Thermo-mechanical model of the Dead Sea Transform

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Abstract

We employ finite-element thermo-mechanical modelling to study the dynamics of a continental transform boundary between the Arabian and African plates marked by the Dead Sea Transform (DST), that accommodated ca 105 km of relative transform displacement during the last 20 Myr. We show that in the initially cold lithosphere expected at the DST, shear deformation localizes in a 20–40 km wide zone where temperature-controlled mantle strength is minimal. The resulting mechanically weak decoupling zone extends sub-vertically through the entire lithosphere. One or two major faults at the top of this zone take up most of the transform displacement. These and other modelling results are consistent with geological observations and lithospheric structures imaged along the DESERT seismic line, suggesting that the Arava Valley segment of the DST is an almost pure strike-slip plate boundary with less than 3 km of transform-perpendicular extension. Modelling suggests that the location of the Arava Valley segment of the DST has been controlled by the minimum in lithospheric strength possibly associated with margin of the Arabian Shield lithosphere and/or by regionally increased crustal thickness. Models also show that heating of the Arabian Shield mantle adjacent to the DST, inferred independently by petrological and seismological studies, is required to explain asymmetric Late Cenozoic uplift in the area.

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1. Introduction

The transform plate boundaries are one of the three major types of such boundaries recognized in the classical plate tectonics concept [1,2]. Therefore understanding of their dynamics is a key issue in plate tectonics. Of particular interest is the dynamics of
transform boundaries crossing continental lithosphere, as their activity generates powerful earthquakes strongly influencing human activities. Despite numerous studies and remarkable achievements, the complicated deformation processes along continental transform boundaries are still incompletely understood. Key questions to be addressed are, how is the large shear strain along a transform plate boundary distributed within the heterogeneous and rheologically stratified continental lithosphere, and what are the factors controlling strain localization and partitioning? We address these questions using finite element thermo-mechanical modelling of lithospheric deformation constrained by geological data and high-resolution geophysical observations from the Dead Sea Transform (DST), which is probably the simplest and now one of the best studied continental transform boundaries.

The Dead Sea Transform (Fig. 1) forms the boundary between the Arabian plate and the Sinai sub-plate of the African plate [3,1,4–8]. The left lateral displacement along the DST amounts to about 105 km beginning at 14–18 Ma. The DST, especially its southern section, was the subject of comprehensive geological–geophysical studies ([4,6,9–13] and references therein). Recently the DST was much studied in the framework of the DESERT multi-disciplinary project [14] that focused on a transect crossing the DST in the central Arava Valley shown in Fig. 1.

![Fig. 1. Topographic map of the DST region. The location of the DESERT profile is shown by the black points.](image-url)
modelling study presented here we also focus on the central Arava Valley segment of the DST, located away from the large pull-apart basins and therefore probably the simplest part of the entire DST.

In addition to the key questions related to the transform boundary in general, mentioned above, we will also consider some important questions specific to the DST. Among these are the questions: (i) What is the origin of the narrow “rift-like” valley coinciding with the DST and what is the relative importance of the transform perpendicularly extension (rifting component) at the DST? (ii) What is the origin of the Cenozoic uplift resulting in an asymmetric surface topography at the DST? To answer these questions we will consider two groups of models. First we consider simplified models aimed to investigate effects of different types of lithospheric heterogeneity on localization and partitioning of the strike-slip deformation (first group of models). Then we analyse the evolution of the DST itself considering heterogeneous lithospheric structure around the DST (second group of models).

2. Strain partitioning in the model of simplified continental transform plate boundary

Let us consider first a typical lithologically (and rheologically) stratified continental lithosphere, which consists of a two-layer crust and mantle lithosphere. The anti-symmetric off-plane velocities of ±0.3 cm/yr (similar to the DST) are applied to the sides of the lithosphere thus subjecting it to left-lateral strike-slip deformation (Fig. 2, top). With this model setup we will study the effect of heterogeneities of crustal thickness and lithospheric temperature on the process of localization of the strike-slip deformation within rheologically stratified lithosphere.

The upper section of Fig. 2 shows three models. In the first model (left) the lower crust is thicker in the model left, while the temperature at the model bottom is kept constant. In the second model (middle) the crustal thickness is constant while the temperature at the bottom is elevated in the central part of the model. In the third model (right), both crustal thickness and temperature heterogeneities are present. In this particular model we assume that thickness of the crust is linearly increasing from the model left to the model right. The thermal lithosphere has constant thickness in the left part of the model and then linearly thickens to the model right. This model setup corresponds to the simplified structure of the passive continental margin.

