The origin of the Dead Sea rift

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Received 14 April 1993; revised version accepted 2 February 1994

Abstract

The Dead Sea rift is considered to be a plate boundary of the transform type. Several key questions regarding its structure and evolution are: Does sea floor spreading activity propagate from the Red Sea into the Dead Sea rift? Did rifting activity start simultaneously along the entire length of the Dead Sea rift, or did it propagate from several centres? Why did the initial propagation of the Red Sea into the Gulf of Suez stop and an opening of the Gulf of Elat start?

Using crustal structure data from north Africa and the eastern Mediterranean and approximating the deformation of the lithosphere by a deformation of a multilayer thin sheet that overlies an inviscid half-space, the regional stress field in this region was calculated. Using this approach it is possible to take into account variations of lithospheric thickness and the transition from a continental to an oceanic crust. By application of a strain-dependent visco-elastic model of a solid with damage it is possible to describe the process of creation and evolution of narrow zones of strain rate localization, corresponding to the high value of the damage parameter, i.e. fault zones.

Mathematical simulation of the plate motion and faulting process suggests that the Dead Sea rift was created as a result of a simultaneous propagation of two different transforms. One propagated from the Red Sea through the Gulf of Elat to the north. The other transform started at the collision zone in Turkey and propagated to the south.

1. Introduction

The Dead Sea rift is a plate boundary between the Sinai subplate and the Arabian plate (Fig. 1). It extends over 1000 km from the southern tip of the Sinai peninsula to the Taurus Mountains in Turkey. The southern and central portions of the Dead Sea rift trend north to north-northeast and approximately follow a small circle (Le Pichon and Francheteau, 1978; Joffe and Garfunkel, 1987). This rift offers a unique opportunity to study the process of rift propagation as well as other problems associated with continental rifting. Geological and geophysical evidence from the Gulf of Elat, at the southern part of the Dead Sea rift, and other parts of the rift indicate that two kinds of geodynamical processes occur here simultaneously (Ben-Avraham, 1987). The first is a transform motion which has led to a total displacement of 105 km on the faults along the Dead Sea rift since the time of its formation less than 20 Ma (Freund et al., 1970; Garfunkel, 1981) and to the formation of a series of pull-apart basins along the entire length of the rift zone. The other process is a propagation of actual crustal spreading activity from the Red Sea northward into the Gulf of Elat (Ben-Avraham, 1985).

The origin of the Dead Sea rift is unclear.
Fig. 1. Plate tectonic model for the eastern Mediterranean and surrounding areas (after McKenzie, 1972). The dashed line box indicates the area studied: 1 = Sinai; 2 = Red Sea; 3 = Gulf of Suez; 4 = Gulf of Elat. Arrows indicate motion of plates relative to Eurasian plate. Inset: main faults in northern Levant (after Walley, 1988).
Several models published recently have suggested that rifting activity along the Dead Sea rift propagated from the northern Red Sea in the south to the north (Vink et al., 1984; Steckler and ten Brink, 1986). In the southern part of the rift, gravity, magnetic, seismic refraction and heat flow data indicate a propagation of the spreading activity within the Gulf of Elat, from the Red Sea northward (Ben-Avraham, 1987). Only in the southern part and possibly also in the central part of the Gulf of Elat is actual mantle upwelling now taking place.

The northern part of the Dead Sea rift is located close to the collisional belts along the Cyprian Arc and Taurus Mountains. Stress generated in this zone can propagate a long distance into the interior of the adjacent plates. Based on the analysis of crustal structure variation in the eastern Mediterranean, Ben-Avraham and Grass0 (1991) offer an alternative view on the formation of the Dead Sea rift. They suggested that the initial cracking of the crust and faulting activity have started at the collision zone in the north and propagated southward. Rifting from the Red Sea northward began later. A simulation of a faulting processes along the northern Dead Sea and the Levant margin (Ben-Avraham and Lyakhovsky, 1992) suggests that the formation of the major structural elements could be explained as a result of a simultaneous propagation of the Dead Sea rift from the north and south.

In this paper we describe a simulation of the propagation of rifting along the Dead Sea trend by taking into account both an opening of the Red Sea in the south and collisional processes in Turkey in the north.

