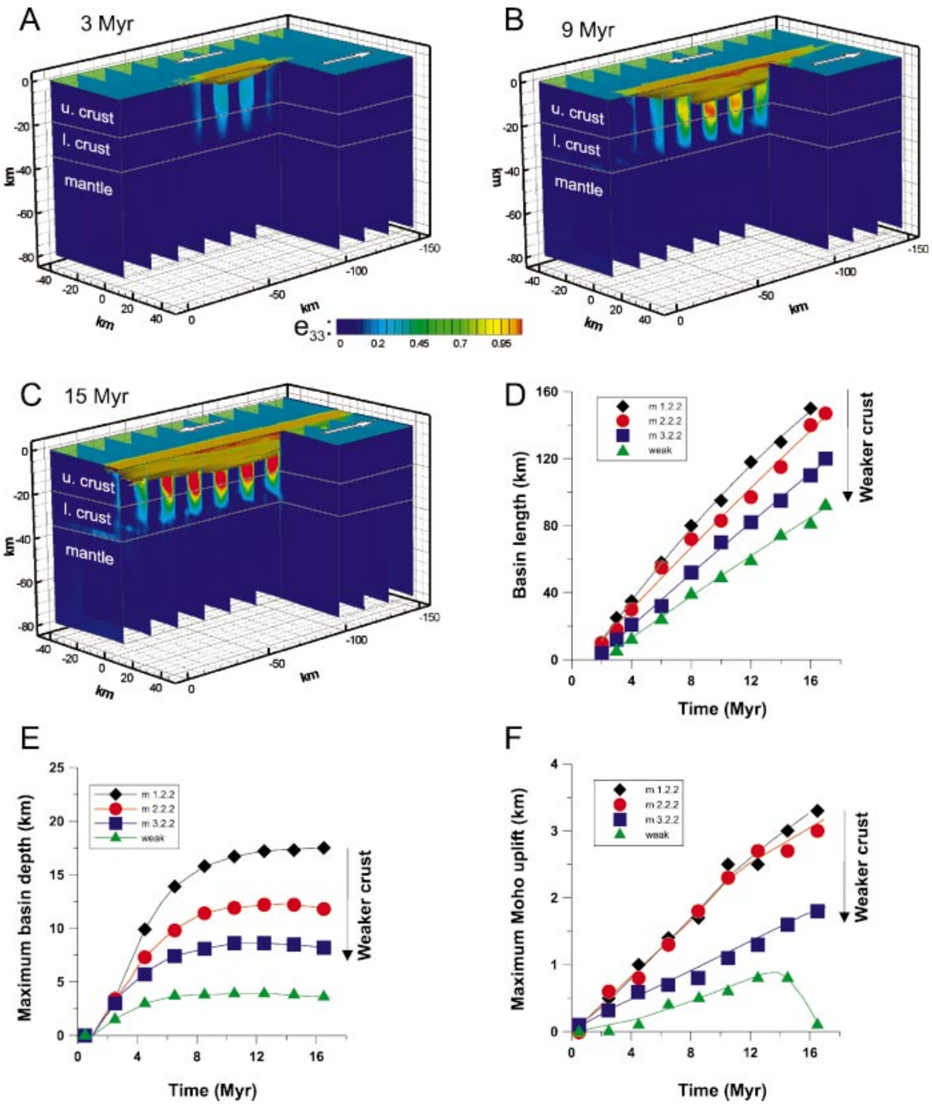

Figure 2. A–C: Growing pull-apart sedimentary basin (brown) together with distribution of transform-parallel extensional strain (e_{33}) shown at a number of cross sections. Time evolution of the length of a pull-apart basin (D), maximum depth of a basin (E), and maximum Moho uplift (F) for different crustal rheologies.