2.1. Modelling technique

The deformation process is modelled by numerical integration of the fully coupled system of 3-D conservation equations for momentum (Eq. (1)), mass (Eq. (2)) and energy (Eq. (3)). These equations are solved together with appropriate rheological relations (Eqs. (4) and (5)) including those for a Maxwell visco-elastic body with temperature and strain-rate dependent viscosity (Eq. (4)), and a Mohr–Coulomb failure criterion with non-associated (zero dilation angle) shear flow potential Eq. (5).

\[
\frac{\partial p}{\partial x_i} + \frac{\partial \tau_{ij}}{\partial x_j} + \rho g_i = 0, \quad i = 1, 3; \quad \frac{\partial}{\partial x_3} = 0 \quad (1)
\]

\[
\frac{1}{K} \frac{dp}{dt} - \frac{a}{C_0} \frac{dT}{dt} = -\frac{\partial \gamma_i}{\partial x_i}, \quad (2)
\]

\[
\rho C_p \frac{dT}{dt} = \frac{\partial}{\partial x_j} \left( \lambda(x_i, T) \frac{\partial T}{\partial x_j} + \tau_{ij} \dot{\epsilon}_{ij} \right) + \rho \dot{A} \quad (3)
\]

\[
\frac{1}{2G} \frac{d\tau_{ij}}{dt} + \frac{1}{2\eta} \tau_{ij} = \dot{\epsilon}_{ij}; \quad (4)
\]

Fig. 2. Setup and results of the simplified strike-slip models. Columns correspond to the models with different initial crustal structure and temperature distribution. These initial conditions are shown in the top row; isolines of the initial temperature are shown by white curves. Lower rows show sequentially distribution of the lithospheric strength before the beginning of the deformation and time snapshots of the distribution of the strain rate demonstrating the strain localization process. (Left) Model with local crustal thickening (shown in the upper section) and minor temperature heterogeneity. (Middle) Model with homogeneous crust and strong temperature heterogeneity. (Right) Simplified model of a passive continental margin, combining both crustal structure- and temperature-heterogeneities. The strain rate (in 1/s) is shown in logarithmic scale.

\[
\sigma_1 - \sigma_3 = \frac{1}{2} \frac{1 + \sin \phi}{1 - \sin \phi} + 2C \sqrt{\frac{1 + \sin \phi}{1 - \sin \phi}} = 0; \quad \eta_s = \sigma_1 - \sigma_3 \quad (5)
\]
\(x_i\) are coordinates, \(t\) — time, \(v_i\) — velocities, \(p\) — pressure, \(\tau_{ij}\) and \(\varepsilon_{ij}\) — stress and strain deviators, \(\frac{d}{dt}\) — convective time derivative, \(\mathbf{g}_t\) — Jaumann co-rotational deviatoric stress rate, \(K\) and \(G\) — bulk and shear moduli, \(\eta\) — non-Newtonian power-law viscosity, \(B\), \(n\), \(E_a\) — parameters, \(I_2\) — second invariant of strain rate tensor, \(R\) — gas constant, \(T\) — temperature, \(\sigma_1\), \(\sigma_3\) — maximal and minimal principal stresses, \(c\) — cohesion, \(C_p\) — heat capacity, \(\dot{\lambda}\) — heat conductivity, \(A\) — radioactive heat production. Material parameters are listed in Table 1. For more details concerning equations and visco-elasto-plastic rheological relations see [15–18].

To simplify the problem we ignore variations of all parameters parallel to some selected direction in the horizontal plain, which we call the strike direction \((x_3)\). For all models discussed in this paper this direction is set to be parallel to the strike-slip velocities imposed at the side boundaries (Fig. 2). In this paper we use an extended 2-D approach, when all variables depend only on two coordinates, horizontal \((x_1)\) and vertical \((x_2)\), but all 3 components of the velocity vector and all 9 components of the strain or stress tensors can be non-zero.

We solve the Eqs. (1)–(5) using a finite-element modeling technique based on the dynamic relaxation algorithm [15–17]. The numerical code we use (LAPEX-2.5D) is an extended version of a parallel 2-D code [18]. This code, as well as its prototype PARAVOZ [16], applies an explicit, time-marching calculation scheme, which allows strongly non-linear visco-elasto-plastic rheologies. At each time step and for each element, the numerical algorithm chooses the energetically more efficient rheological model from the two options of (i) the Mohr–Coulomb elasto-plastic model with strain softening (Eq. (5)) or (ii) the non-linear temperature-dependent visco-elastic model (Eq. (4)). Shear heating is considered in both Mohr–Coulomb friction and ductile flow. Layer densities, thermo-elastic and rheological parameters are taken from published laboratory experimental data and are presented in Table 1. Strain softening in the elasto-plastic deformation mode (Table 1) allows for spontaneous self-generation of faults. The numerical experiments presented here were calculated with a finite element size of 1 km.

2.2. Modeling results

Fig. 2 shows the evolution of the shear strain rate with time for the models with initial crustal thickness heterogeneity (Fig. 2, left), initial temperature heterogeneity (Fig. 2, middle) and simplified model of continental margin (Fig. 2, right). In all cases the deformation process is with increasing time first “searching for” the optimal place for the shear strain localization. Thus, during the first 1–2 Myr multiple faults are generated in the brittle domain in the upper crust. The number of active faults then gradually