2. Plate tectonics of the region

The geometry of the four major plates, African, Arabian, Turkish (Anatolian) and Black Sea (Fig. 1) was derived from a model of the eastern Mediterranean and surrounding areas proposed by McKenzie (1972). The main sources of motion in the region are the stresses exerted on it through the northward motion of the Arabian and African plates. Rifting in the Red Sea, a plate boundary between Africa and Arabia, has been active since the Oligocene (Cochran et al., 1986). The northern Red Sea is still in the late stage of continental rifting (Steckler and ten Brink, 1986). Rifting in the Gulf of Suez, a northern arm of the Red Sea, began in the latest Oligocene/Miocene. A widespread unconformity, the Mid-Clysmic event at approximately 17 Ma (Garfunkel and Bartov, 1977) marks a dramatic change in its tectonic evolution.

Motion along the Dead Sea rift, which is younger than 19 Ma, accommodates 3/4 of the Africa–Arabian opening (Steckler and ten Brink, 1986). There is no clear evidence of rifting activity along the Dead Sea rift prior to the Mid-Clysmic event. Furthermore, the reorganization of the faulted blocks and change in subsidence rate in the Gulf of Suez at this time is consistent with the changes expected from the development of a new plate boundary (Steckler and ten Brink, 1986). Geological evidence on the timing of events in the Gulf of Suez and Gulf of Elat at the southern part of the Dead Sea rift, indicate that the Gulf of Suez initiated prior to the Gulf of Elat as the northernmost part of the Red Sea. However, soon thereafter, the Gulf of Elat developed and accommodated most of the plate motion north of the Red Sea.

Seismic refraction studies in the northeastern and southwestern Red Sea show that the adjacent continental crust is about 35 km thick and that it decreases to about 20 km under the coastline over a distance of 30 km (Le Pichon and Gaulier, 1988). Based on reconstruction of the Red Sea opening, Sultan et al. (1992) suggested that prior to rifting, the continental crust underlying the Red Sea was also about 35 km thick.

The Dead Sea transform runs more than 1000 km from the Gulf of Elat in the south to the Taurus–Zagros convergence zone in the north. A total left-lateral offset of about 105 km in its southern part is determined by matching numerous rock bodies and structures south of Lebanon (Freund et al., 1970; Joffe and Garfunkel, 1987). The southern and central portion of the Dead Sea rift trend north to north-northeast and approximately follow a small circle (Le Pichon and Francheteau, 1978; Joffe and Garfunkel, 1987).
In the northern Dead Sea rift, the Yammuneh Fault strikes N30°E and connects the central Dead Sea rift with the north-trending Ghab Fault (Walley, 1988). Besides the Yammuneh Fault, several other faults can be related to the plate boundary in this area.

Several hypotheses have been forwarded to explain the tectonic framework and faulting processes in the northern Dead Sea rift. Ron (1987) considered the Yammuneh Fault as a restraining bend of the Dead Sea rift and suggested that the regional shear in this area is accommodated by secondary faulting and block rotation. Walley (1988) objected to the assumption that the entire displacement on the Dead Sea rift is transmitted northward along the Yammuneh Fault and suggested that the left-lateral motion is transmitted along a number of faults. Girdler (1990) has suggested that the Dead Sea rift continues northward as the Ed Damur (Roum) Fault, which follows a trend similar to the central Dead Sea rift. He further suggested that the Roum Fault extends into the Mediterranean south of Beirut to meet the Cyprian Arc somewhere east of Cyprus. According to this scheme, the Yammuneh Fault is not a plate boundary. One major problem concerning the northern part of the Dead Sea rift is that the displacement on the Yammuneh Fault or any other fault in this area is relatively small, much less than the displacement on the southern and central Dead Sea rift. According to Walley (1988) only 60 km of displacement took place along the Yammuneh Fault.

Large variations in crustal structure exist in the Levant and eastern Mediterranean (Ben-Avraham and Ginzburg, 1990). The Levant passive continental margin marks a transition from the oceanic crust of the Levantine Basin to a typical continental crust of the Levant (Makris et al., 1983; Ginzburg and Ben-Avraham, 1987). The Levant itself is probably composed of distinct crustal domains of different origins. For example, the crust of the Galilee–Lebanon region, which is bordered on the east by the Dead Sea transform, is markedly different from the areas to the east (the Palmyra zone in Syria) and south (the Judea–Samaria region) (Ben-Avraham and Ginzburg, 1990). These crustal structure variations have probably caused a segmentation of the collision zone in the north along the Cyprian Arc and the Taurus Mountains. The segments are separated by large fault zones (Ben-Avraham and Grasso, 1991). Seismic refraction studies indicate a sharp boundary between continental and oceanic crusts along the northern Levant margin as compared with a wide transition zone along

![Fig. 2. Palaeotectonic map of Turkey of the Late Eocene–Early Miocene (after Sengor and Yilmaz, 1981).](image-url)
the southern Levant margin (Ginzburg and Ben-Avraham, 1987). Seismic reflection data show the existence of probable strike-slip faulting along the northern Levant margin (Ben-Avraham and Ginzburg, 1990).