The strain distribution in all the models is significantly different in the brittle part of the crust and in the ductile lower crust and upper mantle. Most of the fault-parallel extension is concentrated in the brittle part of the crust between the faults, and not much extension is transmitted into the ductile part of the crust and into the upper mantle (Figs. 2A, 2B, 2C, and 3). In the upper brittle crust, the strike-slip strain is localized along the two parallel vertical faults, whereas in the lower crust and mantle, the strain is concentrated within a 20–30 km wide diffuse zone (Figs. DR1 and DR2), similar to the model of the continental transform fault (Sobolev et al., 2005). Domains of different deformation styles are separated by a ductile detachment zone located in the crust (Fig. 3). The presence of such a zone is crucial for the formation of a pull-apart basin (see below).

From Figure 2E, it is clear that the deepest sedimentary basins form in the strongest crust, and that the weaker the crust, the shallower the associated basin. It is also seen that the Moho uplift is small (less than 3 km; Fig. 2F) in all the models, even when the thickness of the sediments is large. The reasons for this particular behavior of the basin structure are discussed in the next section.

Interestingly, in none of our models with a cold lithosphere corresponding to a surface heat flow of 50 mW/m^2 was the pull-apart type of deformation in the crust achieved. The reason was that in such models the cold and viscous lower crust was mechanically attached to the strong upper mantle. In this case, the

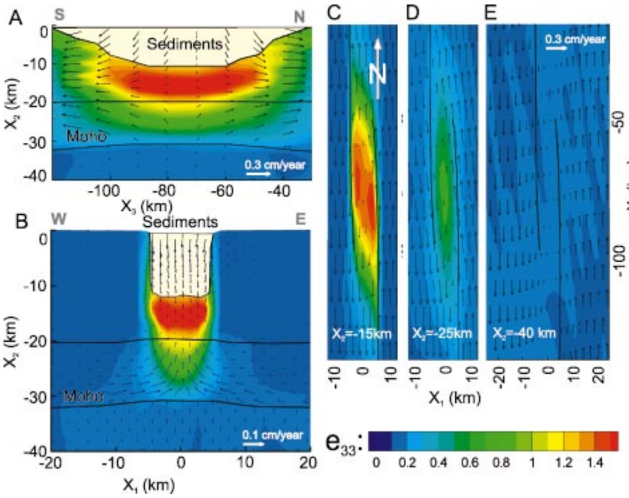

Figure 3. Deformation patterns in model m2.2.1 after 100 km of strike-slip displacement in the sections crossing the central part of a pull-apart basin parallel to the faults (A), perpendicular to the faults (B), and in horizontal cross sections in the upper crust (C), in the lower crust (D), and in the upper mantle (E). The deformation pattern is changing from "classical" pull-apart type of structure (as shown in Fig. 1A) in the upper crust to the diffuse shear zone in the mantle, with the transition pattern in the lower crustal detachment zone.

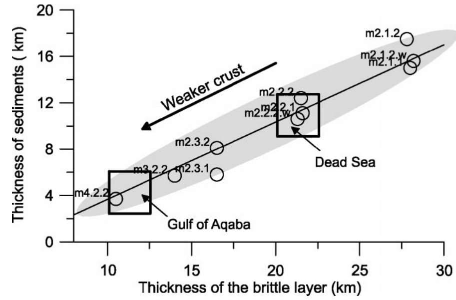

Figure 4. Maximum thickness of sediment fill after 100 km of strike-slip displacement versus thickness of the brittle layer. Each point indicates a particular model with indexes specified in Table DR2. Boxes show the most plausible conditions for the Dead Sea basin and Gulf of Aqaba basin based on observed depths of the maximum seismicity. Solid line corresponds to the simple model, when the extensional thinning of the brittle layer of initial thickness h_0 is compensated by a sedimentary layer of thickness h_{sed} and by uplift of the base of the layer by 3 km (similar to the magnitude of the Moho uplift); $h_{sed} = h_0 \times (1 - L_0/L) - 3$, where $L_0 = 50$ km and $L = 150$ km are initial and final lengths of the basin, respectively. The shaded ellipse indicates characteristic deviation of numerical solutions from the simple model.

effective thickness of the brittle layer dramatically increased, which made it impossible for large deformation of the crust, required to develop a pull-apart basin, to occur.