Table 1
Rheological and thermo-elastic parameters used in the modelling

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Upper crust</th>
<th>Lower crust</th>
<th>Mantle</th>
</tr>
</thead>
<tbody>
<tr>
<td>Elastic and thermal constants [19]</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(\rho), [kg/m^3](at 20 °C and 0.1 MPa)</td>
<td>2700</td>
<td>3000</td>
<td>3300</td>
</tr>
<tr>
<td>2450 (sediments)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(\eta), [K^-1]</td>
<td>3.7 (10^{-5})</td>
<td>2.7 (10^{-5})</td>
<td>3.0 (10^{-5})</td>
</tr>
<tr>
<td>(K), (G), [Gpa]</td>
<td>55, 36</td>
<td>63, 40</td>
<td>122, 74</td>
</tr>
<tr>
<td>(C_p), [J/kg/K]</td>
<td>1200</td>
<td>1200</td>
<td>1200</td>
</tr>
<tr>
<td>(\dot{\lambda}), [W/K/m]</td>
<td>2.5 or (\dot{\lambda}) (T) [20]</td>
<td>2.5</td>
<td>3.3</td>
</tr>
<tr>
<td>2.0 (sediments)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(A), [\mu W/m^3]</td>
<td>1.3 [24]</td>
<td>0.2</td>
<td>0</td>
</tr>
<tr>
<td>Power-law creep constants</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(B),[MPa^-1/\text{s}^-1]</td>
<td>1.0 (10^{-4})/1.0</td>
<td>4.0 (10^{2})/4.0 (10^{4})</td>
<td>4.9 (10^{4})</td>
</tr>
<tr>
<td>(E_a), [J/mol]</td>
<td>2.23 (10^{5})</td>
<td>3.56 (10^{5})</td>
<td>5.35 (10^{5})</td>
</tr>
<tr>
<td>(n),</td>
<td>4.0</td>
<td>3.0</td>
<td>3.5</td>
</tr>
<tr>
<td>Mohr–Coulomb elasto-plasticity with softening</td>
<td>Friction angle: 30°; dilation: 0°; cohesion: 20 MPa; linear decrease of cohesion to 2 MPa at strain 0.1; and of friction to 10 from strain 1.0 to 2.0</td>
<td>The same as for the crust</td>
<td></td>
</tr>
</tbody>
</table>
decreases until one single fault takes over at about 2 Myr model time. Simultaneously, the zone of high strain rate narrows and stabilises in the ductile lower crust and upper mantle.

The strain is localized exactly above the thickest crust in the model with crustal heterogeneity (Fig. 2, left), above the hottest mantle in the model with temperature heterogeneity (Fig. 2, middle) and just to the left of the margin of the thermal lithosphere root in the simplified model of passive continental margin (Fig. 2, right).

To understand what controls location of the localization points in Fig. 2, let us define a quantity which we hereafter call strike-slip lithospheric strength or simply lithospheric strength. This quantity \( S \) is an absolute value of a force required to drive a strike-slip deformation (with given reference strain rate \( \dot{e}_{13}^{\text{ref}} \)) at the vertical plain crossing the entire lithosphere at some point \( x_1 \) along a crossection oriented perpendicular to the direction of the strike slip deformation \( (x_3) \).

\[
S(x_1, \dot{e}_{13}^{\text{ref}}) = \left| \int_H \tau_{13} \left(x_1, x_2, \dot{e}_{13}^{\text{ref}} \right) dx_2 \right|
\]

Normalized variations of the lithospheric strength along model cross-sections are shown in Fig. 2 in second row from the top. The localization point always corresponds to the global minimum of the lithospheric strength, which is in turn controlled by the thickness of the crust and temperature distribution. For laterally constant thickness of the thermal lithosphere, the lowest strength is achieved where the crust is thickest. This happens because the top of the mantle lithosphere (its strongest part) is deepest and therefore has highest temperature and lowest strength beneath the thickest crust. In the thermally heterogeneous model with constant crustal thickness the strength minimum is located above the hottest part of the mantle, i.e., above the thinnest lithosphere. In the simplified model of the passive continental margin where both crust and lithosphere are thickening to the right, the localization is a result of competition of two effects. Due to the thickening crust strength tends to decrease to the right, while thickening lithosphere leads to the opposite effect. As a result, lithospheric strength first decreases and then increases to the right. The point of minimal strength in this case is located just to the left of the margin of strongly thickening lithosphere.

Not only position of the localization point, but also the rate of localization depends on the lithospheric strength. As it is seen from Fig. 2, the localization rate is the lowest where the strength minimum is least pronounced (continental margin model).

Partitioning of the deformation within the stratified lithosphere is also rather interesting. In the brittle upper crust the deformation is localized in a vertical fault (one-finite element column). In the ductile lower crust and upper mantle the shear strain localizes in a 20–40 km wide zone, with the high strain core some 5 km wide. Additional model runs (not presented here) with different thicknesses of the lithosphere and its rheology, with and without shear heating, demonstrate that the width of the shear zone in the ductile domain does not depend much on the thickness of the lithosphere. Instead, it is controlled by shear heating (in accordance with previous studies [25,26]).

All simplified models considered above contain relatively thick (80–120 km) lithosphere, similar to the DST case (see below). From these models we conclude that in such lithosphere and for a realistic temperature-dependent non-linear (power-law) rheology, the strike-slip deformation tends to localize in a relatively narrow zone (20–40 km) crossing the entire lithosphere with one major fault atop of it. The main factor controlling the position of the strain localization zone is the temperature of the mantle lithosphere right below the crust that, in turn, is controlled by the thicknesses of the crust and the lithosphere. In the case when both crust and lithosphere are thickening in the same direction the localization takes place before the place where the lithosphere begins to thicken rapidly, i.e., close to the margin of a shield.