The Turkish (Anatolian) microplate is located between the Eurasian, African and Arabian plates. Its northern boundary, the north Anatolian Fault, is one of the major tectonic features with well defined fault trace. The age of the north Anatolian Fault is not well known. Ketin (1968, 1969) suggested that the fault has been active since Pliocene–Quaternary time with a total displacement of a few tens of kilometres. Sengor and Canitez (1982) suggested a time of initiation between the Burdigalian and the Pliocene and cumulative offset of 85–90 km. Creep measurements and earthquake studies (Toksoz et al., 1979) indicate that the current average slip rate of the fault is about 2 cm/year.

The east Anatolian Fault is generally considered to be a major strike-slip fault bounding the Anatolian block and Arabian plate (McKenzie, 1978; Jackson and McKenzie, 1984; Sengor et al., 1985). This region has been and is still subject to compressive tectonics (Lyberis et al., 1992). The sedimentary cover displays a large, 5–10 km wide, curved fold structure, located along the east Anatolian Fault. Further south the Arabian plate seems to be undeformed. Kasapoglu and Toksoz (1983) suggested that the initiation of motion along the Dead Sea transform is older than the beginning of the strike-slip motion on both the east and the north Anatolian Faults. A palaeotectonic map of Turkey by Sengor and Yilmaz (1981) of the late Eocene–early Miocene (Fig. 2) suggests that the Africa–Eurasia convergence took place along a sinuous, uniformly north-dipping subduction zone beneath southern Turkey. This subduction zone was consuming very young oceanic lithosphere of the Cungus basin along its eastern extent, prior to the collision of Arabia and Eurasia, whereas south of central and western Anatolia much older lithosphere was being subducted. Subduction along the arcs south of central and western Anatolia is still taking place at present.

3. Stress field simulation

Simulation of the faulting processes in the crust raises two main problems: (a) What is the stress field distribution in the crust? (b) Which rheological model of the crustal material is able to describe the faulting process?

To calculate the stress field distribution, following England and McKenzie (1982, 1983) and Villette et al. (1986), we approximate the deformation of the lithosphere by the deformation of a thin sheet that overlies an inviscid half-space. Because the horizontal dimensions of the sheet are much greater than the vertical dimension, and the shear stresses are assumed to be zero on the top and the bottom of the sheet, the deformation is approximated by neglecting the vertical gradients of horizontal velocity components. This allows the deformation of the sheet to be expressed in terms of vertical averages of the strain rate and the deviatoric stress.

The lithosphere itself has also its internal structure. Typical continental lithosphere consists of a number of layers. They are: a sedimentary layer, granitic and basaltic layers (upper crust and lower crust), and an upper mantle (Fig. 3). The thickness of each layer in the simulated litho-

![Fig. 3. Model of the layered structure of the continental crust. See Fig. 5 for the location of the profile in the simulation area.](image-url)
sphere is variable and the rheology also varies from layer to layer. Thus, it is some possibility to introduce a depth-dependent structure and rheology into two-dimensional in-plane stress simulation. Under the lithosphere we assume the existence of an inviscid asthenosphere. Governing equations of the in-plane stress approximation were derived by a perturbation technique (Nayfen, 1981) for the integration of the equations of motion of the lithosphere along a vertical axis. To realize this procedure we defined the boundary conditions and the rheology of each layer. The upper boundary (topography) was assumed to be free, i.e. forces acting on this boundary are equal to zero. Forces and velocities are assumed to be continuous on all internal boundaries. The existence of an inviscid, as compared with lithosphere, asthenosphere provides a state of a zero shear force at the boundary between the asthenosphere and the lithosphere. The normal stresses at the base of the lithosphere are defined in accordance with the assumption of the local isostasy equilibrium. The rheology of the sedimentary layer was described by a linear viscous Newtonian body; for all the other layers, a viscoelastic model of a material with damage is described below. The effective viscosity of the sedimentary layer is assumed to be much smaller than the average viscosity of the crystalline crust.