WHAT CONTROLS LENGTH OF A PULL-APART BASIN, THICKNESS OF ITS SEDIMENTARY COVER, AND THE MOHO UPLIFT?

As mentioned above, the length of a pull-apart basin grows faster than the strike-slip displacement (Fig. 2D). The common view, also supported by our modeling, is that a pull-apart basin grows along with the lengthening of its strike-slip border faults (e.g., Garfunkel and Ben-Avraham, 1996). In turn, these faults grow (1) due to the strike-slip displacement and (2) due to the rotation of the brittle blocks bordered by the end sections of the faults. The first mechanism leads to the uniaxial extension of the brittle crust, with a rate equal to the rate of the strike-slip displacement. The second, rotational mechanism increases the rate of the basin growth proportionally to the thickness of the rotating blocks, i.e., proportionally to the thickness of the brittle layer. As the depth to the brittle-ductile transition is larger in the stronger crust, the total rate of the basin growth appears to be higher for the stronger crust (Fig. 2D).

The crustal rheology also controls the thickness of the sedimentary fill of a pull-apart basin. Figure 4 shows that the maximum thickness of the sedimentary cover of a pull-apart basin is proportional to the thickness of the brittle layer. As a proxy for the latter value, we accept the depth of maximum energy dissipation beneath the model basin (see Fig. DR3). The depth of the bottom of the brittle layer, defined in this way, can be directly compared with the depth of maximum energy release by earthquakes, which is known for seismogenic pull-apart basins. The general trend in Figure 4 does not change much if the density of the sediments changes from 2200 to 2450 kg/m^3 and the distance between the faults changes from 10 to 15 km. The reason for this trend is that the extension within a pull-apart basin is almost entirely concentrated in the brittle layer (Figs. 2 and 3). In this case, according to the mass conservation law, the extensional thinning of the brittle layer, Δh, which causes the basin subsidence, is proportional to its initial thickness h_0; $\Delta h \approx h_0 \times (1 - L_0/L)$, where L_0 and L are initial and current lengths of the basin, respectively. Subsidence of a narrow (~10 km) basin is largely uncompensated isostatically in the relatively cold and thick lithosphere. Therefore, most of the thinning of the brittle layer, Δh, is compensated by the accumulating sediments, explaining the trend of Figure 4. The lack of isostatic compensation also explains why the significant variation of the density of the sediments and the width of the basin (keeping it still narrow) do not affect the deformation process much. This is also a reason why the Moho does not uplift much during the subsidence of the basin.

APPLICATION TO THE DEAD SEA BASIN

The Dead Sea basin is associated with the Dead Sea Transform, which has taken up ~105 km of left-lateral strike-slip motion during the last 16–17 m.y. (Garfunkel and Ben-Avraham, 1996). The only available seismic data (Ginzburg and Ben-Avraham, 1997) suggest that the thickness of the pull-apart-related sedimentary fill in the deepest part of the basin is between 8 and 14 km. Modeling of the gravity data also suggests that the maximum thickness of the pull-apart sediments in the Dead Sea basin is at least 8 km (ten Brink et al., 1993; Garfunkel and Ben-Avraham, 1996). The maximum seismicity is observed at a depth of 20–22 km beneath most of the basin, but reaches 27 km depth at ~31.3°N (Aldersons et al., 2003). Using these data and Figure 4, we can estimate the maximum depth of the sedimentary fill in most of the Dead Sea basin to be 8–12 km and up to 15 km at 31.3°N, in agreement with the observations (ten Brink et al., 1993; Ginzburg and Ben-Avraham, 1997). The closest fit to the Dead Sea basin case is given by models m2.2.1, m2.2.2, which result in 140–150-km-long basins after 100 km of strike-slip displacement (Fig. 2D). This is in agreement with the length of the Dead Sea basin, as well as with the estimation of the associated strike-slip displacement (Garfunkel and Ben-Avraham, 1996).