In the next sections we apply our modeling technique to the DST itself, using geological and geophysical observations to constrain the models.

3. Constraints for the DST model

As we have seen above, lateral variations of the lithosphere temperature and crustal thickness play a key role in localizing strike-slip deformation. Therefore, here we discuss various geological and geophy-
sical data constraining lithospheric structure and thermal state in the DST region.

3.1. Thermal state of the lithosphere

The DST formed as a result of the Mid-Cenozoic break-up of the Arabian–Nubian platform that separated the Arabian from the African plate. The fractured area is underlain by a Late Proterozoic basement [27–29]. During most of the Phanerozoic this region was a part of a stable platform that experienced only a few episodes of igneous activity and mild deformation. The last important pre-breakup igneous activity was in the Early Cretaceous (140–120 Ma, [30]), so there was enough time for the thermal effects of this event to have largely decayed before the DST formed. The surface heat flow of 50–55 mW/m² [31,32] or 60 mW/m² [24], similar to the Protoerozoic shields worldwide, suggests a relatively cold crust and mantle lithosphere and the absence of large long-lived thermal perturbations.

However, during the Cenozoic continental breakup, widespread igneous activity resumed, also in the region crossed by the DST, especially on its eastern side [30,33,34]. Though sporadic activity has taken place since 20–18 Ma, the most voluminous igneous rocks formed 8–5 Ma ago. Volcanism of this age also occurred along the DESERT transect [35]. Mineral compositions of mantle xenoliths from these Neogene–Quaternary basalts near the DST indicate high upper mantle temperatures [36,37] that appear to be higher than expected for a steady-state conductive geotherm corresponding to the observed surface heat flow. High upper mantle temperatures below the young volcanics adjacent to the DST and the Red Sea are also indicated by low shear wave velocities [38,39]. A recent receiver function study [40] suggests a low S-velocity zone beneath the DST and adjacent Arabian Shield with the top at 60–80 km depth, which could be the lithosphere–asthenosphere boundary.

These data imply that, when continental break-up began, the lithosphere in the region around the DST was still cold and thick. However, as continental breakup progressed this initial state was modified and the lithosphere was likely thinned and a warmer geotherm established. This occurred quite recently (likely later than 10 Ma) so that there was not enough time to allow elevated heat flow to reach the surface yet.

The above scenario is supported by the history of uplifting which also suggests significant modification of the underlying lithosphere. Considerable, but variable, regional uplifting accompanied the continental break-up. Before break-up at the end of the Eocene (40–35 Ma), a shallow sea covered the area crossed by the DST, which shows that it was flat and close to sea level, whereas now the areas flanking the DST are 1–2 km above sea level. Uplifting developed gradually as the transform motion progressed. Elevations reached only a few hundred meters at 12–15 Ma, while about half the present relief is younger than 5 Ma (e.g., [41,35]).

For the modelling an important question is what was the topography of the lithosphere–asthenosphere boundary in the region just before the initiation of the DST at 15–20 Ma? There is good evidence [14] that the crust is now (and likely was at 15–20 Ma) gradually thickening from the Mediterranean margin towards the Arabian Shield. Little surface topography at 15–20 Ma implies that the isostatic effect of the Moho topography was largely compensated by the thickening of the lithosphere to the east of the Mediterranean similar to the simplified model of the continental margin discussed above.

In view of these observations, the initial state envisaged in our model is that of a relatively thick and cold lithosphere whose temperature reflects a steady-state conductive geotherm compatible with the surface heat flow of 50–60 mW/m² and crustal heat generation according to Förster et al. (in preparation) [24]. In this model we also assume that the lithosphere thickens to the east of the Mediterranean margin largely compensating the isostatic effect of the eastwards thickening crust (Fig. 3). Assuming of less than some 0.3 km of isostatically compensated altitude difference between left (SW) and right (NE) ends of the model cross-section, we estimate eastward thickening of the initial lithosphere to be about 50 km.

3.2. Crustal structure and strain distribution

Along the DESERT transect the crust thickens gradually from 27–30 km near the coast to about 40 km over a total distance of 250 km, with the DST in the middle [14]. Farther south and away from the
Mediterranean margin, the crustal thickness does not change so much perpendicular to the DST, although regional variations of crustal thickness by several kilometers cannot be excluded.

An important aspect examined in the modelling is the strain distribution at different depths. The DST is expressed by a belt of prominent faulting that comprises shallow segments dominated by major strike-slip faults that are sometimes transpressional, alternating with basins that are interpreted as pull-aparts [6]. Its southern half is mostly expressed as a topographic depression, much of which is delimited by normal faults. The DESERT transect crosses a saddle (central Arava Valley) between the Dead Sea and Gulf of Elat (Aqaba) basins that are 12 to 25 km wide and up to more than 10 km deep [6,13]. Where the DESERT transect crosses the DST, it consists of two major faults ca. 10 km apart, with most of the lateral motion taken up by the eastern fault — the Arava Fault [42,43]. Farther east a few normal faults, which likely also have a significant strike-slip component, delimit the high standing transform flank.