Thus, we had to simulate stress distribution in two types of layers. The first type is the low-viscosity sedimentary layer, and the second type is all the other layers with similar rheological properties. To describe a viscous flow velocity distribution \( (v_i) \) in the first layer we used the Navier-Stokes equation in the following form:

\[
\rho_1 g_i - \frac{\partial p}{\partial x_i} + \eta_1 \Delta v_i = 0 \tag{1}
\]

\[
\text{div} \ (v_i) = 0
\]

where: \( \rho_1 \) = density of sediments; \( p \) = pressure; \( \eta_1 \) = viscosity of sediments; \( g_i \) = gravity acceleration vector \( (g_i = (0,0,-g)) \); \( i \) = index corresponding to the number of the coordinate axis, where 1,2 = horizontal axes, 3 = vertical axis; \( \Delta \) = Laplace operator.

Neglecting higher-order terms of small parameter \( \varepsilon = H/L \) (the ratio of average vertical size of the lithosphere to its horizontal size), we integrated (1) and obtained the equations that describe the velocity and stress fields in each layer of the model. Velocity of the motion in the sedimentary layer relative to the crustal velocity vector \( V_i \), under the sediments is (Mikhaylov, 1983):

\[
\begin{align*}
  v'_1 &= \frac{\rho_1 g}{2\eta_1} \frac{\partial h_1}{\partial x_1} (h_2 - x_1)(h_2 + x_1 - 2h_1) \\
  v'_2 &= \frac{\rho_1 g}{2\eta_1} \frac{\partial h_1}{\partial x_2} (h_2 - x_1)(h_2 + x_1 - 2h_1) \\
  v'_3 &= \int_{h_2}^{x_3} \left( \frac{\partial v_1}{\partial x_1} + \frac{\partial v_2}{\partial x_2} \right) dx_3
\end{align*} \tag{2}
\]

To take into account erosion and sedimentation during the evolution of the upper surface, the diffusion approach is applied. This approach is widely used in geomorphology and civil engineering to simulate the long-term behaviour of river systems. The diffusion model has been applied to the development of river delta's in fjord and glacial lakes (Syvitski et al., 1988), and to foreland basins (Fleming and Jordan, 1989; Sinclair et al., 1991). Following Kenyon and Turcotte (1985) an equation of motion of the upper boundary \( (h_1) \) is:

\[
\frac{\partial h_1}{\partial t} = v'_3 + \lambda \Delta h_1 + \psi \tag{3}
\]

The Laplace operator of \( h_1 \) with a transportation coefficient \( \lambda \) makes it possible to take into account a process of smoothing of the relief due to denudation, and a function \( \psi(x_1,x_2,t) \), to introduce into the model a description of the erosion and sedimentation processes.

Substituting (2) into (3) gives an equation for the partial temporal derivation of the relief:

\[
\begin{align*}
  \frac{\partial h_1}{\partial t} &= -V_1 \frac{\partial h_1}{\partial x_1} - V_2 \frac{\partial h_1}{\partial x_2} + V_3 + \lambda \Delta h_1 + \psi \\
  &= \frac{\rho_1 g (h_1 - h_2)^3}{3\eta_1} \Delta h_1 + \frac{\rho_1 g (h_1 - h_2)^2}{\eta_1} \\
  &\times \left[ \left( \frac{\partial h_1}{\partial x_1} \right)^2 + \left( \frac{\partial h_1}{\partial x_2} \right)^2 \right] - \frac{\partial h_1}{\partial x_1} \frac{\partial h_2}{\partial x_1} - \frac{\partial h_1}{\partial x_2} \frac{\partial h_2}{\partial x_2} \tag{4}
\end{align*}
\]
Eq. (4) is solved numerically for the known motion of the crust \( (V_1) \). The first approximation of the stress distribution in the sedimentary layer gives a pressure equal to a lithostatic pressure \( p = \rho g (h_1 - x_3) \) and absence of a deviatoric component due to the low viscosity.

According to the Maxwell visco-elastic constitutive relationship, the deviator of the stress tensor \( (s_{ij}) \) for the layers 2, 3, 4 is equal to:

\[
s_{ij}^n = \exp \left( -\frac{\mu^n}{\eta^n} t \right) \int_{-\infty}^{t} \mu^n \left( \frac{\partial V_i}{\partial x_j} + \frac{\partial V_j}{\partial x_i} \right) \times \exp \left( \frac{\mu^n}{\eta^n} t' \right) dt'
\]

where the index \( n = 2, 3, 4 \) denotes the consecutive layers, \( \mu \) is the effective shear modulus and \( \eta \) the effective viscosity, which are functions of the parameters of the layer, and \( t \) is time.