Now we can test our model by comparing its prediction with the observed thickness of sediments in another pull-apart basin at the Dead Sea Transform, the Gulf of Aqaba basin, which has accumulated only up to 5 km of sediments (Ben-Avraham, 1985). Seismicity beneath the Gulf of Aqaba is significantly shallower than beneath the Dead Sea (Aldersons et al., 2003), which is consistent with the higher surface heat flow than at the Dead Sea (Ben-Avraham and Von Herzen, 1987). The largest earthquake ever recorded in the Middle East occurred in 1995 at a depth of 12 km beneath the Gulf of Aqaba (Hofstetter, 2003). Taking this depth as the depth of the maximum seismicity beneath the Gulf of Aqaba, we estimate the maximum depth of the basin as 3–6 km (Fig. 4), in agreement with the observations (Ben-Avraham, 1985).

Another intensively discussed feature of the Dead Sea basin is the apparent absence of a significant Moho uplift beneath the deepest part of the basin (Al-Zoubi and ten Brink, 2002). According to our modeling, this is an expected feature of a narrow, 10–15-km-wide, pull-apart basin formed in relatively cold continental lithosphere. Our models generate no more than 3 km uplift of the Moho beneath such a basin due to the lack of isostatic compensation, without the hypothetical intensive lower crustal flow suggested by Al-Zoubi and ten Brink (2002).

Another common point of view is that the surface heat flow at the Dead Sea basin is ~40 mW/m^2 (Ben-Avraham et al., 1978; Eckstein and Simmons, 1979). However, in our modeling we obtain the closest fit to the Dead Sea parameters (depth of maximum seismicity, basin length, and thickness of sediments) with the model corresponding to a surface heat flow of 60 mW/m^2. From our models corresponding to a heat flow of 50 mW/m^2, we infer that no pull-apart type of deformation would occur in such cold lithosphere, due to mechanical attachment of the lower crust and strong mantle lithosphere. We consider these results as an indication that the surface heat flow at the Dead Sea might have been underestimated. Note that recent revision of some heat-flow data in the Dead Sea Transform region suggests a surface heat flow of 60 mW/m^2 (Förster et al., 2004), which is consistent with the Late Proterozoic age of the crust in the area (Artemieva and Mooney, 2001). Earlier, Ben-Avraham (1997) also suggested that the estimations of the heat flow at the Dead Sea might have been biased due to the presence of huge salt diapirs, and estimated the unbiased value of the heat flow to be ~60 mW/m^2.

Finally, we can formulate model predictions for the Dead Sea basin, which will be tested in the near future by the new interdisciplinary project similar to the DESERT Project carried out recently south of the Dead Sea (DESERT Group, 2004). Based on data on the depth of maximum seismicity beneath the Dead Sea (20–22 km, and 27 km at 31.3°N) and on our modeling results, we can predict that the maximum thickness of the pull-apart sedimentary fill of the Dead Sea basin is 8–15 km, maximum Moho uplift is less than 3 km, and surface heat flow (corrected for sedimentation and salt diapirs) is closer to 60 mW/m^2 than to 40 mW/m^2.

In conclusion, we infer from our three-dimensional thermomechanical modeling that the key factors controlling basin length, thickness of sediments, and deformation pattern beneath a pull-apart basin are (1) magnitude of the strike-slip displacement and (2) thickness of the brittle layer beneath the basin, which is, in turn, controlled by the temperature and rheology of the crust. Another important parameter is the friction coefficient at major faults. The effect of this parameter on the evolution of a pull-apart basin will be discussed in another paper.

We also conclude that a necessary condition for a pull-apart basin to form is a mechanical detachment between the brittle part of the crust and the rheologically strong uppermost mantle. The lack of this condition precludes development of a pull-apart basin in lithosphere with a surface heat flow below some 50 mW/m^2.

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