The DESERT project [14,44,45] revealed that the shallow structure dominated by discrete faults changes into a relatively broad zone of deformation deeper in the lithosphere. The SKS splitting study [44] suggests that a ca. 30 km wide anisotropic zone is located in the mantle to the west of the Arava Fault with the direction of the fast S-wave parallel to the fault and the delay time between fast and slow S-waves about 1.5 s. Another possible indication of the lithospheric deformation is a small (ca. 1–2 km) but clearly visible flexure of the Moho just beneath the Arava Valley [14]. The seismic data, however, do not show any evidence of a strong crustal thinning typical for a rift setting [14]. These observations suggest that deformation likely penetrates the Moho and becomes more diffuse with depth, similar to the results of the simplified models discussed in the Section 2.2. Comparison of these results with the results of the DST model will provide an important test of the model.

Analysis of the structure along southern half of the DST indicates the occurrence of some divergence (transtension) there [6]. Its magnitude varies along the transform, depending on minor irregularities in shape in plan view, and it appears to have increased during the last third of the transform history. Along the DESERT transect the transtension is estimated to be negligible, but in the Dead Sea and Gulf of Elat (Aqaba) basins north and south of the transect, the transtension reaches 5 km or more. This minor component of plate separation should cause some extension also at depth, allowing asthenospheric upwelling between the diverging plates [47] or it may lead to some other type of flow. In any case, because of the strength of the lithosphere, especially of its upper part, the effects of such movement will be felt over some distance along the transform. Therefore, in one of the
4. Model of the DST

4.1. Model setup

Our model setup, representing the situation along the DESERT transect, is shown in Fig. 3. The lithosphere is lithologically layered and thermally heterogeneous, including a three-layered crust thinning to the west and a mantle lithosphere. Initial crustal structure is fixed from the seismic observations [14], ignoring small Moho flexure. We do not correct the present day structure for the 105 km transform motion, because of small gradients of crustal structure parallel to the DST. In this approximation and because model properties do not vary parallel to the transform strike, the model represents an average picture of a ca. 100 km long segment of the DST. We consider the evolution of such a domain subjected to left-lateral transform motion with a velocity of 0.6 cm/year, leading to a total offset of 105 km in 17.5 Myr, and to a slight transform-perpendicular extension in one of the models. Layer densities and thermo-elastic parameters are taken from published laboratory experimental data and are consistent with a recent analysis for the area [24] (Table 1). All calculations have been done using the same rheological model for the mantle and two different rheological models for the crust (Table 1). One model is based on recent laboratory data for naturally hydrous quartz [21] and plagioclase [22] (hereafter called strong crust model) and another model is its modification where the effective viscosity at fixed strain rate is reduced by 10 times (hereafter called weak crust model). The numerical modelling technique is the same as described in Section 2.1.

The initial temperature distribution is assumed to be conductive steady state within the thermal lithosphere (defined as domain with temperature below 1200 °C) and constant below it. The expected shape of the 1200 °C (i.e., bottom of the thermal lithosphere) isotherm in the initial model, simulating thickening of the lithosphere towards the Arabian shield, is schematically shown in Fig. 3. The depth to this isotherm is calibrated in order to fulfill the following constraints: (1) Initial surface heat flow must be in the range of 50–60 mW/m² and (2) maximal initial altitude variation along profile must be less than 0.3 km. The location of the margin of the thickening lithosphere is constrained by the known position of the major fault at the surface (see also simplified models, Section 2.2). In a series of models we have also perturbed the initial temperature structure replacing a large portion of the lithosphere by the asthenosphere with $T=1200$ °C.

4.2. Modelling results

In the first model, presented in Fig. 4, we study pure strike-slip motion which starts at $t=0$. In this model (model 1) the lithosphere remains cold during the entire modelling time (17 Myr). Model 1 as well as all other DST models presented in this paper has two versions corresponding to the strong and weak crust. Similar to the simplified models of Section 2.2, the DST model 1 (and other models discussed below) show that the deformation process is largely controlled by the deformation of the strongest (upper) part of the mantle lithosphere. Shear strain localizes in a 20–40 km wide zone in the mantle, where the temperature-controlled lithospheric strength is at a minimum (see upper sections of Fig. 4), in accordance with an earlier suggestion [48,49]. The site of the localization zone is mostly controlled by the margin of the thickening shield lithosphere, similar to the simplified model of the passive continental margin discussed in Section 2.2. However, if in simplified model the major fault was localized just before the tip of the lithospheric root (see Fig. 2, right section), in model 1 it is shifted by about 15 km to the west. The reason of this shift is temperature increase caused by the westwards thickening sediments (see Fig. 3) which have relatively low thermal conductivity.