Using Eq. (5) we get the equations of motion for the visco-elastic medium analogous to the Navier-Stokes equation (1):

\[
\rho \frac{\partial v_i}{\partial t} - \frac{\partial p}{\partial x_i} + \exp \left( -\frac{\mu^n}{\eta^n} t \right) \int_{-\infty}^{t} \frac{\partial}{\partial x_j} \left[ \mu^n \left( \frac{\partial V_i}{\partial x_j} + \frac{\partial V_j}{\partial x_i} \right) \right] \exp \left( \frac{\mu^n}{\eta^n} t' \right) dt' = 0
\]

and ignoring a compressibility of material during a long term deformation we add:

\[
\text{div} (v_i) = 0
\]

In a similar way to the analysis of Eq. (1), we get the first approximation of the velocity components:

\[
v_{11}^n = V_1(x_1, x_2)
\]

\[
v_{12}^n = V_2(x_1, x_2)
\]

\[
v_{13}^n = -\left( \frac{\partial V_1}{\partial x_1} + \frac{\partial V_2}{\partial x_2} \right) x_3
\]

deviatoric stress components \( (s_{ij}) \):

\[
s_{11}^n = \exp \left( -\frac{\mu^n}{\eta^n} t \right) \int_{-\infty}^{t} \mu^n \left( 4 \frac{\partial V_1}{\partial x_1} + 2 \frac{\partial V_2}{\partial x_2} \right) \times \exp \left( \frac{\mu^n}{\eta^n} t' \right) dt'
\]

\[
s_{12}^n = \exp \left( -\frac{\mu^n}{\eta^n} t \right) \int_{-\infty}^{t} \mu^n \left( \frac{\partial V_1}{\partial x_2} + \frac{\partial V_2}{\partial x_1} \right) \times \exp \left( \frac{\mu^n}{\eta^n} t' \right) dt'
\]

and lithostatic pressure:

\[
p = -\int_{-\infty}^{x} \rho g dx
\]

These formulas allow the first approximation of the deformation of the lithosphere to be expressed by neglecting vertical gradients of horizontal velocity components in terms of strain rate and deviatoric stress components. This method also permits us to deduce an analytical solution for \( s_{13} \) and \( s_{23} \) components of the stress tensor. Using all components \( s_{ij} \) and the absence of shear forces on the boundary between lithosphere and asthenosphere it is possible to achieve equations of the motion of the lithosphere:

\[
\frac{\partial s_{ij}}{\partial x_j} + f_i = 0
\]

where:

\[
s_{ij} = s_{ij}^2 (h_2 - h_3) + s_{ij}^3 (h_3 - h_4) + s_{ij}^4 (h_4 - h_5)
\]

and \( f_i \) effective internal forces caused by variations in layer thickness:

\[
f_i = \rho_1 g \frac{\partial h_1}{\partial x_i} (h_1 - h_5) + (\rho_2 - \rho_1) g \frac{\partial h_2}{\partial x_i} (h_2 - h_5)
\]

\[
+ (\rho_3 - \rho_1) g \frac{\partial h_3}{\partial x_i} (h_3 - h_5)
\]

\[
+ (\rho_4 - \rho_1) g \frac{\partial h_4}{\partial x_i} (h_4 - h_5)
\]

For the steady-state case Eq. (10) gives an effective viscosity of the lithosphere \( \eta = \eta_2 (h_2 - h_3) + \eta_3 (h_3 - h_4) + \eta_4 (h_4 - h_5) \). This equation
and Eq. (11) for the internal forces are the multilayer analog of the one-layer approximation used by Sonder and England (1989).

Eqs. (8-11) are the mathematical analog of a 2-dimensional visco-elastic problem and are solved numerically by the "FLAC" algorithm described below.