Due to the strong mechanical coupling between the crust and upper mantle in the model 1 with strong crust, the zone of largest crustal deformation is located just above the mantle deformation zone and is almost symmetric (Fig. 4, middle left section). In the upper brittle crust, shear strain localizes in one vertical fault, i.e., in one-element-wide column in the model, reaching to a depth of ca. 15 km. Deeper, in the lower, ductile part of the upper crust the fault transforms into the diffused deformation zone which focuses again in the upper (semi-brittle) part of the lower crust. Crustal
deformation is significantly different in the model 1 with weak crust. In this case the lower crust is partially decoupled from the upper mantle and the upper crust from the lower crust (Fig. 4 upper right section). As a result, deformation pattern becomes more complicated and asymmetric and more faults are generated in the upper crust. Quite interesting is calculated distribution of viscosity (Fig. 4 lower section). Due to the dependency of the viscosity on strain rate and temperature and due to the shear heating, the actively deforming lithosphere becomes weak, i.e., behaves mechanically like asthenosphere. This weak zone effectively decouples mutually moving plates. One consequence of this is that thermal and mechanical definitions of the lithosphere do not coincide within this decoupling zone.

The shortcoming of model 1 is that it does not explain neither magmatic activity in the region nor the Late Cenozoic uplift, nor the seismic observations suggesting present day thickness of the lithosphere.
to be about 70 km beneath the DST and its surroundings. Therefore we introduce here another model, model 2. In this model (Fig. 5) the lithosphere is thinned at $t=12$ Myr model time (ca. 5 Ma), i.e., the mantle lithosphere is replaced by the asthenosphere with a temperature of 1200°C within the domain below the white dashed curve in Fig. 5. The depth of this curve below the DST proper and east of it is taken according to the receiver function observations [40]. Shape of this boundary closer to the Mediterranean Sea is unknown, and we simply assume that lithospheric thickness reduction gradually reduces towards the Mediterranean Sea. The deformation pattern after 105 km of strike slip displacement in model 2 (Fig. 5) is almost the same as in model 1 (Fig. 4, middle section). However, model 2 explains better general topographical offset between the eastern and western flanks of the DST (Fig. 6). This suggests that changing the temperature at depth is essential for producing uplift. As most of the uplift occurred in the later stages of the transform development (mostly since 5 Ma, or at 12 Myr model time), we infer that most of the temperature change also occurred at that time, like it is assumed in model 2.

Although model 2 explains many features of the DST structure well, it fails to explain the “rift-like” topography of the Arava Valley. For instance model 2 predicts an elevation of more than 0.5 km immediately west of the master fault, although in reality there is a 0.4 km deep basin filled with young sediments.

Fig. 5. Cumulative finite strain at $t=17$ Myr in the pure strike-slip model with the lithosphere thermally perturbed at $t=12$ Myr. Left — strong crust model, right — weak crust model.

Fig. 6. Calculated surface topographies for models 1, 2 and 3 together with the observed topography along the DESERT line (solid red curve). Models and observations are combined assuming the same location of the master fault in the models and the Arava Fault.
ments immediately west of the Arava Fault (see Fig. 2 in [14]).

The Arava Valley topography is better fitted by model 3 (black curves in Fig. 6) in which we add a few kilometres of transform-perpendicular (east–west) extension, with all other conditions being the same as in model 2. Note, however, that as we consider neither erosion and sedimentation nor the 3D effects, we may not expect very good fit of the observed surface topography. We assume that extension starts at 12 Myr model time and continues till 17 Myr. model time. Total extension is 2 km in the strong- and 3 km in the weak-crust modifications of model 3. As Fig. 6 shows, the addition of only 2–3 km of extension (less than 3% of the strike-slip displacement) dramatically changes the topography close to the major transform fault. If more than 2–3 km of extension is implemented, the model generates deeper basin than observed (0.4–0.5 km) [14]. This allows to estimate maximal possible DST perpendicular extension at the DESERT line (see also Section 5.3 for discussion of this point).

The fact that small extension produces a large topographic effect in model 3 is because the most of extensional deformation is localized within narrow (20–40 km wide) upper mantle and lower crustal shear zones, where viscosity is reduced due to the high strike-slip strain rate (see the bottom section of Fig. 4). Fig. 7 shows time snapshots of the distribution of the strain rate norm (square root of the second invariant of strain rate tensor) for the strong-crust and the weak-crust versions of model 3. From Fig. 7 it is seen that extension tends to localize with time and combined with strike-slip deformation (transtension), it activates more faults than strike-slip deformation alone, especially in the case of the weak-crust model. Note also significant difference in deformation patterns between strong- and weak-crust models.

![Fig. 7. Distribution of the strain rate norm (square root of the second invariant of the strain rate tensor) for models with transtension (model 3) and pure strike slip (model 2) deformation, weak and strong crusts.](image)
5. Discussion

The successful geodynamic model must satisfy robust geophysical observations revealing lithospheric structure, with the seismic and seismological data from the DESERT project being an outstanding example of such observations. Therefore we begin with a discussion of the model results versus these data. Then, we focus the discussion on the factors controlling localization and depth distribution of the deformation at the DST i.e., the key questions at which the modelling was aimed. Finally we discuss the origin of the spectacular “rift like” surface topography variations across the DST.

5.1. Model results versus seismic observations

Fig. 8 shows deformation fields after the 105 km of the strike-slip motion predicted by model 3 with strong crust (left section) and weak crust (right section). It is seen that the deformation patterns at depth for these models are quit different. Model with strong crust generates rather simple and symmetric deformation field (Fig. 8b,c). No visible deformation of the...
Moho is associated with this model (Fig. 8a). In contrary model with weak crust generates rather complicated asymmetric deformation field with small (1 km) but visible flexure at Moho (Fig. 8d,e,f) consistent with seismic observations [14]. The weak-crust model 3 also predicts a zone of high horizontal shear strain in the lower crust east of the major fault (Fig. 8e). This zone is related to lower crustal flow, which accommodates deformation and it may partially account for the high lower crustal reflectivity observed east of the Arava Fault [14]. Apparently the weak-crust model 3 fits seismic data better than strong-crust model 3.

All our models predict a 20–40 km wide zone of high finite strain in the upper mantle. In the olivine-dominated rocks such a zone must be associated with the anisotropy of seismic waves with the direction of the fast S-wave sub-parallel to the strike-slip. We estimate the delay time between slow and fast S-waves to be 1.4 s for model 1 and 1.3 s for models 2 and 3. These numbers are obtained by calculation of S-wave anisotropy in the mantle from the calculated cumulative shear strain (Fig. 5) using the following simplified piecewise-linear relation between the strain and S-wave anisotropy, which is consistent with experimental data [50] and modelling results [51]: 
\{shear strain; anisotropy\} = \{0,0\}, \{0.5, 5\%\}, \{1.0, 7\%\}, \{2.0, 8.5\%\}, \{infinity, 8.5\%\}. Both the orientations of the fast S axis and the magnitudes of the delay times in the models agree well with the SKS splitting observations [44].

Although transtention model (model 3) with weak crust is generally consistent with most of seismic and seismological observations along DESERT line, some of the observed features remain unexplained. None of our models can replicate the bright seismic reflector and converter 10 km above the Moho observed in both seismic [14] and seismological [45] data. Obviously this reflector/converter demands some feature currently not present in the model, such as an inherited lithological heterogeneity possibly also associated with the localized shear strain. Another unexplained feature is westward shift of the mantle anisotropic zone relative to the Arava fault [45]. In all our models the major fault is located either above the central part of the mantle deformation zone or is even slightly shifted to the west of it (Fig. 4), not to the east as observed. We suggest two possible explanations of this phenomenon. First, possibility is that the Arava fault at the DESERT line is relatively young feature, and most of the strike-slip displacement is taken by other faults west of it, located above the zone of strongly anisotropic mantle. This idea is in line with seismic observation suggesting that the uppermost section of the Arava fault is very narrow (10 m) [46], but apparently contradicts some geological observations [43]. Another possibility is a 3-D effect of presence of the Dead Sea pull-apart basin less than 100 km to the north of the DESERT line. If the mantle deformation zone is located right beneath the Dead Sea basin and continues to the south parallel to the strike of the DST, then it must cross the DESERT line indeed west of the Arava fault.

5.2. Controls of the position of the DST and strain distribution at depth

The question of what controls the position of the DST is among the basic questions concerning this plate boundary. Steckler and Ten-Brink [48] suggested that the DST north of the Dead Sea follows the zone of the minimum lithospheric strength at the Mediterranean margin, while to the south of the Dead Sea, where this margin becomes too oblique to the plate motion direction, the DST is parallel to the plate velocity. In this hypothesis the lithospheric heterogeneity south of the Dead Sea did not play a role in the localization of the DST. Our modelling shows that the strike-slip deformation does localize in the zones of minimal lithospheric strength, confirming the idea of Steckler and Ten-Brink [48] in this respect. However, we also show that possible localizing factor was a margin of the thick lithospheric root of the Arabian shield located below the part of the Arava valley at the pre DST time or regional thickening of the crust at the trace of the DST. Therefore we suggest that between the Dead Sea and the Red Sea the location of the DST might have been significantly influenced by the inherited regional minimum of the lithospheric strength.

In all our models the strike-slip deformation localizes in the sub-vertical zone crossing the entire lithosphere. Strain partitioning within the lithosphere is controlled by its rheology. For realistic crustal and mantle rheologies (although weaker than those implied by recent laboratory experiments for the natu-
rally hydrous feldspar and quartz [21,22]), our models generate deformation patterns, which are consistent with geological and geophysical observations at the DST. In the brittle upper 10–20 km of the crust the deformation is localized in a few sub-vertical strike-slip faults with one of them taking up most of the strike-slip deformation. Deeper, the faults merge into a diffused, 20–40 km wide, ductile deformation zone which widens slightly with depth and has a 5–10 km wide high-strain core. The width of this deformation zone does not depend much on the thickness of the lithosphere and is mostly controlled by shear heating together with temperature- and strain rate-dependency of the viscosity of the rocks. Note, that the similar width (40 km) of the lower crustal shear zone for the DST was suggested previously based on elastic modelling of the subsidence of the Dead Sea basin [52].