4. Model of a medium with damage

A long-term action of stress and temperature will cause growth or healing of microcracks resulting in a changing of the rheological properties of a solid. Recent investigations of a fracturing process in granite by Yukutake (1989) and Lockner et al. (1991) show that this process can hardly be described in terms of a classical Griffits model of a single crack propagation. Lockner and Madden (1991) put forward a multiple-crack model of brittle fracture. Many attempts have been made to define and estimate suitable variables to describe the extent of fracturing. Such variables should be able to characterize the processes of fracturing quantitatively and should be suitable for modeling from the point of view of continuum mechanics. Among such approaches are Robinson's (1952) linear cumulative creep damage law, Hoff's (1953) ductile creep rupture theory, Kachanov's (1961, 1986) brittle rupture theory, Rabotnov's (1969) coupled damage creep theory and many modifications of these theories. Following this approach and using the strain-dependent moduli model, a description of the damage evolution of a fractured medium under long-term stress has been put forward (Lyakhovsky and Myasnikov, 1985; Myasnikov et al., 1990; Lyakhovsky et al., 1993). This model has made it possible to simulate the evolution of fracturing under tectonic stress, resulting in the appearance of seismic boundaries of a rheological nature (Mints et al., 1987; Lyakhovsky and Myasnikov, 1987, 1988) and investigate a faulting process in the northern Dead Sea rift (Ben-Avraham and Lyakhovsky, 1992).

In order to simulate a faulting process in terms of continuum mechanics a nondimensional damage parameter $\alpha$ is incorporated in a visco-elastic Maxwell model. This parameter lies in the interval from 0 to 1 and describes the evolution of the medium from an ideal undestroyed solid to an absolutely destroyed material. Elastic moduli are supposed to be linear functions of $\alpha$. Assuming the process of elastic deformation to be reversible and following the principle of maximum rate of entropy production, a phenomenological equation for damage evolution was suggested (Lyakhovsky et al., 1993):

$$\frac{d\alpha}{dt} = aI_1^2 + bI_2 + cI_{v1}I_{v2}$$

where: $t =$ time; $I_1 = \epsilon_{ii}$ and $I_2 = \epsilon_{ij}\epsilon_{ij}$ = the first and the second invariants of the elastic strain tensor $\epsilon_{ij}$. According to the multilayer approximation, an average value of the elastic deformation for each layer was taken to calculate the damage distribution in each layer.

To simulate any flow it is necessary to define an effective viscosity as a function of parameters of a solid. Power law viscosity is usually assumed for rocks in the ductile regime. Not much is known about the viscosity in the brittle regime, except that it is very strongly dependent on the level of fracturing. To take into account a stronger variation of viscosity in comparison with elastic moduli, we assume the viscosity $\eta$ to be an exponential function of the damage parameter $\alpha$:

$$\eta = A \exp(-B\alpha)$$

where $A$ and $B$ are constants.

Variation of $\alpha$ between destroyed and undestroyed zones causes changes of the effective viscosity of about $10^4$ times. Viscosity of the upper crust, for instance, changes from about $10^{27}$ Pa s for an undestroyed block up to $10^{23}$ Pa s for a weak zone. A detailed description of the model is presented by Lyakhovsky et al. (1993).

The application of this model makes it possible to simulate the rheological transition from a ductile to a brittle regime and the process of creation and evolution of narrow zones of strain rate localization, corresponding to a high value of damage parameter, i.e. the evolution of fault zones.
5. Computation techniques

Numerical simulation has been done using the "FLAC" algorithm (Cundall and Board, 1988; Cundall, 1989). The formulation is explicit in time, using an updated Lagrangian scheme to provide the capability for large strains. FLAC is believed to offer advantages over conventional finite element schemes in cases where flow and material instability occur. Physical instability can be modeled without numerical instability if internal terms are included in the equilibrium equations. The FLAC method has an advantage over implicit methods because it is computationally inexpensive for each time step and it is memory-efficient because matrices storing the system of equations are not required. An application of this method to the simulation of a visco-elastic flow is described in Poliakov et al. (1993).

The computational mesh consists of quadrilateral elements, which are subdivided into pairs of constant-strain triangles, with different diagonals. This overlay scheme ensures symmetry of the solution by averaging results obtained on two meshes (Cundall and Board, 1988). Linear triangular element shape functions \( L_k \) are defined as:

\[
L_k = a_k + x b_k + y c_k
\]

where \( a_k, b_k, c_k \) are constants, and \( x \) and \( y \) are grid coordinates. These shape functions are used to interpolate the nodal velocities \( v_i^{(k)} \) within each triangle element (e):

\[
v_i(x, y) = \sum_{k=1}^{3} v_i^{(k)} L_k
\]

This formula enables the calculation of the strain increments in each triangle and element stresses using a constitutive law. All the parameters of the media inside the element, such as damage, viscosity and elastic moduli are approximated by a constant. When stresses in each triangle are known, the forces at the node number \( n \) (\( F_i^{(n)} \)) are calculated by projecting the stresses from all elements surrounding the node:

\[
F_i^{(n)} = -\sum \frac{1}{2} \alpha_{ij}^{(e)} \left( n_j^{(1)} \Delta l^{(1)} + n_j^{(2)} \Delta l^{(2)} \right)
\]

where \( n_j \) is the \( j \) component of the unit vector normal to each side of the element adjacent to the node \( n \). The length of each side is denoted by \( \Delta l \). The minus sign is a consequence of Newton's third law. New velocities are computed by integrating the equation of motion over a given time step \( \Delta t \):

\[
v_i^{(n)}(t + \Delta t) = v_i^{(n)}(t) + \frac{F_i^{(n)}}{\rho} \Delta t
\]

If a body is in a state of mechanical equilibrium, the net forces on each node are zero; otherwise, the nodes are accelerated. The new coordinates of the grid nodes are computed by:

\[
x_i^{(n)}(t + \Delta t) = x_i^{(n)}(t) + v_i^{(n)} \Delta t
\]

and then the calculations are repeated for a new configuration. Together with the deformation a new damage distribution is calculated using known stresses. Thus, a visco-elastic deformation of a damaged material is simulated.

When the mesh becomes too distorted it is necessary to apply a special remeshing procedure. An interpolation from the old nodes to a new mesh must be done only at this stage. This is in contrast to the Eulerian method where this interpolation must be performed at every time step. An interpolation of velocities from the old mesh to a new one is made by using shape functions \( L_k \) (Eq. 14). To interpolate media parameters, which are constant inside each element, these parame-

Fig. 4. The location of an element of a regular mesh with additional points for reinterpolation from a distorted mesh.
ters are interpolated from the old elements to an additional 15 points inside the new element of the regular mesh (Fig. 4). An average value calculated over these points is projected onto the new element. This method of interpolation makes it possible to reduce the numerical diffusion caused by a discrete mesh approximation.

6. Results of computer simulation

In order to simulate the plate tectonic evolution of the region boxed in Fig. 1 two types of lithosphere were introduced.

(1) The lithosphere of the Africa, Arabia, Turkey and Black Sea microplates (Fig. 5) is assumed to have a continental crust of about 35 km thickness. It has a 15 km thick granitic layer with 2.7 g/cm³ density and a 20 km thick basaltic layer with 2.9 g/cm³ density. The density of the upper mantle is assumed to be 3.4 g/cm³. A very thin sedimentary layer of about 0.5 km with 2.4 g/cm³ density initially covers a continental-type lithosphere. It should be mentioned here that the African and Arabian plates were assumed to be initially similar and homogeneous. Crustal structure variations between Galilee-Lebanon, Judea and Sinai were not taken into account. Therefore it is difficult to expect an exact simulation of the geometry of the newly created Dead Sea rift, which locally must follow some pre-existing weak zones and boundaries of blocks with different crustal structure.

(2) An oceanic lithosphere with a 4 km thick sedimentary layer and a 7 km thick basaltic layer represents the eastern Mediterranean, which is probably an oceanic part of the African plate, and the Cungus basin, which was an oceanic northern part of the Arabian plate. The north Anatolian Fault and the Red Sea rift were introduced initially as the destroyed zones with a high value of the damage parameter \( \alpha \). The initial block geometry is based on the palaeotectonic map shown in Fig. 2 and the rigid block kinematic reconstruction made by Lyberis et al. (1992).

The boundary conditions shown in Fig. 5 correspond to the northward motion of Africa and Arabia with a different rate. A free slip condition is enforced on the African and the Mediterranean parts of the western boundary. Therefore the boundary of this section can move only in a north-south direction. This boundary condition is in agreement with a current northward motion of the African plate and the modern tectonic stress field in the Mediterranean region (Rebai et al., 1992). The Turkish part of the western boundary is assumed to be free, whereby westward motion of the Turkish microplate is possible. The Black Sea microplate is taken as fixed in its place. This condition implies that all motions in the simulated area are defined with respect to the fixed Black Sea microplate. The eastern boundary of the simulated region south of the Black Sea microplate moves northward with a constant rate and without any force acting in the western direction. These boundary conditions are in agreement.
with the results of a finite element model of the collision of the Arabian and Eurasian plates (Kasapoglu and Toksoz, 1983).