5.3. Origin of the variations of surface topography across the DST

Our modelling suggests that modification of the mantle temperature is required to generate the observed asymmetrical regional uplift of the lithosphere around the DST. Based on the tectonic history of the region we suggest the following scenario. A big portion of the Arabian Shield and adjacent Mediterranean was tectonically stagnant following the Mesozoic, probably due to its location far from active mantle convection flows. This might have resulted in cooling and over-thickening of the lithosphere as well as in reduced temperature of the sub-lithospheric mantle in the entire region. The situation changed at some 30 Ma with the appearance of the Afar plume and the rifting and spreading of the Red Sea. We speculate that at this time, the rejuvenation of the asthenosphere, and destabilisation and destruction (i.e., thermal erosion due to small-scale asthenospheric convection) of the over-thickened mantle lithosphere began. This process likely peaked during the last 5 Myr and was marked by the surface uplift. The western shoulder of the DST, which is mechanically linked to the heavy oceanic lithosphere of the Mediterranean, did not uplift much during this process. In contrast, the eastern shoulder, which is largely decoupled from the Mediterranean lithosphere by the mechanically weak DST, was isostatically uplifted. First results of the DST modelling with consideration of the mechanical instability of the Arabian Shield lithosphere [53] support this scenario.

The specific feature of the DST topography is a narrow (20–30 km wide) valley, which was actually one of the reasons to call the DST the Dead Sea Rift. The 2-D thermo-mechanical models of the rifting processes (e.g., [54]) imply that the width of the rift valley is proportional to the initial thickness of the lithosphere. For a lithospheric thickness of more than 100 km (consistent with the surface heat flow of 50–60 mW/m²) the expected width of the rift valley is much larger than that actually observed at the DST. Our modelling suggests a possible explanation of this contradiction. As we have shown in the previous section the strike-slip deformation localizes in a relatively narrow zone, whose width does not depend on lithospheric thickness. When some subordinate extension is added (model 3) it is concentrated in the same narrow zone where the major strike-slip deformation is localized (see Fig. 8). The reason for this concentration is reduced lithospheric viscosity within this zone due to the high strike-slip strain rates. Accordingly, the surface expression of the lithospheric extension, the rift-like valley, is narrow and its width does not depend on the lithospheric thickness. What is in fact happening in this process is a major coupling of the strike-slip deformation and the fault perpendicular extension, which is simulated by our extended 2-D method but cannot be replicated by usual 2-D models.

A related question is how important is the rifting (transform-perpendicular extension) deformation component at the DST (e.g., [6,55,56])? We have seen above that model 3, which combines transtension (105 km strike-slip and 2–3 km of transform perpendicular extension) and lithospheric thinning, fits reasonably well with most of the observations along the DESERT line. If extension is larger than 2–3 km the model generates basin deeper than 0.4 km, observed at DESERT line [14]. Note, however, that our model does not consider transform-parallel extension, which may also contribute to the subsidence of the Arava Valley [52,57,58]. We expect that the transform-parallel extension will also concentrate in the same narrow deformation zone as the transform-perpendicular extension (for the same reasons discussed before) and will lead to similar consequences for the crustal structure and surface topography. Therefore the amount of transform-perpendicular extension neces-
sary to explain the observed topography and crustal structure may be in fact less than 3 km or even absent at all. From this we conclude that in the Arava Valley part of the DST, the rifting was relatively unimportant in accordance with geological observations [6].

6. Conclusions

1. We have developed fully coupled thermo-mechanical models in an extended 2-D approximation to study evolution of the continental lithosphere subjected to transtensional deformation during 20 Myr. The models have been focused on the geodynamic setting of the DST in the region between the Red Sea and the Dead Sea (Arava Valley), crossed by the DESERT transect. The geological data as well as geophysical and petrophysical observations in the DESERT experiment have been used to constrain initial and boundary conditions and to choose the thermal and rheological parameters. The model, which combines plate-scale transtension (strongly dominated by strike-slip deformation component) with thinning of the mantle lithosphere of the Arabian Shield at 5–10 Ma, and has relatively weak crust, replicates well most of the geological, geophysical and geodetic observations.

2. The shear strain at the DST in the Arava Valley is localized in a sub-vertical shear zone, which crosses the entire lithosphere. In the upper crust the deformation localizes at one or two major vertical faults located at the top of this zone. The width of this zone in the lower crust and upper mantle (20–40 km) is controlled by shear heating and temperature- and strain-rate-dependence of the viscosity of the rocks.

3. The strike-slip deformation localizes in the zone of minimal lithospheric strength, usually in the region of the highest temperature of the uppermost mantle. This temperature in turn depends on the thickness of the crust (the thicker the crust the higher is the sub-crustal temperature) and the thickness of the lithosphere (the thinner the lithosphere the higher is the sub-crustal temperature). The modeling suggests that the location of the DST between the Dead Sea and the Red Sea might have been controlled by the inherited regional minimum of the lithospheric strength.

4. Modelling without consideration of the possible transform parallel extension in the Arava Valley shows that the rifting deformation component (transform-perpendicular extension) at the DST in this region is likely less than 3 km. We expect that with consideration of the transform parallel extension, for which full 3-D modelling is required, the transform-perpendicular extension may even become unnecessary to fit the observations.

5. Uplift of the Arabian Shield adjacent to the DST requires young (<10 Ma) thinning of the lithosphere at and east of the plate boundary. Such lithospheric thinning is consistent with seismological observations, with the low present-day surface heat flow and with the high temperatures derived from mantle xenoliths brought up by Neogene–Quaternary basalts.

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