At the initial stage of evolution at about Mid-Miocene time (Fig. 6) the boundary conditions and the geometry of the lithospheric units cause a northward motion of the Arabian and African plates with different rates. This motion results in a zone of very high gradient of the horizontal velocities (arrows in Fig. 6) or high horizontal deformation that lies between the eastern Mediterranean and the Cungus basin in the south and the undeformed Turkish microplate in the north. The thin oceanic crust subducts the thick and rigid continental crust along the southern margin of the Turkey plate. There is also a rotation of the Arabian plate associated with the opening of the Red Sea and its propagation in a northwestern direction. This propagation results in the creation of a zone of low topography (Fig. 7), which may be interpreted as the opening of the Gulf of Suez.

The beginning of the collisional process between Arabia and Turkey influences the stress distribution in the whole simulated area. Northeast of the Red Sea a small anomaly in the topography is developed. The rheological model indicates that there is a transition zone between undestroyed and initially destroyed crust along the Gulf of Suez. As a result the new direction of the rifting process along the Dead Sea rift starts not at the tip of the Gulf of Suez, where the weakening only begins, but slightly southward, where the crust is already weak. The “partly rifted” zone at the north end of the Red Sea rift may be interpreted as the future Gulf of Suez, whose formation started before the creation of the Dead Sea rift.

Continuation of the collisional process (Fig. 8) results in dramatic changes both in the evolution...
of the topography and in the distribution of horizontal velocities. There is no further propagation of the Red Sea into the Gulf of Suez. On the other hand, a newly created northeastern trending of the propagation of the Red Sea into the Gulf of Elat becomes active. At the same time in the north a new transform zone starts its propagation to the south at the boundary between the zone of continental collision and the closing Cuncus basin. The simulation, thus, suggests a simultaneous propagation of the Dead Sea rift from the Red Sea in the south through the Gulf of Elat northward and from the collision zone in the north southward. In the simulated case, the Dead Sea rift (Fig. 9) propagates along the shortest straight line joining the northern Red Sea and collision zone. Horizontal deformation of the simulated area corresponding to the latest stage is shown in Fig. 10. There is a newly created high deformation zone with a left-lateral motion, which corresponds to the Dead Sea transform. A westward motion of the undeformed Turkish microplate accumulates on the east and north Anatolian Faults. There is also a newly created right-lateral strike-slip fault at the base of the northern Levant margin. Similar to the northern Dead Sea, this zone starts its propagation to the south at the boundary between the zone of continental collision and the eastern Mediterranean basin. The combination of the northern Dead Sea transform and the fault zone along the Levant margin causes the northward motion of the northwestern part of the Arabian plate to be at a reduced rate as compared with the areas east and west of it (Ben-Avraham and Lyakhovsky, 1992).

It should also be noted that in spite of the existence of a destroyed zone along the north...
Fig. 10. Horizontal deformation at the last stage of simulation. Note the development of a left-lateral shear zone along the Dead Sea rift and a right-lateral shear zone along the northern Levant margin. 1 = Dead Sea rift; 2 = north Anatolian Fault; 3 = east Anatolian Fault; 4 = shear zone at the Levant continental margin.

Anatolian Fault, the simulation shows no development of any significant deformation and topographic features along this zone until the closing of the Cungus basin, which happened at about the same time as the creation of the Dead Sea rift.

7. Discussion and conclusion

This mathematical model of the evolution of the Dead Sea rift suggests that the rift was created as a result of the propagation of two fracture zones at its northern and southern ends towards each other. Due to the collision of Arabia with Eurasia and the strike-slip motion along the east Anatolian Fault a new fracture zone was created and formed the northern part of the Dead Sea rift. It propagated from the north to the south. Active tectonics in the northern part of the Arabian plate also causes a dramatic change of the stress field around the northern Red Sea. As a result, rifting activity changes its direction of propagation from the Gulf of Suez into the Gulf of Elat and a new plate boundary of the transform type was created.

The simulation suggests that the Dead Sea rift is made of two different fault zones, which later on were combined into one feature. The simulation outlines only the general trend of the Dead Sea rift. The exact geometry, especially in the northern part, is affected by local crustal structure variations. The resulted geometry of the Dead Sea rift follows the actual location of the southern and central Dead Sea rift, but is derived from the northern Dead Sea rift. This suggests that the crustal structure of the African and Arabian plates in the south is homogeneous, while in the north it is variable.

Acknowledgements

Critical reading by B. Colletta, Y. Podladchikov, R. Guiraud, J. Chery and F. Lucazeau improved the manuscript. Work by V. Lyakhovsky was supported by grants from the Israel Ministries of Science and of Absorption.

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